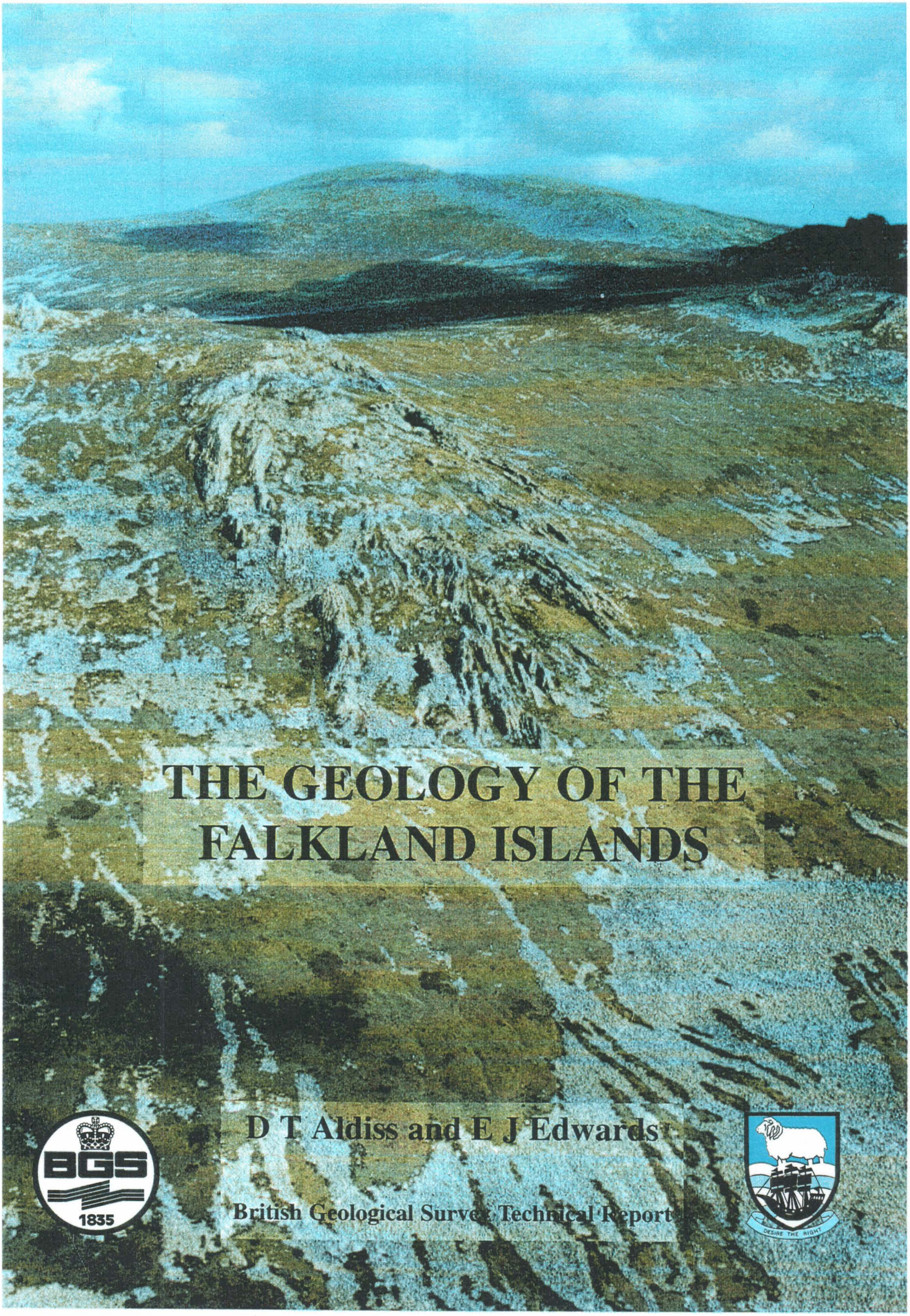

THE GEOLOGY OF THE FALKLAND ISLANDS

D T Aldiss and E J Edwards

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D T Aldiss and E J Edwards

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Cover photograph: Medium-scale anticline in Devonian quartzites of the Port Stanley Formation with stone runs in the Wickham Heights, East Falkland. Mount Usborne (705 m) is in the distance.

British Geological Survey, Keyworth, Nottingham, NG12 5GG, UK
Department of Mineral Resources, Falkland Islands Government, Stanley, Falkland Islands

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1999

SUMMARY

This report is complementary to the 1:250 000 scale geological map of the Falkland Islands compiled in 1998. The report and map are products of the Falkland Islands Geological Mapping Project (1996-1998).

Geological observation and research in the Islands date from 1764. The Islands were visited during two pioneering scientific cruises in the 19th century. Subsequently, many scientists visited en route to the Antarctic or Patagonia. Geological affinities to other parts of the southern continents, especially South Africa, were noted early in the 20th century. There have been two previous attempts to create a geological map of the Islands, both motivated primarily by the search for economic mineral deposits onshore. In the last few decades much effort has been directed to understanding the Falklands' place in Gondwana, the processes by which the Islands have moved to their present position by continental drift and the concomitant development of offshore sedimentary basins. Considerable progress in describing the superficial deposits was made in the 1970's, and during the last ten years.

The stratigraphic subdivisions of the geological sequence shown on the previous geological maps have been substantiated and defined more rigorously than before. In addition, several new stratigraphic units have been recognised. Each unit is described with an introductory summary of composition and distribution, followed by comments on nomenclature and stratigraphic relationships, associated landforms, distinguishing characters, and the criteria used to locate and survey the stratigraphic base. Detailed descriptions of composition, sedimentary structures and fossil content then lead to brief comments on the environment of deposition, age and correlation.

The bedrock geological formations ('solid geology') can be divided into four age groups. The Proterozoic granites and amphibolite facies gneisses of the Cape Meredith Complex (about 1150 to 1000 million years old) are overlain in turn by sedimentary sequences of the ?Silurian to Devonian West Falkland Group and the Carboniferous to Permian Lafonia Group. Jurassic igneous rocks are widespread but only locally abundant.

The West Falkland Group is dominated by sandstones, with some siltstones and mudstones. The oldest of four formations, the Port Stephens Formation, is divided into seven members, representing marine to fluvial environments. The basal member on East Falkland is probably the oldest part of the sedimentary sequence and might be latest Ordovician in age, but is more probably Silurian. The overlying Albemarle Member is notable for abundant trace fossils, mainly *Skolithos,* but also contains a new ichnospecies of *Heimdallia.* The succeeding marine Fox Bay Formation contains the Early to Middle Devonian Malvinokaffric invertebrate fossil fauna. One proximal facies member is recognised in the west. The Port Philomel Formation represents deltaic facies. It is notable for abundant fossil plant debris, most conspicuously lycophyte stems. The Late Devonian Port Stanley Formation, which includes the Stanley Quartzite, marks a return to marine conditions. The sandstones in the West Falkland Group are mostly quartz arenites and subarkoses, consistent with derivation from an area of stable continent crust. The West Falkland Group can be correlated with parts of the Cape Supergroup of South Africa.

The Lafonian Group is lithologically more varied and is divided into five formations. The fine-grained marine sediments of the Late Carboniferous Bluff Cove Formation are confined to East Falkland. They are followed by the thick tillites and thin mudstones of the Permo-Carboniferous Fitzroy Tillite Formation. One distinctive bedded interval is defined as a member. The Port Sussex Formation, with three members, includes carbonaceous mudstones with up to 40 per cent total organic carbon. The Brenton Loch Formation is also divided into three members. It is composed of sandstones with siltstones and mudstones, which represent combinations of basinal and turbiditic deposition in a large lake. The basinal sediments include varvites and some striking examples of the trace fossils *Umfolozia* and *Undichnia.* The Bay of Harbours Formation is generally similar lithologically, but was deposited in deltaic environments. It includes a sandy proximal facies member and has common plant fossils, including *Glossopteris*. Most of the sandstones in the Lafonia Group contain detritus derived from a distant contemporary volcanic terrane and there are also sporadic thin tuffs. The lowest part of the Lafonia Group can be correlated with

the Dwyka Group of South Africa, and the remainder with the parts of the Ecca Group and the lowermost Beaufort Group.

Better understanding of the distribution and structure of the sedimentary sequences has allowed their thickness to be estimated more reliably. In general, each stratigraphic unit is thinner on West Falkland than on East Falkland. It is likely that West Falkland lay near one margin of the Permo-Carboniferous depositional basin, and that no part of the Lafonia Group was ever deposited in the southernmost part of West Falkland. Mid-Carboniferous sedimentary dykes occur locally in the south-west of West Falkland.

Jurassic igneous dykes are widespread, especially in West Falkland. They are mostly composed of various types of dolerite, but more evolved rocks, including felsite, also occur. Those to the west of Falkland Sound can be divided by their field characteristics into six dyke swarms. Most of the dykes date from the Early Jurassic (about 190 million years ago) but Middle or Late Jurassic dykes could also occur. The latter are likely to be of similar age to Jurassic rhyolitic volcanic rocks which are inferred to occur offshore, close to the Islands. These could be accompanied by basalts, which might be the source of the agates known as 'Falklands pebbles'.

An east-west to NW-SE trending fold and thrust belt in the northern part of the Islands formed during the Permo-Triassic Gondwanide regional deformation. That was followed by at least four episodes of more localised deformation. A series of large-amplitude drape folds trending between north-south and ENE-WSW formed during the second phase as a result of passive deformation of the sedimentary cover over deep-seated faults. Each fold pair represents up-to-the-west fault movements in the underlying basement. During the third deformation, dextral strike-slip faulting with minor folding occurred in Falkland Sound and the adjacent parts of both main islands. About five to ten kilometres of dextral displacement occurred on the Falkland Sound Fault Zone. The fourth phase of deformation is represented onshore by a thrust sheet in the far north of the Islands. This thrusting is possibly younger than the Early Jurassic dolerite dykes. During the final phase of deformation the eastern part of East Falkland was uplifted by about three kilometres relative to the rest of the Islands. This uplift caused the formation of the Goose Green Graben, one of the several fault structures marking the Goose Green Axis, a long-lived tectonic lineament extending the length of East Falkland. The configuration of the Lafonia Group depositional basin was partly controlled by faulting on the Goose Green Axis and along Falkland Sound.

Metamorphic grade in the sedimentary sequences has been determined by measurement of illite crystallinity and vitrinite reflectance. Most of the strata exposed in the Islands have experienced only very low grade metamorphism and remain in the late diagenetic zone. The central and southern portions of East Falkland reached the anchizone, but only the centre of that island is in the high anchizone, very locally reaching the epizone (greenschist facies). The outer Jason Islands show the highest grade of metamorphism, in the epizone.

The Falkland Islands lie within the 'Falklands Microplate', part of an assemblage of crustal blocks at the southern end of the South American continent. In Early Jurassic times, these blocks lay between what are now South Africa and East Antarctica. Several lines of evidence show that the Islands were then inverted north to south relative to their present position, but the mechanism by which they were rotated remains very little understood.

Superficial deposits are widespread and varied, but all except the Neogene West Point Forest Bed are no older than the late Quaternary. Stony clays deposited by solifluction are almost ubiquitous on the higher ground. Stone runs (blockslopes and blockstreams) formed at the same time as the solifluction deposits during the last ice age. Both are the product of repeated freezing and thawing of the ground under periglacial conditions. The stone runs are formed by a type of frost-related stone sorting in which a fine-grained matrix is gradually displaced by ice. Stone runs can be classified according to their shape into patches, streams, stripes, terraces, fans and rivers. There is also a type of 'high-level' stone run which appears to be less well-developed than the other kinds. Blockfields occur in some summit areas, as does some small-scale patterned ground. Scree and landslips are relatively minor.

Pingo scars occur locally, demonstrating that even the low-lying ground was subject to permafrost at times during the Quaternary. However, glacial deposits and landforms are confined to a few of the highest hills. Aeolian sand is mostly found on the leeward side of lakes and of west-facing beaches. Peat is widespread and occurs both on high ground and in the valleys. Tussac peat is confined to coastal localities. Most peat has formed during the last 14 000 years, but there are rare thin deposits from between 36 000 and 26 000 years ago. Alluvium and river terrace deposits are relatively restricted. Lakes are numerous and formed by a variety of processes including aeolian deflation and thawing of ground ice. Lacustrine deposits are widespread but seemingly nowhere very extensive. Raised beach deposits and raised marine erosion levels are known from many parts of the Islands. There is evidence that past sea levels have been as much as 50 metres lower and 80 metres higher than at present.

The Falkland Islands have not been thoroughly prospected for economic mineral deposits. Traces of gold occur in stream sediments. It is possible that economically significant concentrations of rutile, zircon and other heavy minerals occur in raised beach deposits. Some smooth lacustrine clays may be of local use for pottery. Although parts of the offshore sedimentary basins are prospective for hydrocarbons, there is no significant chance that oil or gas occur onshore. Only very sparse traces of coal are present.

Devonian quartzite and Permian sandstone have been used in small amounts as building stone. Devonian quartzite is the main source of crushed stone but Permian sandstone and Permo-Carboniferous tillite have also been used. Small deposits of flagstone are fairly widespread but at present it is quarried only at Fox Bay. Wind-blown sand is generally used for building purposes. Small deposits of lime gravel (calcareous bryozoan debris) found on some beaches could meet some local needs for agricultural lime, but there is no other indigenous source of lime. Some water shortages could probably be alleviated by the development of groundwater resources.

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Acknowledgements

The Falkland Islands Geological Mapping Project could not have been carried out without the support of many people in many different ways. We would like to thank all those who have helped and encouraged the project, particularly the Islanders without whose goodwill, interest and ready assistance we would have made little progress.

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D T Aldiss
British Geological Survey
Keyworth
Nottingham
NG12 5GG
UK

E J Edwards
Falkland Islands Government
Stanley
Falkland Islands

1. INTRODUCTION

Even on a World map, the Falkland Islands draw the eye. The eastern seaboard of South America curves gently southwards from Rio de Janeiro until a scatter of islands appears offshore at latitude 52°S. These are obviously not part of the tenuous chain of mid-Atlantic volcanic islands, nor of the elegant trace around the Scotia Sea connecting the Antarctic Peninsula with Tierra del Fuego. This anomalous position is most clearly seen in the topography of the sea floor in the Falklands region (Figure 1.1). The Falklands lie on the elongated submarine Falkland Plateau. This is connected to the Maurice Ewing Bank, on which water depths of less than 2000 metres are encountered up to 1800 kilometres east of the South American coast, well beyond the continental shelf to the north. A view of a larger scale map presents more intriguing matters. Here are two large islands of similar size, while the rest are a good deal smaller. While much of the coastline is highly indented, some sections are remarkable for their linearity. The southern part of East Falkland is attached to the rest only by a narrow isthmus. East Falkland is bisected east-to-west by a compact chain of hills, but on West Falkland the distribution of high ground is far less regular. On the ground, the open skylines reveal yet further questions: Why are some hills craggy and others not ? Why can one find fossil shells in some places and fossil wood in others ? Why do so many ponds lie on top of hills ? What are the stone runs ? And so on according to which piece of ground the enquirer knows best.

The answers to these questions and many others lie in a study of the geology of the Islands: what they are made of and the processes by which they were formed. This report describes the earth materials found in the Falkland Islands and discusses their formation in the context of the geology of the southern continents. It is complementary to the 1:250 000 scale geological maps of the Islands produced by the Falkland Islands Geological Mapping Project (1996-1998), which was funded by the Falkland Islands Government (Aldiss, 1997; Aldiss and Edwards, 1998) (Appendix 1).

Apart from the quest to satisfy a sense of curiosity about the Earth beneath our feet, there are several general reasons to carry out a geological survey, and these each apply in the Falkland Islands to some extent. Geological maps assist mineral exploration of all kinds. They can provide information to the construction industry about ground conditions. They help locate constructional materials and groundwater supplies. They provide data for environmental protection and land-use planning. They provide an essential frame of reference for geological research of other kinds, research which itself may be of value in one of the practical applications of earth science.

1.1 Geography of the Falkland Islands

The Falkland Islands mostly lie between about 51°S and 52°30'S and 57° 45'W and 61°30'W, with one outlying island, Beauchêne Island, some 50 kilometres to the south, at about 52° 55'S and 59° 11'W. The archipelago is about 700 kilometres north-east of Cape Horn and some 500 kilometres east of the nearest part of the South American continent (Figure 1.1).

The Falkland Islands consist of two large islands (East Falkland, about 6700 square kilometres, and West Falkland, about 5300 square kilometres) with approximately 700 smaller islands, a total land area of some 12 173 square kilometres. The group extends about 250 kilometres east to west and, excluding Beauchêne, 155 kilometres north to south. The most easterly settlement is the capital town, Stanley. It is about 240 kilometres from the most westerly dwellings, on New Island. Although the land area is comparable to that of Northern Ireland, the extent of the archipelago is much greater, similar to that of Wales.

The highest point is on Mount Usborne on East Falkland (705 metres), with Mount Adam (700 metres) being the highest point on West Falkland. There is generally a clear demarcation between the steeper, higher ground, which is largely underlain by Silurian to Devonian sandstones and quartzites, and the gently undulating lowlands. Lines of hills tend to follow the regional structural trends, which are mostly east-west to NW-SE, with NE-SW alignment in the east of West Falkland. There are numerous rivers and streams, mostly occupying peat-lined valleys. Most are short, the principal exceptions being the Chartres River and the Warrah River on West Falkland, and the San Carlos River, the Arroyo Malo and Orqueta Arroyo on East Falkland. As noted by Greenway (1972), the drainage patterns are closely related to the underlying geological structure. Much of the coastline is highly indented, with drowned valleys forming sheltered inlets, but some stretches are markedly linear, reflecting the underlying geological structures.

Although the Falkland Islands are at the same latitude south as southern England is north, they lie within the cold northerly Falklands Current, a branch of the Circum-polar Current. They are also exposed to the westerly winds of the 'Furious Fifties', albeit in the lee of the South American continent. The effect of the consequent 'rain shadow' is that annual rainfall is much lower than might be expected, and the Falklands enjoy more hours of sunshine than London. Variations in temperature are moderated by the maritime climate.

For the period 1986-96 at Mount Pleasant Airport (MPA), the annual mean maximum temperature was 9.9° C, the annual mean minimum temperature 3.1°C, the mean annual rainfall 570 millimetres, with 4.7 hours annual average daily sunshine and an average of 16 days air frost in July. Typical monthly mean temperatures for the Falklands are 19°C (Jan) and 2° C (July) and the range is from about 30° C to −10°C. Snow is usual in the winter months but rarely lies for more than a week. The prevailing wind direction (70 per cent of the time) is in a broad arc from SSW to NNW. Strong winds are commonplace, especially in the summer, with an average of 34 gale days per year recorded at MPA (1986-96). Rainfall varies significantly across the Islands, reaching more than 600 millimetres annually in Stanley but closer to 400 millimetres at Fox Bay and North Arm. The wettest parts of the Falklands get about as much rain as the driest regions of the United Kingdom. Other weather statistics are quoted by Moore (1968), Strange (1983), Roberts (1984) and Richards (1997).

Although the Falklands have a semi-arid climate, gentle relief and widespread impermeable soils combine to keep large areas of ground wet for much of the year. The strong winds and dry summers mean that no trees grow naturally. The natural vegetation is mostly either grassland or dwarf shrub heath (Moore, 1968; Strange, 1983).

1.2 Geological outline

The oldest rock formations found in the Islands are the granites and gneisses of the Proterozoic Cape Meredith Complex, which crops out only at the southern tip of West Falkland (Figure 1.2 and Table 1). The Big Cape Formation is made up of amphibolites with lesser leucocratic gneisses, originally formed as volcanic rocks, and minor metasedimentary gneisses. It was intruded by three phases of granite and granitic gneiss which were in turn intruded by dykes of dolerite and of lamprophyre.

The rest of the Islands are mostly underlain by sedimentary rocks. These can be divided into the ?Silurian to Devonian West Falkland Group and the Carboniferous to Permian Lafonia Group. The West Falkland Group is dominated by quartzose and subarkosic sandstones, with some siltstones and mudstones. It has been divided into four constituent formations (Table 1) representing environments from fluvial to deltaic and marine shelf. The Lafonia Group is lithologically more varied, including a thick tillite sequence and an interval of carbonaceous mudstones in addition to varied sandstones and siltstones, with some mudstones. Most of the sandstones contain detritus derived from a contemporary volcanic terrane (itself not otherwise represented in the Islands). The Lafonia Group is divided into five constituent formations (Table 1) representing a major glaciation followed by basinal, turbiditic and deltaic deposition in a large lake.

There are numerous dolerite dykes (and a few sills) on West Falkland, with a few dykes on East Falkland. They can be divided into seven groups according to their orientation, distribution and field character. Dykes of more evolved compositions, including spherulitic felsites, occur in some areas. Most of the dykes date from about 190 Ma (million years) ago, in the Early Jurassic, although some could be from the Middle or Late Jurassic. South Fur Island is part of a large dolerite sill. Acid and basic volcanic rocks, probably of Middle to Late Jurassic age, are inferred to crop out on the seafloor close to the north of the Islands. Carboniferous sedimentary dykes occur locally.

Permian to Mesozoic deformation in the Falkland Islands can be divided into at least five phases of folding and faulting, or of faulting with negligible folding. The earliest is represented by an approximately east-west belt of regional folding and subparallel faulting which extends throughout the northern part of the Islands, fading out to the south. Metamorphism was mostly of very low grade but reached the epizone (greenschist facies) in central East Falkland and on the Jason Islands. The second phase of deformation created a series of large folds trending between north-south and ENE-WSW and found at intervals throughout the Islands, except in the eastern part of East Falkland. These folds all seem to have formed above deep-seated blind faults. The third deformation appears to have occurred during a dextral rotation of the Falkland Islands but its effects are mainly confined to strike-slip faulting and lateral

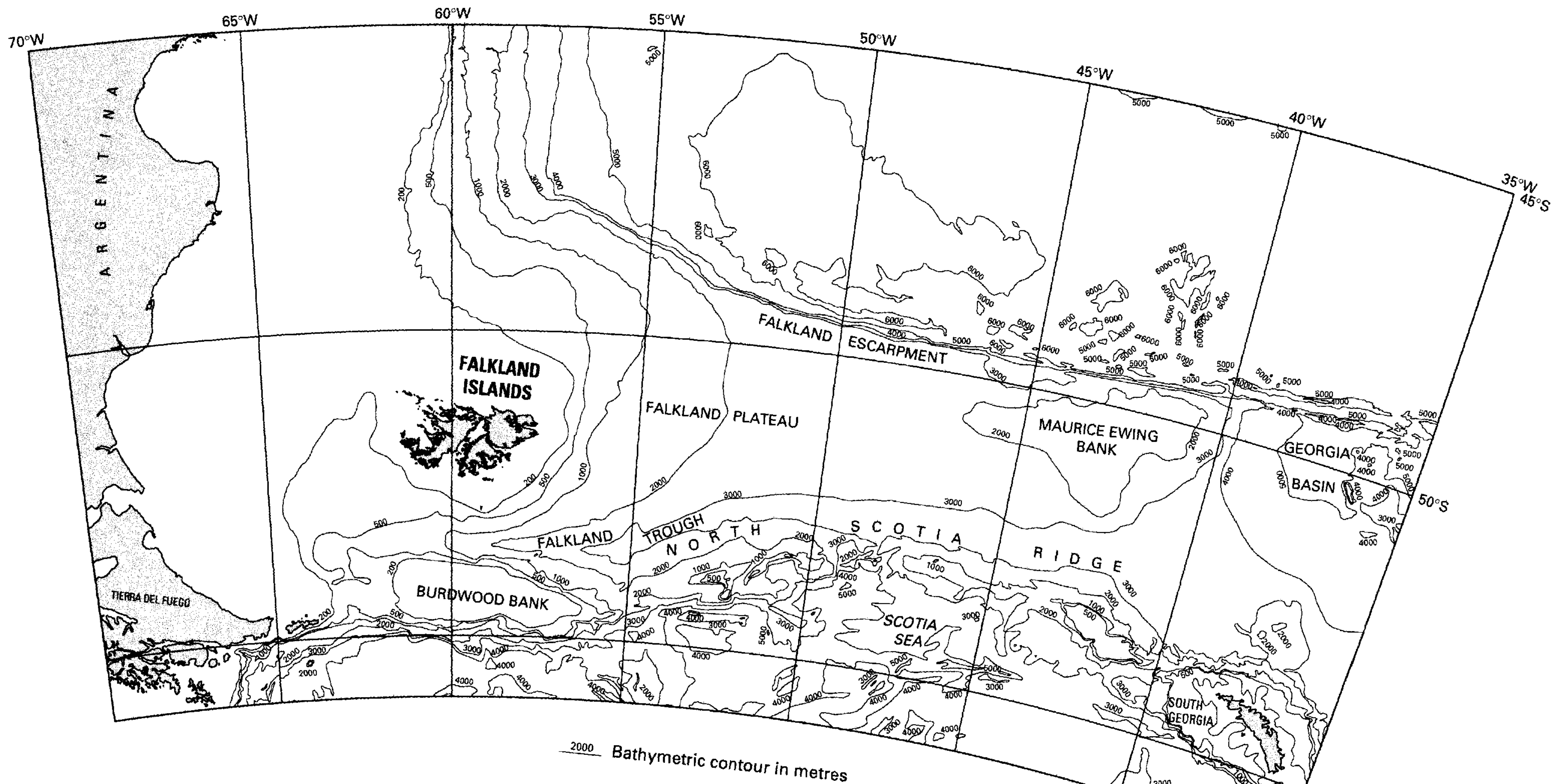


Figure 1.1 Location of the Falkland Islands

Figure 1.2 Simplified Geological Map of the Falkland Islands

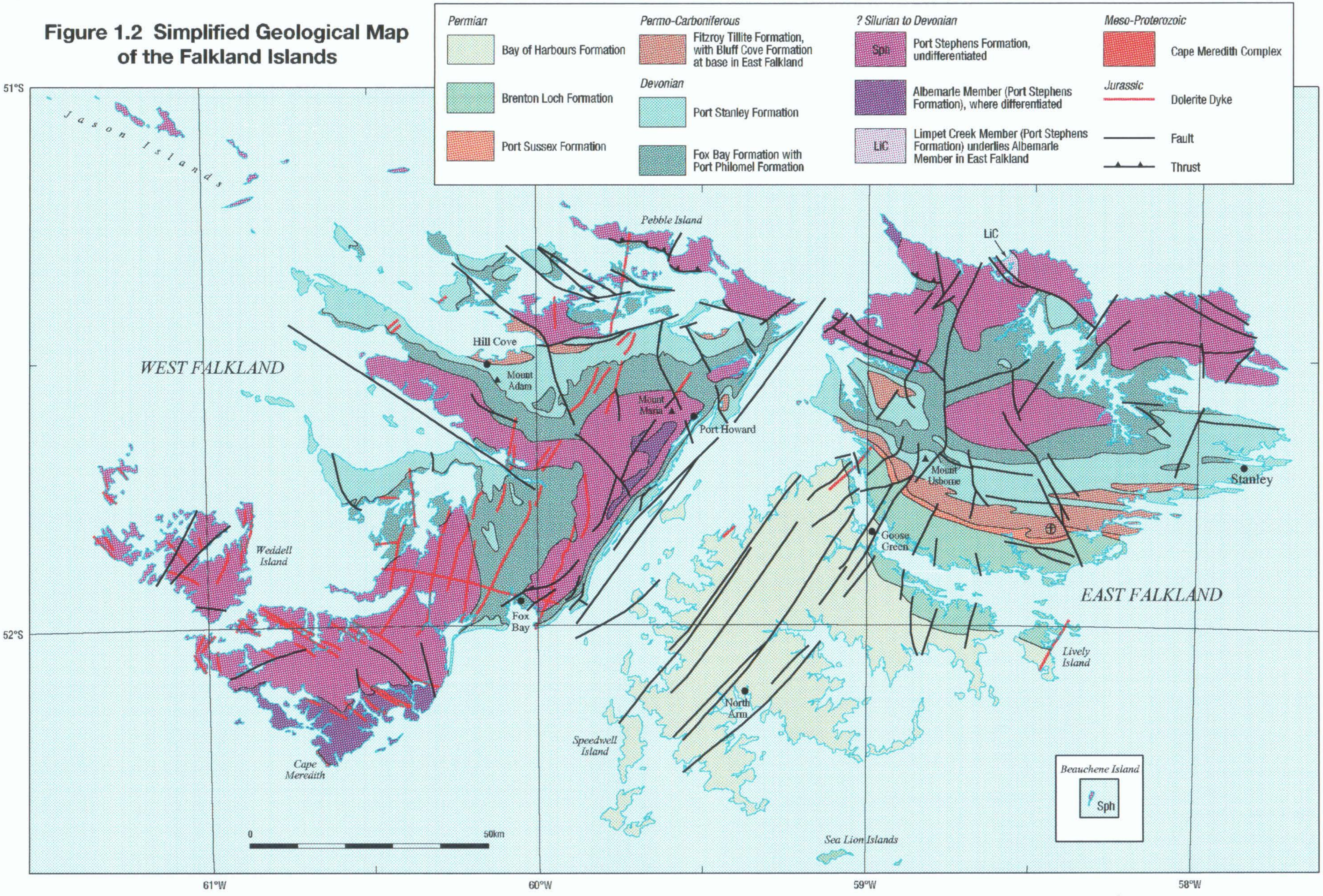


Table 1: LITHOSTRATIGRAPHY OF THE FALKLAND ISLANDS

		Approximate thickness in metres
LAFONIA GROUP (Late Carboniferous to Permian)		
Bay of Harbours Formation	Sandstones, siltstones and mudstones	3500
Egg Harbour Member		
Brenton Loch Formation		3000
Saladero Member	Alternating packages dominated by	
Cantera Member	turbidite sandstones or by laminated	
Terra Motas Member	siltstone/mudstone couplets	
Port Sussex Formation		250
Shepherds Brook Member	Siltstones, fine sandstones and mudstones	
Black Rock Member	Mudstones, some organic	
Hells Kitchen Member	Mudstones and bedded diamictites	
Fitzroy Tillite Formation	Massive diamictite, minor sandstone and mudstone	750
Quark Pond Member	Mudstones and bedded diamictites	5
Bluff Cove Formation	Fine sandstones, siltstones and mudstones	0 - 200
WEST FALKLAND GROUP (?Silurian to Devonian)		
Port Stanley Formation	Sandstones (some quartzitic), mudstones	450
Port Philomel Formation	Sandstones, siltstones and mudstones, some with plant fossils	250
Fox Bay Formation	Fine sandstones, siltstones and mudstones, some with marine invertebrate fossils	800
East Bay Member	Fine to medium sandstones, some siltstones and mudstones	250
Port Stephens Formation		
Fish Creek Member	Mainly subarkosic arenites; with large-scale cross bedding; no trace fossils	350
South Harbour Member	As Fish Creek Member	600
Mount Alice Member	Mainly quartz arenites; alternately cross-bedded, or plane-bedded with *Skolithos* and other trace fossils	700
Albemarle Member	Mainly quartz arenites; typically plane-bedded, with common *Skolithos* and other trace fossils	800
Limpet Creek Member (east only)	Fine to medium sandstones	800
Plantation Member (west only)	Arkosic sandstones and red-coloured siltstones and mudstones	10
CAPE MEREDITH COMPLEX (Proterozoic)	Granitoid gneisses, paragneisses and mafic dykes	

rotation of older structures either side of Falkland Sound. The fourth phase is represented by one or more major SSW-verging thrust sheets in the north of the Islands. The final phase gave rise to a series of WNW through to ENE-trending extensional faults in the north of the Islands, at least in part reactivating major thrusts and other structures. The eastern part of East Falkland was uplifted by several kilometres relative to the rest of the Islands.

Only one onshore deposit is known from the period between the Jurassic and the late Quaternary: the Neogene West Point Forest Bed, which comprises tree trunks and other plant material in organic clay. Superficial deposits of late Quaternary age (younger than about 50 000 years) are widespread and varied. Solifluction deposits are almost ubiquitous on the higher ground, generally together with stone runs, and minor blockfields and scree. Landslips have occurred in a few places. Closed ramparted hollows (pingo scars) occur locally, demonstrating that even the low-lying ground was subject to permafrost during the ice age. On the other hand, glacial deposits and landforms are confined to a few of the highest hills. Deposits of wind-blown sand are mostly associated with west-facing beaches and with lakes. Although peaty soils are very common, and thin peat is widespread, both on high ground and in valleys, large peat deposits are relatively restricted. Tussac peat occurs in coastal localities. Most peat is younger than about 14 000 years, but some is known to have formed between 36 000 and 26 000 years ago. Most river valleys are not wide, so alluvium and river terrace deposits are correspondingly restricted. Lakes are numerous and have formed by a variety of processes, including aeolian deflation and thawing of ground ice. Lacustrine deposits are widespread but seemingly nowhere very extensive. Raised beach deposits and raised marine erosion terraces are known from several different levels all around the Islands.

1.3 History of geological research

Geological observation on the Falkland Islands dates from the records of Pernety (1770), a priest and naturalist who accompanied de Bougainville in 1764. Pernety sketched the distinctive anticline on Long Island Mountain [VC 24 82] and described the Prince's Street stone run in flamboyant terms. Charles Darwin is known to have brought a copy of Pernety's book when he visited the Islands in 1833 and 1834 with HMS Beagle. Although Darwin (1845) wrote that the stone runs formed during earthquakes, it appears likely that he was following Pernety's account with misgivings (Armstrong, 1992). Darwin travelled only on East Falkland. Apart from speculating on the origin of the stone runs, he described aspects of the geological structure, and recorded the presence of fossil brachiopods and other marine invertebrates (described by Morris and Sharpe, 1846). He noted the presence of dolerite dykes in West Falkland which were reported to him by Sulivan, at that time a lieutenant on the Beagle (Darwin, 1846).

A scientific successor to HMS Beagle, HMS Challenger visited the Islands in 1876 during a series of pioneering world-wide oceanographic cruises. While the main objective in the Islands was to investigate (mistaken) reports of rising sea-level, the ship's company also discovered lime-rich mud in Stanley Harbour, collected fossil brachiopods and investigated a fictitious report of coal (Tizard et al., 1885, pp. 883-893; Halle, 1912; Greenway, 1972).

Many explorers and scientists bound for the Antarctic or South America have visited the Islands. Newton (1906) described brachiopods collected on the Falklands by the Scottish National Antarctic Expedition of 1904. Professor JG Andersson, the geologist attached to the Swedish Antarctic Expedition, was able to travel widely in the Islands during 1902. Passing Cape Meredith by schooner, he was the first to observe the presence of a crystalline basement in the Islands. Amongst other disparate observations, his report (Andersson, 1907) includes some splendid engravings of hummocky cross-stratification in the Fox Bay Formation at Port Louis. He deduced the importance of solifluction as a geological agent, both in the Falklands and elsewhere (Andersson, 1906). Nathorst (1906) described some of the plants collected by Andersson, revealing the presence of the Gondwanan *Glossopteris* flora. Halle (1912), a geologist with the Swedish Magellanic Expedition, considerably expanded Andersson's account. He recognised the presence of the Permo-Carboniferous tillite, another characteristic component of Gondwanan geology, and described the West Point Forest Bed in detail. He noted the resemblance of the main geological formations to those found in South Africa.

Clarke (1913a; 1913b) reviewed Early Devonian marine invertebrate fossils from Brazil, comparing them with collections from Bolivia, Argentina, the Falkland Islands and South Africa, which together comprise the 'Malvinokaffric Province'. His material includes some fine specimens collected by Mrs Constance Allardyce, wife of the Governor at that time. He concluded that the faunas from these diverse areas showed a remarkable homogeneity, distinct from those from the northern hemisphere.

The earliest systematic attempt to create a geological map of the Islands was by the first Government Geologist, Herbert Arthur Baker. Baker arrived in the Falklands on Christmas Day 1920. During that and the following austral summer he travelled throughout the Islands by steamer, schooner and horseback, undertaking 'a comprehensive survey for coal, oil and other minerals' and making a 'collection of palaeontological and petrological specimens', in the course of which he obtained 'data for the construction of a geological map of the islands'. While the geological reconnaissance of Andersson (1907) and Halle (1912) guided his efforts, Baker accomplished this work without the benefit of a detailed topographic map of the interior. Given the circumstances of its production, his report (Baker, 1924) is a remarkable document. He deduced that the thick arenitic sequences of the Islands are separated into two by the fossiliferous Fox Bay Formation; and the stratigraphic succession which he described for the Islands remains broadly valid. However, the map which accompanies the report is little more than a sketch in which even the major geological structures can barely be discerned. Seward and Walton (1923) discuss plant fossils collected by Baker and by Halle.

All this work was used to great effect by the great South African geologist, A L du Toit, one of the pioneers of the theory of continental drift and of research into the existence and configuration of the past southern supercontinent, Gondwana. In comparing the geology of South America and South Africa, Du Toit (1927; 1937) concluded that the geology of the Falklands resembled the latter more closely that the former. His reconstruction of the past positions of these two 'wandering continents' placed the Falkland Islands between them.

In the following years, scientists of the British Antarctic Survey (BAS) briefly investigated aspects of the geology of the Islands. Joyce (1950) studied the stone runs. Hattersley-Smith and Hamilton (1950) described an accumulation of sub-fossil bird bones in the peat of West Point Island. Adie (1953) reviewed the evidence for past sea level changes in the Islands, describing a shell limestone at Fitzroy in detail. Barton (1958) made a geological map of the Stanley area. Ashley (1961) carried out the first geophysical work in the Islands, a ground magnetometer survey around Stanley (Mansfield, 1965). Brown (1967) made a detailed description of a Falklands dolerite, drawing comparisons with dolerites from other parts of Gondwana. McNaughton (1972), Martin and Sturgeon (1982) and McGibbon (1988) established a gravity base station in Stanley and made a limited number of gravity measurements at widely spaced localities around the Islands.

Adie (1952a; 1952b) also reviewed the geology of the Islands, emphasising the links with South Africa. He largely confirmed du Toit's findings but realised that the closest correlation between the geological formations in the Falklands and those in South Africa would be made if the Islands once lay to the east of South Africa in a reconstructed Gondwana, completely inverted relative to their present position. This bold interpretation has been substantiated by subsequent work.

A detailed investigation of the Fitzroy Tillite Formation was carried out by Frakes and Crowell (1967), one of a series of studies by them of the Permo-Carboniferous tillites of the southern hemisphere. Dawson (1967), who accompanied Frakes and Crowell, geologically surveyed the Bluff Cove area.

In 1968, Sir Cosmo Haskard, the Governor of the time, provided the stimulus to produce a new geological map of the Islands, most probably as a possible avenue for broadening the economy (Aldiss, 1997). Following his initiative, Mary Greenway produced a new geological map of the Islands by photogeological methods, without any additional fieldwork. Amongst other advances, her work (1970 to 1972) showed clearly the general distribution of the major lithostratigraphic units, revealing the overall geological structure of the Islands. It also showed the general extent of the dolerite dykes, and that these followed at least two major orientations. Greenway's map was an immeasurable improvement over Baker's and together with her report (Greenway, 1972) has remained the fundamental reference on Falkland Islands geology until now. Nevertheless, the limitations of this photogeological interpretation must have become apparent even as the work proceeded. Critical appraisal of Greenway's map reveals its uncertainties about some aspects of the bedrock geology.

Substantial work on the superficial geology of the Islands took place in the 1970's. Clapperton (1971), and Clapperton and Sugden (1976) described evidence for glaciation within the Islands. Clark (1972) made a detailed and comprehensive review of the periglacial landforms in the Falklands, elaborated upon by Clapperton (1975) and Clark (1976). Weller (1975) investigated the geomorphology and biodiversity of ponds located in peat. He published a radiocarbon date on bird bones found in the peat on West Point Island. Research on the Quaternary deposits of the Islands was undertaken in the late 1970's by Roberts (1984), with particular emphasis on glaciation and raised beach levels (Clapperton and Roberts, 1986; Clapperton, 1990), and the radiometric age and palynology of peats (Birnie and Roberts, 1986).

There was also a series of visits to the Islands by Argentine geologists during the 1970's. Cingolani and Varela (1976) made a reconnaissance survey of the Cape Meredith area. They were the first to carry out radiometric age determinations, demonstrating the mid-Proterozoic age of the Cape Meredith Complex, and an Early Jurassic age for a single dolerite dyke. Jalfin and Bellosi (1983) studied the Permian sequence in the Goose Green area. Bellosi and Jalfin (1984b) investigated the Lafonia Group between Swan Inlet and Ceritos, and performed morphometric analysis of a nearby stone run (Bellosi and Jalfin, 1984a). Scasso and Mendia (1985) carried out similar detailed stratigraphic studies around Shag Harbour, Cape Meredith and Stanley.

Instrumental analysis became more commonplace during the 1980's. The age of the Cape Meredith Complex was confirmed by Rex and Tanner (1982). Palaeomagnetic analysis of dolerite dykes by Mitchell et al. (1986) and Taylor and Shaw (1989), supported by geochemical and petrological studies of the dolerites (Mitchell, 1988) confirmed Adie's (1952a) proposal for the inverted position of the Falkland Islands in a reconstructed Gondwana.

Requirements for construction projects in the 1980's and 1990's, most notably for the Mount Pleasant Airport Complex (1983-1986), led to a number of geotechnical papers (Rosenbaum, 1984; Rosenbaum, 1985; Kenrick, 1987; Skene and Brice, 1987) and unpublished geotechnical reports.

Although preceded by some seismic work in the 1950's and by three Deep Sea Drilling Programme boreholes on the Maurice Ewing Bank in 1974, hydrocarbons exploration in the Falklands area can be said to have started in the late 1970's, when there were two speculative regional seismic surveys offshore (Richards et al., 1996a). An onshore oil exploration licence was granted to Firstland Oil and Gas plc in 1984, but little work was carried out. Two more offshore reconnaissance seismic surveys were made in 1993 and 1995. The results of the seismic surveys led to the licensing of seven tranches for commercial exploratory work in the North Falkland Basin (Richards and Fannin, 1994; Richards et al., 1996a; Richards, 1997). Further geophysical surveys in the tranche areas were followed by the drilling of six exploration boreholes in 1998.

With the growing interest in the possibilities offered by hydrocarbons exploration in Falkland Islands waters, research was carried out on diverse aspects of the onshore geology relevant to the hydrocarbons industry (Storey et al., 1995; Geochem Group Ltd, 1996; Macdonald et al., 1996; Macdonald, 1996). Marshall (1994b) performed stratigraphic studies of selected sections through the Devonian sedimentary sequence, carrying out vitrinite reflectance and palynological analysis. He was able to demonstrate that no part of the West Falkland Group is younger than the Devonian, and to support the Gondwana reconstructions proposed by Adie (1952a) and Mitchell et al. (1986). Marshall (1994a) investigated the thermal maturity of Permian mudstones, which are the only probable hydrocarbons source rock onshore. He found that they are overmature in northern East Falkland, but that some potential remains in southernmost East Falkland and West Falkland. Data in Macdonald (1996) show that the Permian is also overmature in southern East Falkland.

Other research more generally relevant to the structural evolution of the Falklands was also carried out on the Islands during the 1990's. Mussett and Taylor (1994) provided more geochronological information which confirmed the age of the dolerite dykes as about 190 Ma, in the Early Jurassic. In 1994, R J Thomas and J Jacobs spent three weeks mapping the Cape Meredith Complex (Thomas and Jacobs, 1995; Thomas et al., 1997) and sampling it for detailed geochronological studies (Jacobs et al., 1999). Thistlewood (1993) recorded sedimentological details of several key areas to support palaeomagnetic studies. Thistlewood et al. (1997) and Thomas et al. (1998) presented petrological, geochemical and geochronological data on the dykes intruding the Cape Meredith Complex. Curtis (1994), Curtis and Hyam (1998) and Hyam (1997; 1998) carried out structural studies in parts of East Falkland, and in the Bold Cove and Carcass Bay areas of West Falkland. They provided further strong support for the previously inverted position of the Falkland Islands in Gondwana. Hyam et al. (1997) described mid-Carboniferous and Permo-Carboniferous sedimentary dykes from the South Harbour, Bluff Cove and Green Patch areas.

Research on fossils collected from Falkland Islands and held in overseas museums has been carried out by Boucot and Gill (1956) (brachiopods), Cramer et al. (1972) (palynomorphs), Hiller (1987) (brachiopods), Edgecombe (1994) (trilobites), Cocks (1996) (brachiopods) and Cocks et al. (1998) (brachiopods, trilobites, bivalves).

Quaternary studies on the Islands revived with the Cumbria and Lancashire Falklands Expedition of 1989, which studied aspects of bedrock and superficial geology, geomorphology, and ecology (Clark et al., 1990). A similar second expedition took place in 1994 (Clark et al., 1994) and a third in 1997 (Clark and Wilson, 1997). Several published papers arose from this work. Clark and Wilson (1992) discussed the significance of ventifacts found at several localities. Their radiometric age determinations of peat showed that climatic amelioration at the end of the last glaciation took place prior to about 14 000 years ago. Wilson et al. (1993) discuss soil erosion and related aspects of geomorphological evolution in the Falkland Islands. Wilson (1994; 1998) analysed the formation of post-glacial deposits of aeolian and alluvial sands near Blue Mountain and Lake Sulivan, with respect to climate variation. Wilson (1995) describes some unusual forms of patterned ground which are unrelated to frost. Clark et al. (1998) describe a rare occurrence of peat pre-dating the solifluction deposits which formed during the last cold phase of the Late Pleistocene glaciation. This peat yielded radiometric ages of between 36 000 and 28 000 years before present, and is the oldest Quaternary deposit so far discovered onshore. In independent work, Macphail and Edwards (in prep) interpret new palynological data to suggest that the West Point Forest Bed is of Neogene age.

As geological research on the Islands proceeded in the early 1990's, the deficiencies of the existing maps became increasingly apparent. There was a growing realisation within the Falkland Islands Government not only that its own knowledge of the geological structure and mineral resource potential of the Islands was based on inadequate information, but also that further research work was being hindered by the lack of a more detailed and accurate map. This led to the Falkland Islands Geological Mapping Project (1996-98) (Appendix 1 and Aldiss, 1997).

Geological research on the Islands continuing at the time of writing includes mineral exploration (Cambridge Mineral Resources plc), sedimentological studies in the West Falkland Group (British Antarctic Survey) and the Lafonia Group (University of Aberdeen), structural studies in northern East Falkland (University of Birmingham), Permian trace fossils, and the geochemistry and petrology of dykes (University of Aberdeen), and tephrochronology and other Quaternary studies (University of Ulster at Coleraine).

Much research from the work of du Toit onwards has been devoted to demonstrating the similarities of the geology of the Falkland Islands with other parts of the southern continents, especially that of South Africa. Papers by Adie (1952a), Mitchell et al. (1986), Marshall (1994b) and Curtis and Hyam (1998) have demonstrated the correlations which place the Falkland Islands within Gondwanaland in a position close to Port Alfred, in south-eastern South Africa (Section 6). Nevertheless, contributions by Thomas et al. (1997) and Hyam (1997; 1998) show that uncertainties remain in the reconstruction. It seems that a new phase of research may be starting, in which the significance of the *differences* between the sequences and structures in the Falkland Islands and adjoining parts of South Africa and Antarctica can be investigated. This should further constrain the pre-Jurassic position of the Islands relative to both adjacent continents and to the structures that developed during the break-up of Gondwana. For example, Jacobs et al. (1999) have found that the gneisses of the Cape Meredith Complex are somewhat younger than their equivalents in the Natal Metamorphic Complex, but are of a similar age to those in Western Dronning Maud Land (East Antarctica). Conversely, the correlative basement in East Antarctica was metamorphosed at about 500 Ma, whereas the Cape Meredith and Natal basement complexes were not. It is hoped that the new information about the geological formations of the Falkland Islands and their structure presented in this report and the accompanying maps will contribute to this process and will form a sound basis for new geological research in the Islands.

1.4 The geological maps and report

This report complements geological maps of the Falkland Islands prepared during the Falkland Islands Geological Mapping Project (1996-1998). A map of the solid geology at 1:250 000 scale was compiled from 1:50 000 scale manuscript maps and published in two sheets (Appendix 1, Aldiss and Edwards, 1998). The geological information shown on the 1:50 000 scale compilations is available in digital form, together with a digital map depicting the superficial deposits of the Islands at 1:250 000 (Section 5), and digital images of the 1:250 000 solid geological maps. The original compilations and other project records are held by the Department of Mineral Resources, Falkland Islands Government, Stanley.

In large part the rock formations of the Falkland Islands are well-bedded and structurally simple. Considerable areas have only thin or patchy superficial cover. Together with the lack of trees these are very favourable conditions for geological interpretation of aerial photographs, and many dykes, upright folds and high-angle faults can be identified photogeologically. Note, however, that reclined folds, bedding-parallel faults, and low-angle faults are generally difficult to identify on aerial photographs. Note also that the visibility of bedding (and of dolerite dykes) on aerial photographs depends on the amount of differential erosion and on the amount of superficial cover.

Two kinds of symbol are used on the face of the map to indicate the angle of dip of bedding. A strike bar with simple cross-tick, accompanied by a number, shows where bedding orientation has been recorded directly by a geologist on the ground. Similar symbols incorporating an arrow head together with either one, two or three cross-ticks, show where the bedding has been estimated at a distance. In most cases, this is an interpretation from aerial photographs viewed stereoscopically. The dip direction in bedded strata is generally easy to determine photogeologically, either from direct observation of large exposed bedding planes or more usually from the configuration of associated landforms. However, bedding dips are notoriously difficult to estimate accurately from aerial photographs. Therefore, only the approximate *relative* amount of dip is indicated in most places. The broad categories of gentle, intermediate and steep do not represent the same ranges of dip values throughout the Islands. In the south of West Falkland, a dip of 10° can be regarded as moderate, but in northern East Falkland, the 'gentle' dip symbol may indicate true dips up to 15° or 20°. In some places, typically on wave-cut platforms, the strike of bedding is clearly visible on aerial photographs, but there is insufficient topographic relief for the dip direction to be determined, even in stereoscopic view. In a few places, it is clear from the aerial photographs that bedding closely approximates to the topographic slope, for example in the Dunnose Head area [TC 25 26]. It is then possible to estimate the angle of dip from the topographic map.

On the ground, the dip of bedding exposed in cliffs or other rocky outcrops can commonly be estimated when seen at some distance. These estimated dips may be less than the true dip, but given the tendency for strike to be very consistent over distances of some kilometres in many parts of the Falklands, this is felt to be a reasonable method to use in a reconnaissance survey.

The horizontal sections shown in the margins of the two 1:250 000 scale map-sheets were drawn at the size at which they appear, from information taken from the 1:50 000 scale manuscript compilations. In order to accentuate some of the geological and topographic features, these sections are vertically exaggerated by a factor of five. Consequently, dips are exaggerated by five times the tangent of the angle. Thus a bedding plane with a true dip of 20° appears on the section to be dipping at about 60°. Where the dip of a layer changes from steep to gentle, its thickness apparently increases as a consequence of this distortion.

With a very few exceptions, place-names used in this report are those appearing on the current version of the 1:50 000 scale topographic maps of the Falkland Islands. In some instances these may differ from those which appear on the 1:250 000 scale sheets. The origin of many Falklands place-names is given by Munro (1998).

The word 'camp', derived from the Spanish '*el campo*' (the countryside), is used in the Islands to denote the area outside settlements. In addition, the major subdivisions of individual farms are known as 'camps', whose names are shown on the 1:50 000 topographic maps in small uppercase letters. In many instances, camps are named after a local geographic feature (e.g. Main Point). In this report, due care has been taken to avoid ambiguity, but where a place-name has not been otherwise qualified, it should be assumed that reference is made to the geographic feature rather than to an eponymous camp.

Note that many of the larger inlets in the Falklands coastline are named as 'ports', for example Port Purvis, Port Harriet, Port Pleasant. In some instances, such as Port Howard, the same name is used for the adjacent settlement, but in most the name does not imply that any berthing facility ever existed. In present usage in the Islands, 'Port Stanley' is taken to be the whole of the inlet on which the town of 'Stanley' lies. 'Stanley Harbour' is commonly used for that part of Port Stanley where ships are usually moored.

Where compass bearings are given to describe the orientation of geological structures, these are clockwise relative to grid north (so that N270° indicates 'due west'). Grid references are quoted in the form 'Public Jetty, Stanley [VC 4105 7282]', or 'the eastern part of the town of Stanley [VC 41 72]', and so forth. These refer to the UTM Grid Zone 21 as printed on 1:50 000 scale topographic maps of the Islands. In some instances, the letter code forming the first part of the full grid reference is omitted. *Permission from the relevant landowner or manager must always be sought before visiting sites mentioned in this report. The description of a site in this report should not be taken as an indication that such permission will be granted.*

The reader should note that some words which are in general use in English take on a specific meaning when used in a geological context. For example, 'strike' refers to the orientation of a horizontal line drawn on a bedding plane, or other planar surface. In this report the noun 'outcrop' refers to the area of a given geological formation as shown on the geological map, whereas 'exposure' indicates a place where the bedrock can actually be seen at the surface. The corresponding verb is 'to crop out'. In other words, outcrops crop out at exposures.

2. PRE-CENOZOIC STRATIGRAPHY

2.1 Introduction to the pre-Cenozoic stratigraphy

In describing the geology of an area, a distinction is commonly made between the 'solid geology', which broadly corresponds to 'bedrock', and the 'superficial geology' (or 'drift' geology), which encompasses material lying between the soil and the bedrock. Solid geological formations generally extend to depth and form the geological framework. In the Falklands they entirely pre-date the Cenozoic Era. Four major divisions are represented (Figure 1.2): the Mesoproterozoic Cape Meredith Complex (Section 2.2), the ?Silurian to Devonian West Falkland Group (Section 2.3), the Carboniferous to Permian Lafonia Group (Section 2.4) and Jurassic minor igneous intrusions (Section 3). The structure and metamorphism of the solid geological formations are described in Section 4. Superficial deposits tend to obscure the underlying geological framework. In the Falklands nearly all the known superficial deposits are less than 50 000 years old: they are described in Chapter 5. The geological formations known to occur onshore in the Falkland Islands are summarised in Table 1.

The solid geological map (Aldiss and Edwards, 1998) shows lithostratigraphic divisions which are named according to international conventions on lithostratigraphic nomenclature (Eysinga, 1970; North American Commission on Stratigraphic Nomenclature, 1983; Whittaker et al., 1991). Where possible, existing stratigraphic names have been used (with any adaption necessary for use in a lithostratigraphic scheme), so granting precedence and avoiding proliferation. In some cases an existing name is used here in a slightly different stratigraphic concept to that of its author. Such changes should cause little confusion, partly because most units were originally very loosely defined. New names have been chosen either to designate newly recognised units or to resolve conflicts between the existing nomenclature and conventions of stratigraphic nomenclature. The stratigraphic names incorporate place-names in common use in the Falkland Islands at the time of writing and which appear on currently available topographic maps. Spellings are those shown on the current version of the 1:50 000 scale topographic maps.

The lithostratigraphic rank (i.e. group, formation, member, bed) to which each unit is assigned reflects the authors' understanding of the stratigraphy of the Falkland Islands from the viewpoint of a mapping geologist. The divisions reflect apparently natural groupings of parts of the sedimentary sequence which can be observed in the field or on remote images. They do not depend on the results of biostratigraphic or mineralogical analysis (although in some instances they have been substantiated by the results of such analysis), nor on an interpretation of the environment of deposition of the units, nor on their interpretation in terms of sequence stratigraphy. The lithostratigraphic ranks adopted in equivalent sequences in the surrounding continents have not been allowed precedence in attribution of rank in the Falkland Islands.

In future, the lithostratigraphic rank of these units may change. Therefore the symbol used on the geological map to denote each unit does not include any component to indicate rank. Note that these symbols may therefore be considered unsuitable as abbreviations in running text. For example, although the Cape Meredith Complex is designated 'CMr' on the map-face, 'CMC' is used in running text.

Descriptions of each stratigraphic unit in this report designate a type area and one or more reference sections, including where possible a basal stratotype. These reference sections include those which seem most suitable to be adopted as type sections, or as parts of composite type sections.

People wishing to undertake geological fieldwork should be aware that in the Falkland Islands permission to enter land for that purpose should always be obtained beforehand. Designation of a type area, type section or reference section should not be taken to imply that such permission will be granted. The names of relevant landowners or managers can be obtained from the Department of Mineral Resources, Stanley.

The base of some stratigraphic units, such as the Bay of Harbours Formation and most members of the Port Stephens Formation, have been defined here primarily according to a contrast in terrain type, or 'texture' appearing on aerial photographs and satellite images. This differs from the normal practice of defining the base of a unit according to a specific change in the lithological sequence, usually one that is perceived to be sedimentologically significant. However, in the particular circumstances of the Falklands, where inland exposure of bedrock is rare, a unit boundary must have a reasonably distinctive photogeological expression in order for that boundary to be traced away from the coast. Except where superficial cover is particularly thick, contrasts in topography will reflect lithostratigraphic changes in the sequence, even though the nature of the change may not be fully known. Unit boundaries based on large-scale variations expressed in topography are likely to correspond to changes visible in geophysical borehole logs.

Not all the stratigraphic units described in the following sections appear on the geological map. Some are too thin to be shown at 1:250 000 scale, and in some instances cannot be shown in their correct proportions even at 1:50 000 scale. Some cannot be delineated sufficiently reliably in some parts of their inferred outcrop. The Plantation Member (Section 2.3.1.1) is thin and crops out only in steep ground near Cape Meredith. Even the outcrop shown in the 1:25 000 scale inset map is somewhat exaggerated in width. No mappable boundary between the Mount Alice Member and the South Harbour Member has been identified and so these units are shown by the same colour on the map, but with different symbols to indicate their approximate distribution (Section 2.3.1.5). The other members of the Port Stephens Formation cannot everywhere be separated and so in some areas are shown together as Port Stephens Formation (undifferentiated). The East Bay Member is probably more extensive in West Falkland than shown, but outside its type area it cannot at present be separated from the rest of the Fox Bay Formation with confidence (Section 2.3.2.1). Strata exposed on Beauchêne Island resemble parts of both the Port Stanley Formation and the Port Stephens Formation but at present cannot be correlated with either with confidence (Section 2.3.5).

The Quark Pond Member is less than 10 metres in thickness (Section 2.4.2.1) and so is shown on the map as a single line within the outcrop of the Fitzroy Tillite Formation. It has also been observed in the Bold Cove and Black Rock House areas, and probably continues through the intervening ground, but its outcrop cannot be traced with confidence. The Hells Kitchen Member (Section 2.4.3.1) is thin and occurs everywhere at the base of the Black Rock Member, with which it is shown on the face of the map. On West Falkland, the outcrop of the Port Sussex Formation is too narrow to be divided to show any of the component members.

The two thickest members of the Brenton Loch Formation (Section 2.4.4) cannot be distinguished with confidence in the Goose Green Graben, or to the east. The Terra Motas Member (Section 2.4.4.1) cannot be reliably distinguished east of the Canada Runde area. The base of the Egg Harbour Member (Section 2.4.5.1) is clearly defined only in the north of its outcrop and elsewhere is indicated only very approximately.

2.2 Cape Meredith Complex

The Cape Meredith Complex (CMC) is known only in a coastal section some five kilometres long at the southernmost point of West Falkland (Plate 1). It comprises amphibolites, paragneisses, orthogneisses, granites and pegmatites. The CMC is of Mesoproterozoic age, between about 1120 and 1000 Ma old. It was intruded by lamprophyre and dolerite dykes. These mafic dykes are of several ages, including Jurassic and, probably, Lower Palaeozoic. Rock fragments presumed to have been derived from the CMC or from closely allied terranes occur in the Fitzroy Tillite Formation, and on beaches. The northerly regional dip of the Port Stephens Formation indicates that relatively extensive seafloor outcrops of the CMC probably occur south-west of Cape Meredith and south of Bird Island.

The presence of Precambrian rocks at Cape Meredith was first recognised by Professor J Gunnar Andersson, a geologist of the Swedish Antarctic Expedition of 1902-04 (Andersson, 1907). He observed stratified schists intruded by granite and pegmatite, and noted the unconformity at the base of the overlying sandstones. Baker (1922; 1924) described the main components of his 'Cape Meredith Series' in more detail. Like Andersson, he assigned it to the Archean. His 1924 report includes several photomicrographs and a splendid monochrome photograph from a viewpoint close to that of Plate 1. Adie (1952b; 1958) redesignated this occurrence of the crystalline basement as the Cape Meredith Complex.

During 1994, the CMC was mapped in detail by R J Thomas and J Jacobs (Thomas and Jacobs, 1995). Thomas et al. (1997) recognised a sequence of supracrustal gneisses, the Big Cape Formation, which was intruded by three generations of granitoid. Further descriptions of the CMC are given by Thistlewood et al. (1997). Together with most of the mafic dykes which intrude the CMC, all these components are unconformably overlain by gently dipping sediments of the Plantation Member, where it is present, or the Albemarle Member of the Port Stephens Formation (Sections 2.3.1.1, 2.3.1.3). The appearance of the Cape Meredith Complex on aerial photos is not very distinctive but the base of the overlying sediments can be traced, given adequate ground control.

Plate 1: Coast near Cape Meredith

Photograph by Don Aldiss, BGS. MN28232

View south-east towards Meredith Hill and adjacent coastline. Lower part of cliffs formed by granite gneisses of the Cape Meredith Complex (contact with Big Cape Formation is out of sight, directly below viewpoint). The unconformity at the base of the overlying Port Stephens Formation dips to the left, overlain by reddish-purple strata of the Plantation Member, then by plane-bedded quartz arenites of the Albemarle Member.

Plate 2: Granite gneiss of Cape Meredith Complex

Photograph by Don Aldiss, BGS. MN28233

Foliated granite (G2 granite gneiss) of the Cape Meredith Complex, with two pegmatite sheets and a dark enclave, probably part of a disrupted syn-plutonic microgranitoid dyke.

The Big Cape Formation occurs at the north-western end of the outcrop of the Cape Meredith Complex, on Big Cape camp, and as rafts within the granitoids. Four lithological units are recognised in the Big Cape Formation (Thomas et al., 1997). These are amphibolites with subordinate felsic gneisses, grey biotite-hornblende gneisses, and minor metasedimentary gneisses, including calc-silicates and metapelites. The calc-silicates (quartz-plagioclase-diopside-titanite) and metapelites (garnet-biotite-sillimanite-feldspar) form thin, discontinuous layers and lenses within the amphibolites and grey gneisses. They represent sedimentary rocks deposited within a predominantly volcanic sequence. Geochemical analysis shows that the composition of the amphibolites is basaltic, that of the grey gneisses, andesitic, and the felsic gneisses, rhyolitic. (Thomas et al., 1997) estimate a total apparent thickness of about 600 metres for the Big Cape Formation. They note that there is no indication of the older basement on which this sequence was originally deposited.

The earliest granitoids, termed G1 by Thomas et al. (1997), are foliated medium-grained grey biotite granodiorite sheets up to three metres wide, which intrude the Big Cape Formation. They are locally folded within the sequence, suggesting that they were emplaced during deformation.

Granitic gneiss (G2) is the most extensive unit in the CMC. It intrudes the Big Cape Formation, with the main contact lying concordant to foliation. It is predominantly a coarse-grained, grey or pink, K-feldspar-rich biotite granite gneiss, essentially a deformed granite. It contains generally sparse but variable proportions of K-feldspar megacrysts, and a small area of augen gneiss occurs east of Cape Meredith. There is no marked segregation of mafic and felsic components in the granite gneiss. Observations during the present survey found that it locally includes elongated, weakly foliated mafic enclaves, a few centimetres thick but up to 80 centimetres long, some of which also contain K-feldspar megacrysts. These are interpreted as remnants of syn-plutonic dykes (Plate 2).

Leucogranite (G3) forms the south-eastern part of the outcrop and intrudes granitic gneiss as multiple sheets and dykes. It is a pink, K-feldspar-rich, highly leucocratic garnetiferous granite, locally with primary igneous layering. It includes straight but diffuse-margined pegmatite veins and pods up to 10 centimetres thick and a few metres long. Thistlewood et al. (1997) observed large rafts of augen gneiss enclosed within the granite.

The older granites and the Big Cape Formation are intruded by numerous simple pegmatite veins and dykes up to four metres wide. A thick pegmatite sheet which intruded the contact between the G1 granite gneiss and the Big Cape Formation contains enclaves of foliated amphibolite. In addition to quartz and feldspar, these pegmatites contain biotite, muscovite and very rare tourmaline (Thomas et al., 1997). Baker (1924) noted the occurrence of accessory xenotime in a pegmatite from Cape Meredith. Some veins and dykes of pegmatite display graphic texture. These contain mica 'books' up to three centimetres across.

A pervasive tectonic foliation is found in all these units except for the G3 granite and the pegmatites. According to Thomas et al. (1997), in most of the CMC the foliation dips consistently to the north or north-west at between 40° and 85°. It typically is associated with a mineral stretching lineation which plunges in a westerly direction at low to moderate angles (Thomas et al., 1997). Deviations from this general pattern occur in coarse megacrystic granitic gneiss about one kilometre east of Cape Meredith, where the foliation has been tightly folded about an axis plunging WSW at about 35° within a subvertical axial plane. Thistlewood et al. (1997) also note south-dipping foliations in the south of the outcrop. They describe a subvertical 15 metre-wide ductile shear zone, trending roughly east-west, just east of Cape Meredith. Mineral assemblages in the Big Cape Formation show that metamorphism reached the amphibolite facies (Thomas et al., 1997).

Four mafic dyke suites that intrude the CMC were described by Thistlewood et al. (1997), designated Groups A to D. All post-date the tectonic fabrics and all except D pre-date the Port Stephens Formation. Although it is convenient to describe them in this section, these dykes are not strictly part of the CMC.

Group A comprises lamprophyre dykes between 25 centimetres and 250 centimetres thick, dipping towards the north-east at between 8° and 40°. Locally they were emplaced along pre-existing faults. These lamprophyres are petrographically diverse, including pyroxene-rich spessartite, minette and vogesite (Thistlewood et al., 1997) but their geochemistry suggests that they are co-genetic (Thomas et al., 1998). Thomas et al. (1998) show that the Group A dykes post-date those of Group B.

Group B dykes are strongly altered fine-grained dolerites which cross-cut both G2 and G3 granites. They are less than two metres wide with chilled margins, and are vertical or subvertical, striking NW-SE or north-south (Thomas et al., 1998; Thistlewood et al., 1997).

Group C dykes are olivine dolerites with some alteration. They cross-cut the G2 granite and also dykes of Groups A and B. They strike east-west, dipping at angles of 68°-80° to the north (Thomas et al., 1998; Thistlewood et al., 1997).

Group D dykes are relatively fresh dolerites, ranging from one metre to 13 metres across. Like the Group B dykes they have chilled margins. Within the CMC they form intersecting linear segments along joints, giving the dykes a zig-zag appearance. They are taken to be part of the Jurassic dyke swarms which are widespread in West Falkland (Section 3.1).

Thistlewood et al. (1997) present a range of geochemical data for these dykes. Thomas et al. (1998) present some additional analyses in support of their argument that the Group B dolerites of Thistlewood et al. (1997) pre-date the Group A lamprophyres.

Erratic pebbles, cobbles and boulders occur in the Fitzroy Tillite (Section 2.4.2) and on beaches, including some on raised beaches (Section 5.2.10). While much of this detritus can be matched with components of the CMC, some is composed of rock-types not seen onshore, but which could nevertheless form part of the CMC offshore. For example, beach pebbles composed of mylonite have been found near Cape Meredith. It is also possible that the tillite includes material carried considerable distances, perhaps from what are now terranes in East Antarctica, closely allied to the CMC (Section 2.4.2). For example, small rounded boulders of garnet gneiss occur in the tillite at Hill Cove. Petrographic examination shows that this rock had been subjected to more than one phase of metamorphism and had developed incipient mylonitisation. A complex mineral assemblage including orthopyroxene and sillimanite suggest very high grade metamorphism, probably granulite facies, of a pelitic precursor containing quartz. This was followed by shearing and retrogression (Hards, 1997b). No rocks similar to this occur in the CMC but are found in East Antarctica and South Africa.

Age of the Cape Meredith Complex and associated dykes

The probable mid-Proterozoic age of the CMC was first recognised by Cingolani and Varela (1976) who reported K-Ar ages of about 1100 Ma for hornblende from three amphibolites, and an Rb-Sr 'isochron' of about 980 Ma for an assemblage of granitoid rocks. Rex and Tanner (1982) performed a K-Ar radiometric age determination of hornblende from an amphibolite, finding an age of 977 ± 40 Ma. Biotite from a felsic gneiss gave an age of 953 ± 30 Ma. These were regarded as minimum ages.

Thistlewood et al. (1997) report Sm-Nd data on a garnet-whole rock pair representing the pink leucogranite (G3) which give an age of 992 ± 50 Ma for garnet growth.

Jacobs et al. (1999) carried out a series of U-Pb determinations on zircons from the CMC using the very reliable SHRIMP technique. They interpreted a determination of 1118 ± 8 Ma for felsic gneisses from the Big Cape Formation as the age of extrusion of the volcanic sequence from which the gneisses were formed. Zircon cores in the G2 granite dated at 1135 ± 11 Ma are thought to have been inherited from an older portion of the Big Cape Formation. The G1 gneisses gave SHRIMP ages of about 1090 Ma, the G2 granite gneisses 1067 ± 9 Ma and the G3 leucogranite 1003 ± 14 Ma, taken to be the respective ages of intrusion. Metamorphic overgrowths on zircons from the felsic gneisses gave an age of about 1000 Ma, coeval with the G3 granite.

^{40}Ar-^{39}Ar plateau ages of hornblende separated from the amphibolites of the Big Cape Formation of 1009 ± 14 and 1015 ± 6 Ma, along with muscovite (989 ± 3 Ma) and biotite (989 ± 7 Ma) from G3 pegmatites show that the CMC cooled relatively rapidly to below 300°C. There is no evidence for later metamorphism, for example during the Pan-African Orogeny at about 500 Ma.

These results place the igneous and metamorphic events in the CMC at the end of the Mesoproterozoic (1600 Ma to 1000 Ma).

The lamprophyre dykes were dated by Cingolani and Varela (1976) who obtained biotite/amphibole K-Ar ages of 554 ± 25 Ma and 496 ± 15 Ma.

Thistlewood et al. (1997) report whole rock and biotite K-Ar ages for three lamprophyre dykes as 523 ± 9 Ma (for one of the same dykes analysed by Cingolani and Varela, 1976), 473 ± 12 Ma and 306 ± 8 Ma. They note that the K-Ar content of any of these dykes could have been disturbed since formation and so that these dates should be regarded with caution. Indeed, as it appears from field relations that the lamprophyres probably all pre-date the Port Stephens Formation, they must be older than the c. 400 Ma Emsian macrofauna in the Fox Bay Formation, possibly a good deal older. Nevertheless, Thistlewood et al .(1997) took these results to indicate that at least some of the lamprophyres are of Palaeozoic age and that they could have emplaced over a considerable span of time.

Thomas et al. (1998) take issue with this interpretation. They present their own K-Ar data for biotite from two of the lamprophyre dykes, which gave very similar dates of 503 ± 6 Ma and 520 ± 5 Ma. They conclude that the lamprophyres are all of a similar age greater than about 520 Ma.

Thistlewood et al. (1997) found K-Ar whole rock ages of 470 ± 12 Ma and 494 ± 14 Ma for two of the Group B dykes, which they regarded as minimum ages. Thomas et al. (1998) concluded from their own work that the Group B dykes are older than the lamprophyres, but probably not much older than about 520 Ma.

A single Group C dyke yielded a K-Ar whole rock age of 422 ± 39 Ma, which places it in the Silurian (Thistlewood et al., 1997). This seems to place an important constraint on the age of the overlying Port Stephens Formation, but it is also likely to be a minimum age and should be interpreted with caution (cf. Thomas et al., 1998).

In summary, the precursors of the Big Cape Formation were deposited as a volcano-sedimentary sequence between about 1146 and 1110 Ma ago. They were intruded by granodiorites and porphyritic granites and metamorphosed in the amphibolite facies between about 1090 and 1000 Ma. Those syn-tectonic granites were themselves intruded by a late- to post-tectonic leucogranite and pegmatite at about 1000 Ma.

The CMC was intruded by dolerite dykes and then by lamprophyre dykes prior to about 520 Ma. A later phase of dolerite intrusion occurred prior to about 422 Ma, and before uplift and erosion of the basement complex and deposition of the Port Stephens Formation. More dolerite dykes were emplaced during the Jurassic.

It is widely accepted that the Falklands Microplate (Section 6) lay between south-east South Africa and East Antarctica both during the Palaeozoic and the Mesoproterozoic, in the Gondwana and the Rodinia supercontinents, respectively (Wareham et al., 1998; Jacobs et al., 1999). Correlatives of the CMC occur in the eastern Cape Province of South Africa, in Antarctica (both East and West), and probably also on the Maurice Ewing Bank at the eastern end of the Falkland Plateau (Section 6).

The lithologies found in the CMC and their relationships are very similar to those found in the southern part of the Natal Metamorphic Province. If the Falkland Islands are restored to their inferred position prior to Gondwana break-up (Section 6), the orientation of metamorphic fabrics in one area is consistent with that seen in the other (Thomas et al., 1997). Radiometric age determination shows, however, that the Big Cape Formation is significantly younger than its equivalents in the Natal Metamorphic Province, which have yielded ages of about 1240 Ma (Jacobs et al., 1999; Thomas et al., in press). Jacobs et al. (1999) found that the syn- to post-tectonic granites in the CMC and the Natal Metamorphic Province are of similar age, and that the two areas experienced similar low-temperature histories.

The Cape Meredith Complex is also of similar age and composition to rocks in Western Dronning Maud Land (East Antarctica). By contrast to the Cape Meredith and Natal basement complexes, however, much of Western Dronning Maud Land received a later thermal overprint during the c. 500 Ma old 'Pan-African' metamorphism (Jacobs et al., 1999). Thomas et al. (1997) note that the c. 500 Ma dates obtained from the lamprophyre dykes at Cape Meredith suggest the dykes formed during an alkaline magmatic event of 'Pan-African' age which is also represented in the Sverdrupfjella, Dronning Maud Land, but not in South Africa.

The only exposure of Proterozoic rocks in West Antarctica occurs at Haag Nunataks, within a small crustal block to the north-east of the Ellsworth Mountains. It is mainly composed of granodioritic orthogneisses with large concordant rafts of mafic and ultramafic rock. Rb-Sr (whole rock) dating techniques found that these orthogneisses are 1176 ± 76 Ma old. They are cross-cut by post-metamorphic microgranite sills dated at 1058 ± 53 Ma and aplogranite sheets dated at 1003 ± 18 Ma. Similar Mesoproterozoic rocks are thought to continue beneath the

Ellsworth-Whitmore Mountains and like them to have lain between South Africa and East Antarctica within Gondwana (Storey et al., 1994; Curtis and Storey, 1996; Wareham et al., 1998).

Wareham et al. (1998) note that similar basement rock types to those found in the Cape Meredith Complex were intercepted in the DSDP borehole 330 on the Maurice Ewing Bank (Figure 6.1). These gneisses and pegmatites were metamorphosed at about 500 Ma, but their formation age remains unknown.

2.3 West Falkland Group

The West Falkland Group underlies nearly all of West Falkland and the adjacent islands, together with the northern part of East Falkland, and Beauchêne Island. It is dominated by sandstones, some quartzitic, with subordinate siltstones and mudstones.

Baker (1924) recognised two broad divisions in the bedded rocks in the Islands. He divided the older into four units, but gave it no overall name other than 'Devono-Carboniferous Series'. Although Greenway (1972) accepted Baker's subdivisions, she used the informal term 'Devono-Carboniferous Group'. Although this group is represented in both East and West Falkland, it dominates the west. In addition, West Falkland includes the type areas for most of the component formations and their members. 'West Falkland Group' therefore seems to be the most appropriate name to encompass the four formations which were described in concept by Baker (1924) and which are defined in the following sections. Further subdivisions might be recognised in future and by analogy with equivalent sequences in South Africa (parts of the Cape Supergroup), it seems possible that the units described here could eventually be promoted in lithostratigraphic rank.

'West Falkland Group' is synonymous with 'Grupo Gran Malvina' (Borello, 1963; Borello, 1972), which was translated from the original Spanish as 'Gran Malvina Group' by Marshall (1994b). However, this translation employs a name for the island of West Falkland which is not used in the Falkland Islands themselves and which is not shown on maps in current use in the Islands. Therefore 'Gran Malvina Group' is here considered to be an inappropriate name. However, the name 'Gran Malvina' is in general use in South America and it is expected that the term 'Grupo Gran Malvina' will continue to be used in Spanish language publications.

As discussed in the following sections, the age range of the component formations of the West Falkland Group is not closely constrained. The oldest strata are here considered to be most probably of Silurian age, but they could be as young as early Devonian, or as old as Ordovician or possibly even Cambrian. Each formation in the West Falkland Group is thinner in West Falkland than in the east, and the base of the sequence could be older in East Falkland than in the west. The youngest strata were deposited during the latest Devonian.

As discussed in the following sections, correlatives of the West Falkland Group are found in South Africa (parts of the Cape Supergroup), Antarctica (Taylor Group, Beacon Supergroup) and South America.

2.3.1 Port Stephens Formation

The Port Stephens Formation is the thickest and most widespread subdivision of the West Falkland Group and its overall distribution is very similar to that of the group as a whole. It typically consists of medium to very coarse-grained sandstones, quartzites and quartz conglomerates with rounded granules and pebbles of white quartz. Some fine-grained sandstones and minor mudstones are also present. The sandstones vary from pure quartz arenites to subarkoses, with very little mica compared with those of the Fox Bay Formation. The Port Stephens Formation is apparently unfossiliferous except for trace fossils in the middle and lower part of formation, and rare plant debris near the top. The sparse mudstones might contain microfossils.

This formation was originally named the 'Port Stephens Beds' by Baker (1924), who recognised it only in the southern part of West Falkland and adjacent islands. The type area is here regarded as the country between Cape Meredith, Port Stephens, South Harbour and East Bay. The Port Stephens Formation is divisible into at least five members (Plantation in West Falkland or Limpet Creek in East Falkland, Albemarle, Mount Alice, South Harbour, Fish Creek), the upper four being separated largely on photogeological criteria. No mappable boundary between the

Mount Alice Member and South Harbour Member was identified and it is possible that none exists (M Hunter oral communication, 1998). Indeed, on the basis of current information, it is not everywhere possible to separate the other members of the Port Stephens Formation, as indicated on the face of the map. Nevertheless, in due course it may be possible to re-define some of these members as formations, converting the Port Stephens Formation into a group.

In West Falkland, the Port Stephens Formation rests unconformably on an eroded surface of the Cape Meredith Complex. In East Falkland, the base is not seen. The Port Stephens Formation is everywhere overlain conformably by the Fox Bay Formation. There is no evidence that the Port Stephens Formation, as mapped, includes any unconformities. However, parts of the sequence are poorly exposed, even at the coast, and so such breaks could be overlooked.

The age of the Port Stephens Formation is very poorly constrained. As with the West Falkland Group, the base could be as old as Cambrian or as young as Devonian. It is unlikely to be older than about 520 Ma, the minimum age of lamprophyre dykes that cross-cut the Cape Meredith Complex (Section 2.2). It could be younger than about 422 Ma, the minimum age of the Group C dolerites at Cape Meredith (Section 2.2). Based on correlations with probable equivalents in South Africa, it is here considered most likely to be Silurian. The top of the formation is most probably Devonian, on the basis of its conformable upward passage to the Fox Bay Formation, and its likely regional correlation.

The Port Stephens Formation can be correlated with the Siluro-Devonian Nardouw Subgroup of the Table Mountain Group (Cape Fold Belt, South Africa) (Broquet, 1992). Both fluvial and marine depositional environments are represented in this thick arenaceous sequence, which overall reflects falling sea levels during the Silurian. The Limpet Creek Member (Section 2.3.1.2) might have been deposited during the latest Ordovician marine transgression which led to the deposition of the Cedarburg Formation (immediately pre-dating the Nardouw Subgroup) but otherwise this fossiliferous shale unit appears not to be represented in the Falklands.

2.3.1.1 Plantation Member

The Plantation Member comprises a thin sequence of reddish or grey sandstones with red mudstones and siltstones at the base of the Port Stephens Formation. It occurs only in the coastal section extending some 3.5 kilometres north-west of Cape Meredith.

Baker (1924) noted that the junction of the Port Stephens sandstones with the underlying granite at Cape Meredith is marked by 'a thin band of red micaceous shale'. He and subsequent authors who have offered brief descriptions of these basal red-coloured beds have given them no name. The name 'Plantation Member' is here proposed to describe them. The name is derived from 'Middle Plantation', which is a fenced area re-planted with tussac grass immediately adjacent to the thickest part of the unit.

The Plantation Member rests unconformably on an eroded and weathered surface of the Cape Meredith Complex. The granite underlying the Plantation Member is stained red, presumably due to penecontemporaneous weathering. The unconformity surface displays local irregularities of about one metre amplitude, but overall it appears essentially planar, dipping very gently east to north-east. The Plantation Member is conformable with the overlying Albemarle Member (Plate 3).

The unit typically comprises purplish-red to pale grey, medium to very coarse, moderately to poorly sorted, arkosic sandstones with thin beds, lenses and intraclasts of red or pale grey micaceous silty shales and fissile muddy silts. The sandstones are thinly to medium bedded, and locally cross-bedded. Oligomict conglomerates occur locally at and near the base, typically in lenses from 30 centimetres to one metre thick and a few metres wide, but up to 50 metres wide in one instance. Near the Lighthouse these conglomerates are represented by numerous very well-rounded cobbles and small boulders up to 40 centimetres long scattered over the ground near the base of the Port Stephens Formation. Most are composed of medium to coarse pink quartzite, some are vein quartz. Some are still partly coated by traces of their matrix sandstone and so are clearly not recent beach pebbles. Petrographic examination shows that the quartzite is composed of moderately sorted, very well-rounded quartz grains (mostly monocrystalline) with a ragged quartz overgrowth occluding the intergranular spaces. The original grains have a thin ferruginous coating (Lott, 1999). This is similar to lithologies found in the Albemarle Member, suggesting that the base of that unit is quite strongly diachronous.

The Plantation Member is interpreted as a transgressive fluvial deposit. Regionally, such a facies is likely to be strongly diachronous.

The Plantation Member is probably up to about 15 metres thick, diminishing towards the north and the south, and wedging out laterally near the Lighthouse. Thin lenses and pockets of conglomerate essentially identical to that in the Plantation Member occur at and near the base of the typical Albemarle Member sandstones exposed near the base of the cliff about 600 metres north-east of the Lighthouse.

The type area for the Plantation Member is the central section of the cliffs between Cape Meredith and Big Cape camp. The most complete section is assumed to be in the area west of Meredith Hill, between [TC 4950 0600] and [TC 5020 0500]. Both the base and the top of the Plantation Member are exposed at the base of cliff near the southern boundary of the Big Cape Formation outcrop [TC 4925 0630], where the unit is about five metres thick. This reference section is easily accessible from the rock platform immediately below the level of the basal unconformity, although descent to the platform itself requires care.

The age of the Plantation Member is not known but is assumed to be Silurian, as for the Albemarle Member (Section 2.3.1.3).

Baker (1924) noted similarities between the Plantation Member and a basal facies of the Table Mountain Group in the west of Cape Province, South Africa. However, such basal units are likely to be highly diachronous and there is no reason to suppose that these two occurrences are of similar age.

2.3.1.2 Limpet Creek Member

The Limpet Creek Member is composed of sandstones, mostly fine-grained, which form the lowest part of the Port Stephens Formation exposed in East Falkland. It crops out only in a small area of northern East Falkland around the head of the Limpet Creek tidal inlet.

The occurrence of a unit of fine-grained sandstones beneath the Albemarle Member was recognised for the first time during the present survey. The proposed name is taken from the large tidal inlet where the sandstones are best exposed. The base was not seen during the survey and is probably not exposed onshore. The Limpet Creek Member is apparently conformable with the overlying Albemarle Member, although the latter has not been mapped separately from the overlying parts of the Port Stephens Formation in East Falkland.

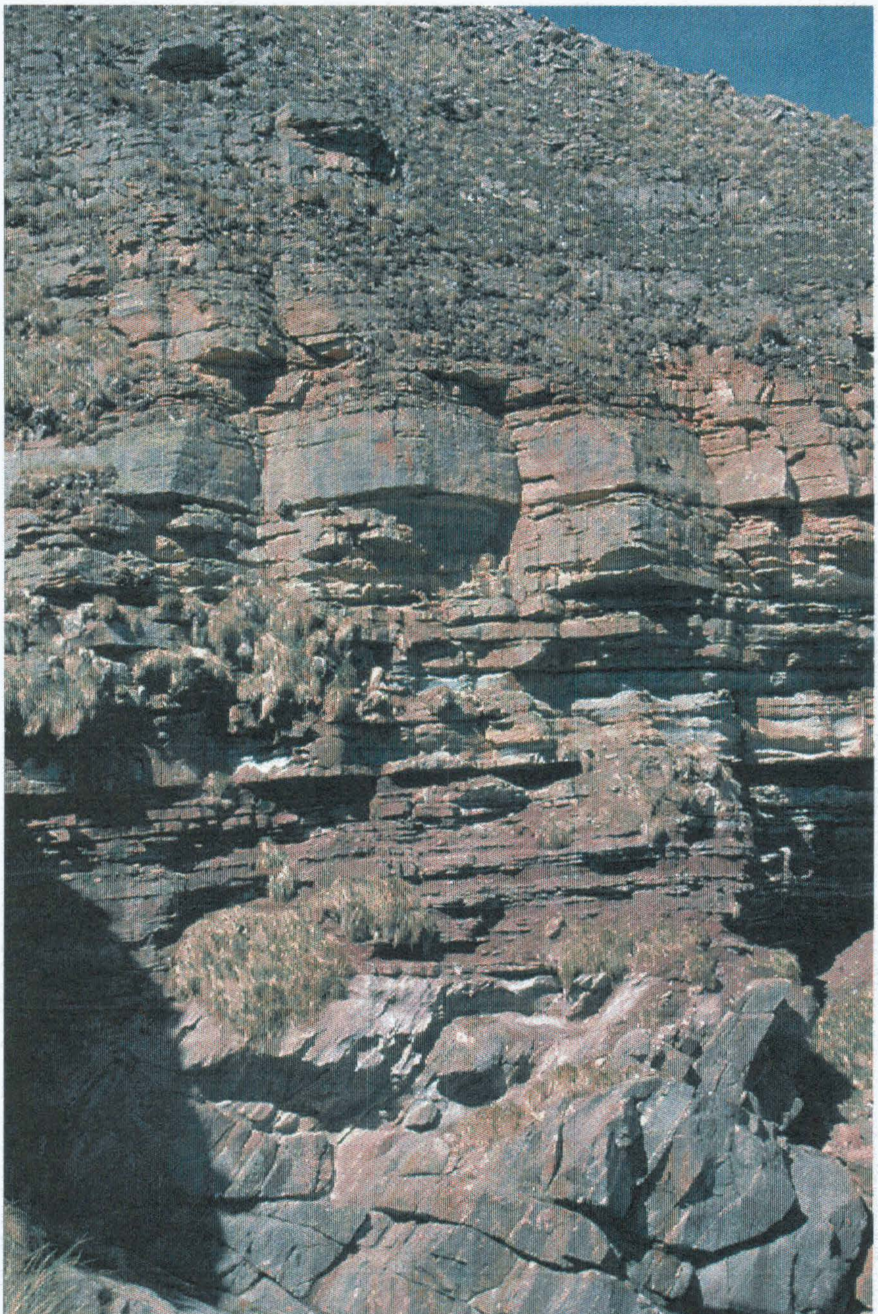
Exposures in the type area at the east end of Limpet Creek [UD 9065 0970] to [UD 9035 1175] are mainly composed of fine-grained sandstones, with some very fine or medium-grained sandstones. Traces of mica were observed. Plane lamination or thin plane bedding is usual but there is also a range of sedimentary structures indicating relatively rapid rates of deposition under a vigorous flow regime, including cross-lamination (some formed in climbing ripples) and de-watering structures. Both trough and tabular cross-beds are locally present. Bioturbation occurs in some beds. Palaeocurrent indicators point to a northerly direction. No fossils were found. These sandstones are thought probably to have been deposited in a near-shore marine environment, although other interpretations are possible.

The exposures lie within a low-lying area bounded to the east and north by a conspicuous break of slope visible on aerial photographs. Exposures of the Albemarle Member occur just above this break of slope. If the outcrop and structure are as mapped, the Limpet Creek Member could be up to 1000 metres in thickness.

There could be some doubt over the stratigraphic identity of the fine-grained sandstones at Limpet Creek. In the field they resemble some parts of the Fox Bay Formation, although no body fossils and no distinct trace fossils were observed. If they are part of the Fox Bay Formation, a considerable throw on faults bounding the outcrop would be indicated, possibly with over-thrusting by the Port Stephens Formation, as on the Sand Grass Thrust just to the east of Foul Bay (Section 4.2.4.2). However, the Albemarle Member sandstones overlying the Limpet Creek Member to the east of Limpet Creek are apparently conformable with the inferred contact and are not conspicuously recrystallised in the style seen above the Sand Grass Thrust. There is no evidence for high-angle faulting.

Also, exposures along the north coast of East Falkland near Cape Bougainville show very long east-dipping sections through the Port Stephens Formation. *Skolithos* sandstones typical of the Albemarle Member occur near Cape Bougainville and it is quite possible that the base of that member comes to the surface between Cape Bougainville and Limpet Creek. There is no particular reason why a fine-grained facies should not occur amongst the marine deposits in the lower part of the Port Stephens Formation. The photogeological and outcrop information is consistent with the fine sandstones at Limpet Creek lying conformably beneath the rest of the Port Stephens Formation. Their composition, sedimentary structures and lack of distinct trace fossils contrast strongly with those generally found in the Albemarle Member, indicating they should be considered as a separate unit.

Plate 3: Base of Port Stephens Formation near Cape Meredith



Photograph by Don Aldiss, BGS. MN28234

G2 granite gneisses of Cape Meredith Complex (with north-dipping foliation) are overlain by reddish-purple shales with thin sandstones of the Plantation Member, then by plane-bedded grey sandstones of the Albemarle Member (Port Stephens Formation).

The Limpet Creek Member is likely to be older than any of the Port Stephens Formation exposed at Cape Meredith, and so would be the oldest part of the sedimentary sequence exposed on the Islands. The Cedarburg Formation of the Table Mountain Group in South Africa comprises a relatively thin sequence of fossiliferous marine mudstones of latest Ordovician age. Together with the overlying, predominantly sandy succession of the Silurian it represents a coarsening-upwards sequence from offshore marine to fluvial deposits (Broquet, 1992). The Limpet Creek Member might be a sandy lateral equivalent to the Cedarburg Formation but the simplest interpretation is that it belongs in the overlying Silurian succession.

2.3.1.3 Albemarle Member

The Albemarle Member is typically composed of medium-grained quartz sandstones, often with the trace fossil *Skolithos*. As shown on the geological map, it crops out in the far south of West Falkland (the type area), in an inlier adjacent to the Hornby Mountains, and on Cape Dolphin. It also occurs as part of many of the undifferentiated outcrops of the Port Stephens Formation; in the Arch Islands, probably in the far south of Weddell Island, possibly at the east end of Byron Sound, on Sedge Island, Pebble Island and the adjacent north-eastern extremity of West Falkland, around Cape Bougainville, and from Macbride Head nearly as far south as Cow Bay and at least as far west as Shanty Ridge [VD 24 00].

The Albemarle Member is a newly recognised subdivision in the lower part of the Port Stephens Formation. The name is taken from 'Albemarle Station', the farm which very approximately coincides with the outcrop in the type area. There is some lithological variation within the member and it probably could be further subdivided. The Albemarle Member is conformable with the overlying South Harbour Member, and with the underlying Limpet Creek Member in East Falkland, or Plantation Member, where present in West Falkland. Where the Plantation Member is absent, the Albemarle Member rests directly on the CMC with strong unconformity. The base is likely to be strongly diachronous, at least locally near Cape Meredith.

The Albemarle Member is generally associated with a relatively low-lying topography of rolling hills with sporadic rocky exposures. One interval in the type area gives rise to numerous rocky outcrops. These include the 'Indian Village' (near Port Stephens), Knoll Island, Cauliflower Rocks and, probably, Grandfathers Rock, although the last coincides with a broad dolerite dyke. In the north of East Falkland, the base of the Albemarle Member is marked by a strong negative break of slope above an area of more subdued topography underlain by the Limpet Creek Member.

The most typical lithologies in this unit are moderately to very well-sorted, medium to coarse-grained quartz arenites with rounded to well-rounded grains. Medium to coarse-grained subarkosic sandstones with rounded to subangular grains also occur. Mica is generally absent. Near Cape Meredith, the basal 30 metres or so includes some very poorly sorted subarkosic sandstones, some of which contain granules or pebbles of vein quartz. Where the Plantation Member is absent, small lenses of conglomerate occur near the base and these include pebbles and cobbles of pink quartzite and vein quartz (Section 2.3.1.1). Pebbles occur only sporadically elsewhere in the unit. Boulders of matrix-supported oligomict conglomerate (with pebbles up to five centimetres across of grey and pink quartzite and micaceous granite) found in the Fitzroy Tillite at Hill Cove (Section 2.4.2) might also be a basal facies. In general, the Albemarle Member sandstones are all pale grey but red-coloured (weakly ferruginous) *Skolithos* sandstone occurs at Cape Dolphin (although that too has been bleached to pale grey along joints). M. Hunter (oral communication, 1997) reports that red-coloured sandstones, many with *Skolithos*, are usual on the coast between Cape Meredith and Albemarle Station. She also reports that several thin beds of blue-grey mudstone are exposed in the same section. The common presence of ponds at the base of sandstone knolls such as Cauliflower Rocks suggests that they are underlain by impermeable horizons, perhaps well-cemented sandstones, or mudstones.

The sandstones are typically thinly to thickly plane-bedded, especially in the lower parts of the sequence (Plate 3). Some thin tabular and trough cross-bedded sandstones do occur in places, usually in sets of 10 to 30 centimetres thickness, rarely up to one metre. Possible herring-bone cross-sets occur near Hoste Inlet House, near the top of the unit. The rare indications of palaeocurrent direction are mostly to the north-west and west-north-west but the axis of one trough cross-set was measured as N010°. Near Cape Meredith, Cingolani and Varela (1976) found a bimodal palaeocurrent distribution, with a main NNW direction and a secondary ENE direction.

The Albemarle Member is characterised by common and locally very abundant trace fossils in the sandstones. By far the most common and widespread is *Skolithos*. Rare *Diplocraterion* occurs in places, particularly at the top of the Member (M. Hunter, oral communication, 1997). The most spectacular type is here assigned to a new ichnospecies of *Heimdallia,* and this is locally abundant. In addition to these named forms, there are some bioturbated beds without distinct trace fossils.

Skolithos takes the form of cylinders lying perpendicular to bedding, typically five millimetres in diameter and about 10 to 30 centimetres long with, in some cases, a weathered-out central tube. They are composed of the same material as the matrix. Where present, they are usually very numerous and fairly regularly scattered through the rock typically 0.5 to 3 centimetres apart (Plate 4). They conform to the description of *Skolithos linearis* given by Alpert (1974). Particularly fine examples of this 'piperock' have been noted on the coast at the west end of Port Stephens [TC 367 194] (C. Aldiss, oral communication, 1997) and between Melvern Creek and White Rock Point [UD 459 030]. The second locality has extensive exposures of medium to thickly bedded *Skolithos* sandstone, including one interval of about 30 metres thickness without visible bedding planes.

Well-cemented sandstones near Meredith Hill include locally abundant examples of a conspicuous large trace fossil here assigned to a new ichnospecies of *Heimdallia.* They occur in an interval at least ten metres thick of pale grey to white, medium-grained (verging to coarse) quartz arenites with moderately to well-sorted, rounded to sub-angular grains. This interval is exposed on the slopes north and east of Meredith Hill [TC 4931 0729], [499 069], [5158 0450], [5175 0480].

This large trace fossil occurs as vertical or steeply dipping sheets assumed to represent feeding burrows. As seen on bedding surfaces, the sheets are each of uniform width, generally between 1.5 and four centimetres wide and 30 to 80 centimetres long, but examples up to 150 centimetres long were seen (Plate 5). They are usually straight or gently curved, or wavy in an irregular fashion, but one example describes a right-angled bend. They do not branch but do cross-cut successively and randomly. Most examples appear as ridges on weathered surfaces, while some lie in grooves, but both these appear to be manifestations of the same trace fossil. Good three-dimensional exposure shows that the traces observed on bedding planes extend from about 20 to 40 centimetres vertically, maintaining uniform width throughout. No distinct downward terminations were observed.

Many of the sheets are divided vertically into equal parts by a thin central septum, while some have no such division. Both forms can occur in the same exposure, but it is difficult to tell if the absence of a visible septum is due to variations in weathering, or if they represent distinct types. In the less deeply weathered examples, curved spreite spaced a few millimetres apart can be seen within the sheets. Where there is no central septum these extend the full width of the sheet. Otherwise, the spreite cross each half of the sheet and meet at a rather irregular, stepped surface which forms the septum. Seen from above, the spreite curve along the trace. Seen in side view it seems that the spreite are upwards-convex in the type with a septum and downwards-convex where there is none, but too few examples were seen to be confident that this is consistent. The sides of the sheets display indistinct downwards-convex subparallel curved striations. One example showed one set of striations cross-cutting another on the wall of the same sheet.

In some exposures these sheets occur together with *Skolithos,* which they usually post-date, but a few examples of younger *Skolithos* were seen in one place.

This trace fossil appears to have been mis-identified as *Arthrophycus* by Scasso and Mendia (1985). *Arthrophycus* is widespread in the Ordovician and Silurian, and is common in the Ordovician La Tinta Formation of Argentina (Acenolaza, 1978). This has been the basis of assigning the lowest part of the Port Stephens Formation to the Ordovician or Silurian (Scasso and Mendia, 1985). However, as illustrated by Acenolaza (1978, fig. 25), *Arthrophycus* has a tubular form with fine orthogonal transverse ornamentation, not curved spreite. The trace fossil described by Scasso and Mendia (1985) conforms to what was seen during the present survey, except that they interpreted the form as tubular [*tubulares*], not sheet-like. While some of the denser developments of this trace fossil seen in two-dimensional exposure do resemble *Arthrophycus,* that ichnospecies was not seen during the present survey.

On the other hand, the sheet-like trace fossil is close to the diagnosis of *Heimdallia* (Bradshaw, 1981), although it differs from the sole named ichnospecies, *H. chatwini,* in being generally broader and in usually having a central septum. Changes in direction shown by *H. chatwini* can be more abrupt than seen in the Albemarle Member trace fossils. The type examples were found in the Taylor Group of South Victoria Land, Antarctica. The only other known occurrence of *H. chatwini* is in the Tumblagooda Sandstone of Western Australia (Trewin and McNamara, 1995). In view of the similarities to the ichnogenus, and the differences to the existing ichnospecies, the sheet-like trace fossil with the central septum found near Cape Meredith is here proposed as a new ichnospecies, *Heimdallia meredithensis.* The similar trace

Plate 4: *Skolithos* sandstone, Albemarle Member

Photograph by Don Aldiss,BGS. MN28236
Near Melvern Creek, White Rock Point.

Plate 5: *Heimdallia 'meredithensis'*, Albemarle Member

Near Cape Meredith shanty. Photograph by Don Aldiss, BGS. MN28235 fossil without the central septum is provisionally regarded as a variation of the same ichnospecies. Further work is required on these traces.

Other examples of *Heimdallia* at or near the same horizon as those near Cape Meredith occur on Cross Island, Ten Shilling Bay Island (where they are particularly good), the west side of 'Ten Shilling Bay Rincon' (i.e. the peninsula south-east of Stephens Peak) and on Bird Island (R. W. Woods, written communication, 1999). The occurrence on Bird Island has been photographed by I. Strange (personal communication, 1996). According to Woods, *Heimdallia* also occurs on Tussac Island [TC 60 09]. This is probably at a higher level in the Albemarle Member. It also occurs at Cauliflower Rocks and in the lower part of the Mount Alice Member, near Dean's River.

The Albemarle Member could probably be subdivided by separating the beds that form Cauliflower Rocks and similar tor-like outcroppings. This interval is clearly visible on satellite images and aerial photographs. It comprises medium to coarse sandstone, some with sparse quartz granules. It is probably between 30 and 50 metres thick, and is mostly plane-bedded with sporadic *Skolithos* and *Heimdallia*, but some thin tabular and trough cross-beds also occur. It is generally more feldspathic than the average, but is moderately to well sorted, commonly with well-rounded spherical quartz grains.

The poorly sorted, immature sandstones near the base of the Albemarle Member are probably fluvial deposits. Elsewhere the predominance of plane-bedding and abundance of trace fossils suggests that most of the Albemarle Member was deposited in an inner shelf marine environment (cf. Bradshaw, 1981). The 'millet-seed' texture of the well-rounded quartz grains in many of the arenites, together with the ferruginous staining, strongly suggests aeolian deposition, but none of the sedimentary structures typical of that facies are seen. It is more probable that sand was reworked from contemporary aeolian deposits on the adjacent continent, being carried out to sea by the wind or perhaps by sporadic river flooding. However, Trewin and McNamara (1995) show that parts of the Tumblagooda Sandstone including *Heimdallia* were deposited on a large fluvial outwash plain. As with other parts of the Port Stephens Formation, further work is required to provide a detailed analysis of palaeoenvironment.

The thickness of the Albemarle Member is estimated as 800 metres in the type area. The possibility of fault-repetition suggests this could be an over-estimate. As much as 700 metres could be exposed in the Hornby Mountains inlier, and there is no evidence that the base of the Albemarle Member is exposed there. On East Falkland, near Cape Bougainville, the Albemarle Member is estimated at about 1000 metres thickness.

The type area for the Albemarle Member is the relatively low-lying land between Cape Meredith and the foot of the Mount Alice escarpment. Reference sections for the base of the member are as for the Plantation Member, together with the cliffs east of the small stream which lies about 500 metres north-east of Cape Meredith lighthouse [TC 5219 0424], where the Albemarle Member lies directly on the CMC. A composite section through this member could possibly be constructed from coastal exposures between Cape Meredith and the north side of Hoste Inlet [TC 47 18] to the west, and Albemarle Harbour [TC 56 17] to the east.

The Albemarle Member is tentatively correlated with the Goudini Formation of South Africa (Broquet, 1992), based on its position within the Nardouw Subgroup (Section 2.3.1). The abundance of *'Heimdallia meredithensis'* in the lower part of the Albemarle Member conceivably marks an event horizon which might be used for regional correlation, in the absence of body fossils. Unfortunately, there seems to be no record of this trace fossil in the Cape Fold Belt, possibly reflecting a facies change.

2.3.1.4 Mount Alice Member

The Mount Alice Member is mainly composed of medium to coarse sandstones, some with trace fossils of a variety of types. In its type area it forms the Mount Alice escarpment, extending from near Port Edgar in the east to Calm Head in the west. It also occurs in the Hornby Mountains inlier, in the higher ground on Weddell Island, on Beaver Island and the southern part of New Island. It probably also occurs in other areas of undifferentiated Port Stephens Formation, such as near Cape Bougainville and between Cow Bay and Volunteer Lagoon, in East Falkland.

The Mount Alice Member is a newly recognised subdivision of the Port Stephens Formation. The name is taken from Mount Alice, the highest point on a prominent escarpment in southern West Falkland. The Mount Alice Member is conformable with the Albemarle Member below and the South Harbour Member above. No mappable contact with the contrasting South Harbour Member has been identified.

In its type area, the Mount Alice Member underlies high ground with rather rounded, flat topped hills extending from the Mount Alice escarpment northwards. The escarpment separates the Mount Alice Member outcrop from the strongly contrasting terrain underlain by the Albemarle Member. The base of the Mount Alice Member is taken as the strongest and most laterally persistent negative feature near the base of the Mount Alice escarpment. This can be traced on aerial photographs reasonably consistently from Calm Head to Chaffers Gullet, with some uncertainty between Wood Cove and Port Stephens settlement, on the north-west coast of Albemarle Harbour, and near Port Edgar. No exposures on this feature were observed during the survey but it presumably marks a relatively weak bed, possibly a mudstone interval.

On the east side of the Hornby Mountains Inlier (in the Coast Ridge steep zone, Section 4.2.2.1), a prominent ridge with numerous exposures of medium-grained quartz sandstone with tabular cross-bedding occurs in the interval between typical Albemarle Member strata and coarse cross-bedded sandstones assigned to the South Harbour Member. This ridge is bounded on the west side by a faint negative break of slope separating it from much smoother, lower-lying topography with only sporadic large exposures of sandstone. The ridge is therefore taken to be the equivalent of the Mount Alice escarpment, and the negative break of slope as the base of the Mount Alice Member. The contrast in topography is accentuated by the drainage, which typically branches dendritically immediately upstream (west) of the boundary. This boundary can be traced south around the closure of the Hornby Mountains Anticline into a well-marked negative feature which extends north beneath the steep slopes and crags forming the east-facing edge of Mount Moody and the rest of the Hornby Mountains. Mount Moody is probably formed by the same interval of strata as Mount Alice.

In comparison to the Albemarle Member, the Mount Alice Member is generally coarser and more feldspathic. The trace fossil assemblage is different, and trace fossils tend to be confined to the tops of well-defined sedimentary cycles.

Most of the unit is composed of medium, coarse, and locally very coarse-grained or gravely quartz sandstones and subarkoses. Most are moderately to poorly sorted but some well-sorted quartz arenites, and also some fine-grained sandstones do occur. Petrographic examination of a specimen of medium to coarse sandstone from the top of the escarpment near Hoste Inlet found it is mostly composed of moderately sorted, rounded to well-rounded (millet-seed) monocrystalline quartz grains. Sparse polycrystalline quartz grains, zircon and the remains of deeply altered and leached feldspar are also present. The framework grains are tightly cemented together by quartz overgrowths with sporadic kaolinite filling some pores (Lott, 1999).

The Mount Alice Member commonly shows cyclic deposition. Each cycle is generally of the order of 10 to 30 metres in thickness, commencing with medium to coarse and pebbly sandstones resting on an erosive base. The lower part of a typical cycle has large-scale cross-beds (in tabular, trough or wedge-shaped forms) generally in cross-sets from 50 centimetres to two metres thick. Some graded cross-beds about 10 centimetres thick have been noted. In places, the large cross-sets in-fill channels some 20 to 30 metres wide, for example at the east end of Markhams Ridge [TC 450 226] (Plate 6). The cross-bedding becomes less distinct upwards in each cycle, passing into thinly to thickly plane-bedded, well-sorted sandstones with trace fossils. These can be of similar grainsize to those below but on the whole the cycles are typically upwards-fining. The bioturbated interval is up to about three metres, although it can be as little as 10 centimetres. In some places the upper part of the cycle is built of stacked thin tabular cross-sets in co-sets about one metre thick each extensively bioturbated in the top five to 15 centimetres, or of one to two metres of thin to medium plane bedded or laminated (rarely cross-laminated) sandstone with one or a few bioturbated beds up to 50 centimetres thick. In each case, the bioturbation is concentrated at the top of each bed in which it occurs.

As found elsewhere in the Port Stephens Formation, the bioturbated sandstones tend to be well-cemented and more resistant to weathering than the cross-bedded strata.

Palaeocurrent indicators seen in the area between Mount Alice and the Dean River area show northerly and north-easterly directions.

Several types of trace fossil are found in the Mount Alice Member. *Heimdallia* is locally present near the base of the member. *Skolithos* seems to occur throughout, but is sparser and much less common than in the Albemarle Member. Rare *Diplocraterion* is present in places. Narrow bi-lobed trails some 10 to 13 millimetres across and up to 20 centimetres long occur on the surfaces of some of the thicker foresets in tabular cross-bedded sandstone. The most common types of trace fossil are varieties of sub-horizontal vermiform burrows, typically between five and 18 millimetres in diameter and up to 70 centimetres long. They are arranged randomly. Some are straight, others gently curved. They cross-cut but

Plate 6: Channel-fill structure in sandstones of Mount Alice Member

 MN28239

Plate 7: Cross-bedded pebbly sandstone, South Harbour Member

 MN28238

do not branch. As seen in the type area by M. Hunter (oral communication, 1997) and on New Island, there are at least two forms of these burrows. The more common tend to be found in coarse sandstones, but were filled by fine sand and mud, presumably introduced from an overlying layer. The less common form occurs as raised ridges, less sinuous than the other, and occurring in finer-grained sands. Sand forms the infill to these burrows.

The Mount Alice Member is characterised by rather episodic deposition, with high energy influxes of sediment alternating with quieter periods when the substrate was colonised by animal life. This could be consistent with inner shelf to shoreface deposits influenced by tidal currents, or with low sinuosity fluvial deposits on a coastal plain. The Mount Alice Member is thus regarded as a transitional sequence between the marine deposits below and the fluvial sequences above.

The thickness of the Mount Alice Member is estimated as 700 metres in southern West Falkland.

The type area is the Mount Alice Escarpment, especially where it meets the mouth of One Pond Valley [TC 45 20], and the adjacent hills to the north as far as Watchful Valley and the south end of Carew Harbour. Reference sections can be found at the mouth of a small inlet on the north side of Hoste Inlet close to the mouth of One Pond Valley [460 205], and in a river bluff beside Stewart Brook, near the southern end of Carew Harbour [TC 481 267]. A complete composite section through the member could possibly be constructed from coastal exposures between Carancho Bluff [TC 28 22] and Calm Head [TC 31 18], although much of that coastline might be difficult or impossible of access. Otherwise most of the sequence is poorly exposed. Good exposures of the Mount Alice Member can be found in the southern part of New Island, for example on the coast north-east of South Hill [TC 04 58] and at the base of a gully at the north end of Grand Cliff [TC 0270 5894].

2.3.1.5 South Harbour Member

The South Harbour Member is mostly composed of medium and coarse-grained subarkosic sandstones. Together with the Fish Creek Member, it has commonly been described as typical of the whole Port Stephens Formation. The type outcrop, in the south of West Falkland, extends from Cape Orford in the west to Leicester Falls and Port Edgar in the east. As shown on the map, it also occurs on the northern part of New Island, in the Roy Cove inlier, and in the Hornby Mountains inlier. It has been found in areas of undifferentiated Port Stephens Formation on northern Weddell Island, near Main Point, on the eastern part of Pebble Island and the immediately adjacent part of West Falkland, north and west of Douglas, east of Cape Bougainville, between Foul Bay and Coutts Hill, and adjacent to other outcrops of the Fish Creek Member.

The South Harbour Member is a newly recognised subdivision of the Port Stephens Formation. The eponymous inlet is largely enclosed by the type outcrop. The South Harbour Member is conformable with the Mount Alice Member below and the Fish Creek Member above. No mappable boundary with the Mount Alice Member has been identified. According to M. Hunter (oral communication, 1998) there is a complete gradation between these two members on West Falkland. Nevertheless, the absence of a mappable boundary does not of itself invalidate these two units: a lithostratigraphic member is not necessarily a mappable division. Typical developments of the South Harbour Member and of the Mount Alice Member represent contrasting facies. In contrast to the Mount Alice Member, the South Harbour Member is coarser and more feldspathic, lacks trace fossils and is dominated by large-scale trough cross-bedding. Furthermore, analysis shows that sandstones from the two units contain heavy mineral assemblages of different provenance: that of the Mount Alice Member is similar to the Albemarle Member, while the heavy mineral signature of the South Harbour Member resembles that of the succeeding formations (Knox and Aldiss, 1999). If the two units were combined as, say, the South Harbour Member, at times there would be the need to use cumbersome expressions such as 'the Mount Alice facies of the South Harbour Member'. It is here considered better to name the units separately.

There are also differences in the texture of the terrain underlain by typical developments of the two units. The Mount Alice Member has steeper slopes and rather rounded hills with very broad summit areas. Long dip slopes are not apparent. The landscape of the South Harbour Member outcrop is generally lower-lying and more undulating with gentler slopes, including some long dip slopes. These differences in topographic texture imply that there are differences in the underlying lithostratigraphy.

The South Harbour Member is typically composed of medium to very coarse subarkosic sandstones, although medium to coarse quartz arenites do occur. Some of the sandstones are slightly micaceous. They range from poorly to well sorted with subangular to rounded grains, and some include pebble lags or isolated quartz pebbles up to 15 millimetres diameter, or sporadic mudflake conglomerates and rare sandstone intraclasts. Individual mudstone intraclasts up to 13 centimetres long occur in a very coarse pebbly sandstone near South Harbour. M. Hunter (oral communication, 1998) reports a thick bed of gritty gravel on the north-east side of the entrance to Carew Harbour. No fossils have been found.

Petrographic examination of a coarse-grained sandstone from near the top of the South Harbour Member near Port Louis found ragged, poorly sorted monomineralic quartz grains welded together by quartz cement. The original grain outlines were rarely seen and some grain boundaries were sutured. Sparse strained rock fragments were present, but no feldspar. The few open pores were partially occluded by clays (Lott, 1997). A medium to very coarse subarkosic sandstone from near South Harbour comprised poorly sorted, sub-rounded K-feldspar and ragged mono- and polycrystalline quartz grains, together with altered ?biotite, muscovite, pyrite and patches of clay-rich matrix/cement. Most of the cement comprises ragged quartz overgrowths. A fine-grained sandstone from near Cape Bougainville was made up of well-sorted, sub-angular to sub-rounded monocrystalline quartz grains with subordinate K-feldspar and muscovite mica. The larger mica grains are partly altered. Sparse polycrystalline quartz grains are also present, as is zircon. Pyrite aggregates are concentrated along laminae. There is pervasive illite cementation (Lott, 1999).

Diamictite (then assumed to be tillite) with minor shale has been reported to occur in the Port Stephens Formation near South Harbour (Baker, 1924; Marshall, 1994b), but Hyam et al. (1997) found that these rock-types actually occur in sedimentary dykes (Section 4.4).

In West Falkland, the South Harbour sandstones were typically deposited in large-scale, low-angle trough cross-sets between 10 centimetres and two metres thick, and between two and 20 metres across. The individual cross beds are generally thin and normally have tangential bases and cut-off tops (Plate 7). Some cross-sets are overturned or slumped. There are also some wedge-shaped or tabular cross-sets, which are generally thinner. All types tend to be grouped in planar co-sets between 50 centimetres and three metres thick and so when viewed at a distance can be taken for plane-bedded sandstone. There are also rare intervals of between 30 centimetres and one metre of laminated, cross-laminated or thinly plane-bedded, medium to fine-grained subarkosic sandstone, in at least one case forming the base of a coarsening-up cycle. In contrast, parts of a sequence exposed near Cape Bougainville [UD 9920 1402] consist largely of tabular cross sets between 10 centimetres and one metre thick in well-sorted medium to fine-grained sandstone, with subordinate trough cross-bedding (Plate 8). Intervals of up to two metres of laminated or fine or very fine-grained sandstone also occur there, some with sparse mudstone intraclasts.

Primary current lineation in thinly plane-bedded sandstones near South Harbour is oriented NE-SW, but in adjacent parts of the coast trough cross-bedding generally indicates deposition by more northerly currents. On New Island, however, currents flowed towards the east-north-east or east. Scasso and Mendia (1985) found NNE palaeocurrent directions near Shag Cove.

The immaturity of the sediments, the preponderance of large-scale trough cross-beds, the presence of mudstone intraclasts and the absence of trace fossils together suggest that the South Harbour Member was deposited in a fluvial environment. In the area between Cape Orford and South Harbour, palaeocurrents flowed mainly towards the north or north-east, although some variation from north-west to east-south-east occurs. In spite of this variation, it seems likely that deposition was in a low sinuosity fluvial environment, as on a coastal plain crossed by braided rivers.

The thickness of the South Harbour Member is estimated as 600 metres in southern West Falkland.

The type area lies between South Harbour and Carew Harbour, including South Harbour Rincon [TC 40 30]. Reference sections can be found at Tern Point [TC 36 36] and the adjacent coastline, and on the shores of Anthony Creek and Carew Harbour. A complete section through this member could possibly be constructed as a composite from observation of coastal exposures between the north side of South Harbour Rincon and Cape Orford [TC 20 29], together with the coastline between Cape Orford and Carancho Bluff [TC 28 22]. Note that the southern shore of Port Richards approximates to a strike section. Thus strata exposed on the northern side of South Harbour Rincon, for example at the mouth of Carew Harbour, lie near the top of the South Harbour Member.

2.3.1.6 Fish Creek Member

The Fish Creek Member is mostly composed of medium to coarse-grained subarkosic sandstones. Together with the South Harbour Member, it has commonly been described as typical of the whole Port Stephens Formation. As shown on the maps, it occurs mainly in central West Falkland, around Main Point and on the adjacent islands. It probably also occurs on the fringes of all of the outcrops of Port Stephens Formation in East Falkland, although it cannot be differentiated everywhere. Note in particular that the Fish Creek Member probably underlies the low ground around

Plate 8: Tabular cross-bedded sandstones, Port Stephens Formation

East of Cape Bougainville. Photograph by Don Aldiss, BGS. MN28237

Plate 9: Base of Fox Bay Formation, Rapid Point

Photograph by Don Aldiss, BGS. MN28240

Moss Side, north of Port San Carlos [UC 67 95], as indicated on Section H-J in the margin of the East Sheet of the geological map (Aldiss and Edwards, 1998).

The Fish Creek Member is a newly recognised topmost subdivision of the Port Stephens Formation. The name is derived from Fish Creek camp (west of Fox Bay), the type area. The Fish Creek Member is conformable with the South Harbour Member below and the Fox Bay Formation above. It is composed of very similar sandstones to the South Harbour Member but overall they are somewhat less coarse, and the Fish Creek Member underlies distinctly less rugged topography. The marked contrast in topography provides a means of separating the two members in most areas, and suggests that there is some consistent difference between the lithological composition of the two members. This possibly lies in the style of cementation in the sandstones, but the actual nature of the difference and its cause are not known. The base of the unit coincides with a strong negative break of slope. The same break of slope was taken by Greenway (1972) as the base of the Fox Bay Formation, but this seems to have arisen through a mistaken photogeological interpretation rather than a difference in stratigraphic concept.

The Fish Creek Member is composed of fine to medium-grained sandstones with some coarse subarkosic sandstones. Sorting is mostly moderate to good, but in places it is poor. The larger sand grains are typically moderately to well-rounded. Small mudstone intraclasts occur in sandstones near the top of the unit in Two Pass Stream, near Teal River [TC 8511 7619], in one instance in association with plant debris and quartz pebbles up to three centimetres diameter in a thin lens of matrix-supported lag conglomerate. Laminated mudstones or siltstones, some with detrital mica or plant debris, occur locally, especially near the top.

Most of the sandstones were deposited in trough cross-beds, although intervals of tabular cross-bedding are also present, while some beds are massive, thinly bedded or laminated. Trough cross sets are typically between 10 centimetres and three metres thick, with thin to laminated low angle cross-beds. Palaeocurrent indicators generally point towards the north or north-east.

Laminated or thinly bedded fine micaceous sandstone, some with current ripple cross-lamination, flaser bedding and small-scale cross bedding occurs near the top of the member. These bedforms tend to occur at the top of cyclic units mainly composed of the more usual medium-grained cross-bedded sandstones. In places, the topmost 20 to 30 metres of strata could be regarded as 'passage beds' with the overlying Fox Bay Formation. This is indicated by the appearance both of mudstones in the laminated intervals and of bioturbated sandstone beds up to one metre thick. In Billy Creek, near Main Point [TD 98 00], these beds of bioturbated sandstone occur between the cross-bedded sandstones and the laminated interval in up to five successive cycles of deposition. However, as elsewhere the base of the Fox Bay Formation is marked by an abrupt change.

Poorly preserved fossilised plant debris occurs in laminated micaceous silty fine sandstone near the top of the unit at Doctors Creek, Fox Bay, [TC 9051 4091] and in Two Pass Stream [TC 8511 7619], where it occurs within two metres of the base of the Fox Bay Formation. This debris appears as black streaks up to about 25 millimetres long and three millimetres wide. Some are gently curved and some of the larger fragments show very fine linear ornament.

As with the South Harbour Member, the Fish Creek Member was deposited in a fluvial environment. The thickness of the Fish Creek Member varies from about 300 to 450 metres.

The type area lies between Port Richards and Symonds Harbour, with representative sections near the road crossing near Fish Creek House [TC 6820 4511]. Reference sections for the base occur at the head of Port Richards near Watering Cove [TC 6555 4165], and in House Stream, immediately west of Port Howard settlement [UC 2536 7918]. Other reference sections should be found along the coastline of Queen Charlotte Bay from [TC 6219 4594] southwards; on the south-east side of Edye Creek [TC 6485 4790] to [TC 675 497]; and adjacent to the stream draining Top Hog Ground Pond [TC 954 707]. The top of the unit is well exposed beside Two Pass Stream [TC 8511 7619] and on the south side of Billy Creek (Main Point) [TD 986 007].

2.3.2 Fox Bay Formation

The Fox Bay Formation is mostly composed of fine to medium-grained sandstones, siltstones and mudstones. It is widely distributed in West Falkland and the smaller islands to the north, and in the northern part of East Falkland.

Baker (1924) used the name 'Fox Bay Beds' for the 'fossiliferous series' which underlies the low ground between the major ranges of quartzite hills in both East and West Falkland, regarding Fox Bay as the type area. He was unable to describe the contacts with the adjacent formations. The present survey has shown that the Fox Bay Formation conformably overlies the coarse sandstones of the Port Stephens Formation, and that it is overlain conformably by the Port Philomel Formation. One subdivision, the East Bay Member, is newly recognised at the top of the formation in parts of West Falkland (Section 2.3.2.1). Future work may demonstrate that the Fox Bay Formation can be further subdivided, for example by reference to marker horizons.

The Fox Bay Formation is distinguished from the Fish Creek Member by its general lack of medium and coarse-grained sandstones, the common presence of argillaceous lithologies including black mudstones and shales, the dominance of plane bedding with small-scale cross-bedding instead of large-scale cross-bedding, and its association with subdued topographic features. Bioturbation is common and marine invertebrate fossils are diagnostic. Plant fossils occur in the uppermost part of the formation, as mapped (Section 2.3.3), and very locally elsewhere in the formation.

The Fox Bay Formation everywhere forms low-lying ground and as seen on aerial photographs tends to show more delicate bedding traces than the Fish Creek Member. Some of the sandstones can give rise to persistent topographic features, and where the beds are gently dipping, they produce a characteristic pattern of valleys and ridges lying parallel to strike. Foreshore exposures of the Fox Bay Formation have a finely laminated appearance on aerial photographs, consistent with the relatively thin bedding seen in the sedimentary sequence. By contrast foreshore exposures of the Fish Creek Member generally lack lineations and where present these are much coarser and more discontinuous that in the Fox Bay Formation. Furthermore, many foreshore exposures of the Fish Creek Member display a texture characteristic of very gently dipping cross-bedded sandstones. The Fox Bay Formation is more likely to have bare 'clay patches' on the flat ground of interfluves, but less likely to have peat. River bluffs on the Fish Creek Member are likely to be craggy or 'rough' textured, whereas those on the Fox Bay Formation usually appear smooth.

The base of the Fox Bay Formation is marked by an abrupt change from a sequence of mainly fine to medium-grained cross-bedded sandstone to one which is composed mainly of bioturbated fine sandstone. Although the topmost 20 metres or so of the Fish Creek Member can show a mixture of characters of the two units, the base of the Fox Bay Formation is generally obvious where seen in exposed sections (Plate 9). It can be defined as the base of the first bed of bioturbated sandstone more than one metre thick above the cross-bedded sandstones of the Fish Creek Member. For example, at the south-eastern end of the Mile Ridge beside Two Pass Stream [TC 8511 7619], the Fox Bay Formation commences with more than six metres of medium to thinly bedded bioturbated fine feldspathic sandstone. This is reddish or yellowish brown in colour with irregular, short dark brown to black subvertical burrows and other rather non-descript bioturbation. This rock-type is typical of the base of the Fox Bay Formation in most areas. The basal sandstone of the Fox Bay Formation is often marked by a low ridge, and in unexposed ground the basal contact is placed at the slight negative topographic feature below it. However, the topographic expression of the contact is variable, depending on the degree of cementation in the local sequences and on their attitude.

The Fox Bay Formation typically comprises brownish-yellow and reddish-brown to grey or black fine-grained and fine to medium-grained sandstones. These are interbedded with siltstones and mudstones in varying proportions. Coarse sandstones are a minor component. The sandstones are typically subarkosic arenites, commonly with fine-grained white mica. Some of the sandstones are well sorted, but argillaceous types also occur. Grey bioturbated muddy sandy micaceous siltstone is a common component in the lower part of the formation. Some beds are weakly calcareous and carbonate or pyritic concretions are present in places, especially in the area north of Hill Cove. The localised occurrence of large oblate diagenetic spheroids up to two metres in diameter, as at Kelp Point near Fox Bay [TC 90 37], may also indicate weak calcareous cementation. Limestone is virtually absent but limestone nodules occur locally in mudstones on the south coast of Pebble Island, south of First Mountain. A single piece of crinoidal limestone was collected on a track near Shallow Harbour House, West Falkland. Although this was found to include marine algae of mid-Namurian to mid-Devonian age (N Riley, oral communication, 1998) and therefore could have been derived from the local Fox Bay Formation outcrops, the possibility remains that this fragment had been carried into the area, perhaps from outside the Islands.

Petrographic examination of five specimens of fine, medium or coarse-grained sandstone from the Fox Bay Formation found little variation in composition. The sandstones were moderately well sorted, and composed mainly of monocrystalline quartz, with abundant feldspar, and sparse rock fragments including cherts. Most contain some muscovite mica although the coarser sandstones lack mica. The framework grains tend to have a ferruginous coating, and in some pyrite grains are preserved. The cement is partly quartz and partly ferruginous clay (Lott, 1997).

Table 2: Devonian invertebrate fossils of the Falkland Islands

This table lists all named fossil invertebrates recorded from the Fox Bay Formation to date. Names marked * were assigned by Cocks (1996) and Cocks et al. (1998) from specimens collected during the present survey or held by the Natural History Museum, London. The trilobites marked + are listed by Edgecombe (1994) and # by Cooper (1982). Names given for the brachiopods and trilobites supersede those in the similar lists of Clarke (1913a), Baker (1924) and Greenway (1972), which include some names which are now obsolete. Some names given here for other taxa may also be obsolete. Those marked § are taken from (Clarke, 1913a) and those β from Baker (1924).

Sponges

§ *Clionolithus priscus* (McCoy)

Conulariids

§ *Conularia africana* Sharpe

Cricoconarids

§ *Tentaculites* sp.

Brachiopods

**Australocoelia palmata* (Morris and Sharpe)
**Australospirifer hawkinsii* (Morris and Sharpe)
**Australostrophia mesembria* (Clarke)
**Craniops* sp.
**Meristelloides?* sp.
**Notiochonetes skottsbergi* (Clarke)
**Orbiculoidea falklandensis* Rowell
**Pleurochonetes falklandicus* (Morris and Sharpe)
**Pleurothyrella falklandica* (Clarke)
**Protoleptostrophia concinna* (Morris and Sharpe)
**Schellwienella sulivani* (Morris and Sharpe)
**Tanerhychia* sp.

Gastropods

§ *Bellerophon (Plectonotus) quadrilobata* (Salter)
§ *Diaphorostoma allardycei* Clarke
§ *Loxonema?* sp.
β *Ptomatis moreirai* Clarke
§ *Tropidocylcus antarcticus* (Clarke)

Bivalves

§ *?Cardiomorpha colossea* Clarke
* *Cuneamya?* sp.
β *Janeia* sp.
β *Leptodomus* cf. *ulrichi* Clarke
* *Modiomorpha?* sp.
* *Nuculites* sp.
β *Nuculites* cf. *branneri* Clarke
§ *Nuculites reedi* Clarke
§ *Nuculites sharpei* Reed
§ *Palaeoneilo* sp.
§ *Toechomya ?* sp.

Cephalopods

§ *Orthoceras* cf. *gamkaensis* Reed

Trilobites

+*Bainella falklandicus* (Clarke)
+*Bainella nilesi* Edgecombe
*#*Burmeisteria herscheli* (Murchison)
+*Metacryphaeus allardyceae* (Clarke)
+*Oosthuizenella ocellus* (Lake)
§ *Proetus* sp.

Crinoids

§ *Botryocrinus doubleti* Clarke
Also indeterminate crinoid ossicles, often together with lengths of stem up to 10 cm long.

Much of the sandstone is medium to very thinly plane-bedded or laminated, but hummocky or swaley cross-stratification is a conspicuous component in some beds. Individual beds with hummocky cross-stratification can be expected to be widely developed and some may act as marker horizons. Some of the plane-bedded micaceous sandstones are conspicuously flaggy, with primary current lineation. Some intervals with trough or tabular cross-bedding also occur, with ripple-marks. Examples of low-angle trough cross-bedded sandstone with sparse mudstone intraclasts can be seen in places. Soft sediment structures attributable to slumping, or loading and de-watering, including ball-and-pillow structure, occur locally (Marshall, 1994b). Siltstones and mudstones are thinly bedded, laminated or bioturbated, but low angle cross-lamination occurs in places.

In the islands north of West Falkland an interval of fine, medium and coarse trough cross-bedded sandstones very similar to those of the Fish Creek Member occurs within the lower part of the Fox Bay Formation. This interval forms the shoals of Bold Rocks, east of Keppel Island [TD 98 08], but is better exposed on the southern coast of Pebble Island, due south of Middle Peak [UD 080 140], where some of the cross-sets are capped by plane-bedded sandstone with primary current lineation. The same interval of sandstones occurs on Saunders Island, with at least one horizon of desiccation cracks. The fine-grained, laminated flagstones seen at Fox Bay (Section 7.6.4) and at intervals north to Chartres, and at San Carlos, could represent a similar high-energy episode within the Fox Bay Formation.

Some beds in the lowest third of the formation contain locally very abundant marine invertebrate fossils. Brachiopods and fragments of crinoid stem are the most common in most parts of the Fox Bay Formation, but trilobites, gastropods, bivalves, rare cephalopods, bryozoans and cricoconarids (e.g. *Tentaculites*) also occur. This relatively low diversity fauna lived in shallow seas at high latitudes, in waters too cold to promote the deposition of limestones or the growth of corals (Clarke, 1913a; Scotese and Barrett, 1990; Cooper, 1982; Cocks and Fortey, 1988). Fossils known to occur in the Fox Bay Formation are listed in Table 2. The most common brachiopods and trilobites found in the Falkland Islands are shown in Plates 10 and 11, and many of the papers quoted in the following include illustrations of types found in the Falklands, amongst others.

These fossils occur most commonly as shell drifts (coquinites) up to one metre wide and five centimetres thick, most of which are associated with hummocky cross-stratification in sandstone. The carbonate has been leached from the shells in these storm-generated death assemblages, and the fossils all occur as moulds. Ferruginous cement is usual. Fossils also occur in laminated siltstones and mudstones, and these are more likely to be isolated individuals and to have at least part of the shell preserved. Well-preserved fossils can be found in calcareous or pyritic nodules in mudstones on Pebble Island (on the coast south of First Mountain) [UD 15 13], at Rapid Point [TD 90 01], at Port North [TC 65 90], and Caneja Creek, near Horseshoe Bay [VC 12 91]. Trilobites are better preserved and more abundant at these localities than elsewhere in the Fox Bay Formation. This style of preservation is associated with mudstones and with the thicker sequences of East Falkland, suggesting deposition in relatively deep water.

Descriptions of small numbers of brachiopods from the Fox Bay Formation were made by Morris and Sharpe (1846) and by Newton (1906). Clarke (1913a) reviewed the Early Devonian marine invertebrate fossils of the Paraná Province of Brazil, comparing them with collections from the Falkland Islands, Bolivia, Argentina and South Africa, which together comprise the 'Malvinokaffric Province'. He concluded that the faunas from these diverse areas showed a remarkable homogeneity, distinct from those of the northern hemisphere. Nevertheless, he also found (to his evident surprise) that the Falkland Islands Devonian fauna resembles that of South Africa more closely than that found in South America. His report (Clarke, 1913a) is illustrated with very fine engravings of the fossils. These were also published by Clarke (1913b) but without the report.

The classification of certain brachiopods found in the Falkland Islands is discussed by Boucot and Gill (1956) and Hiller (1987). Modern reviews of the taxonomy of the trilobites of the South African Bokkeveld Group (many of which also occur in the Falklands) has been compiled by Cooper (1982), and of the calmoniid trilobites of the Falkland Islands by Edgecombe (1994). No such review of the other fossil groups has been made. Given that the taxonomic nomenclature of fossils commonly undergoes revision in the light of new knowledge, fossil identifications quoted in older works should be treated with circumspection. Fossil determinations for the present survey were carried out by Cocks (1996) and by Cocks et al. (1998) (Table 2).

Marine invertebrate fossils of these types occur only in the Fox Bay Formation, as far as is known, but there are two records of these fossils having been collected in locations where according to current evidence the Fox Bay Formation is not present.

Baker (1924, p. 11) mentions that the Challenger expedition "obtained fossils at Macbride's Head". HMS Challenger visited the Falkland Islands in 1876. Etheridge (1885, p. 893, footnote) states "Large numbers of the *Atrypa palmata,* M. and S. [*Australocoelia palmata* (Morris and Sharpe)], are scattered about the fossiliferous layers of the blocks from Port Louis, and the mass from Macbride's Head, East Falklands". No further relevant details are given. Following the present survey, however, no occurrence of the Fox Bay Formation is known to exist anywhere close to Macbride Head, which instead exposes *Skolithos* sandstones typical of the Albemarle Member. It is conceivable that the Fox Bay Formation is present in a small block bounded by faults of considerable throw, or that pieces of fossiliferous rock have been washed in by the sea, but it seems more probable that an error of location or labelling was made by the Challenger observers.

Newton (1906) reports that Mr W Felton of West Point Island presented a slab of 'hard sandy rock' with crinoid debris to the 'Scottish National Antarctic Expedition' of 1904. This slab had been collected from Hope Point [TD 45 05]. The description conforms to the crinoid debris common in the Fox Bay Formation, but photogeological interpretation during the present survey indicates that Hope Point is actually composed of the Port Stanley Formation. It is most likely that the specimen had been washed up by the sea.

The occurrence of sparse plant debris in the Fox Bay Formation has been suspected since Anderson (1907) reported a 'minute *Calamites*-like fragment' from Port Louis. However, parts of the laterally elongate brachiopod *Australospirifer hawkinsii* have a plain linear ornament and it seems possible that a small piece could have been mistaken for a plant fragment by Anderson. Nevertheless, Baker (1921; General report 1, p.10, FIG archive) reports that good specimens of a fossil plant 'resembling *Sphenopteris*' have been obtained from shales exposed in the paving stone quarry at Fox Bay. Lycophyte plant debris like that found in the Port Philomel Formation occurs in the topmost part of the Fox Bay Formation, as mapped (Section 2.3.3).

Although bioturbation is common in most parts of the Fox Bay Formation, distinct trace fossils are less usual. *Planolites, Zoophycos,* and *Diplocraterion* are widespread in the siltstones and mudstones. A bi-lobed convolute trail resembling *Cruziana* occurs at Kelp Point, Fox Bay. Marshall (1994b) records an assemblage of vertical trace fossils including *Arenicolites* at Port North.

The thickness of the Fox Bay Formation is estimated as 650 metres (including the East Bay Member) at Fox Bay, 800-900 metres at Port North and Saunders Island and 960 metres at Port Howard. There seems to be an overall increase towards the north and the east in West Falkland. The complexities of the structure in East Falkland, in particular the probable presence of strike-parallel faulting, together with the generally poor exposure and lack of structural information, mean that previous estimates of the thickness of the Fox Bay Formation should be treated as broad indications. The present estimate is based on the length of the Malo Hills anticlinal axis near the south end of Port Salvador, which can be traced almost continuously from the base to the top of the Fox Bay Formation, and which Curtis and Hyam (1998) report as plunging east at 10°. A thickness of some 1600 metres is indicated, by far the greatest in the Islands. The Fox Bay Formation sequence in eastern East Falkland is therefore likely to be the most useful for intercontinental stratigraphic correlation.

Greenway's (1972) geological map purports to show the Fox Bay Formation combined with the Port Philomel Formation, but the revised geological map (Aldiss and Edwards, 1998) shows that in fact the Port Philomel Formation was mostly included with the Port Stanley Formation on the earlier map. Similarly, ground underlain by the Fish Creek Member was included with the Fox Bay Formation by Greenway (1972). These differences should be borne in mind when comparing estimates of formation thickness based on Greenway's map. Furthermore, as described by Scasso and Mendia (1985), the Fox Bay Formation encompasses both the Fish Creek Member and the Port Philomel Formation.

The sediments of the Fox Bay Formation are thought to have been deposited in a range of marine environments, including inner shelf (within storm wavebase) and shoreface. The intervals of plane-bedded sandstone and large-scale cross-bedding demonstrate periods of high-energy transport and deposition, but these could be a consequence of tidal currents in the shoreface environment and do not necessarily indicate a return to fluvial deposition. At least four shallowing-upwards cycles of deposition are represented in the Fox Bay Formation and probably more in some areas, such as Port North and eastern East Falkland.

There is an overall trend to deeper water facies in the north and east of the Islands. Sequences in the Fox Bay area are dominated by sandstones and siltstones, with only thin mudstone beds up to about 30 centimetres thick. The sandy East

Plate 10a: Common brachiopods, bivalves and criconariids of the Falkland Islands

1 *Orbiculoidea falklandensis* Rowell, 1965. x 1.5. BB17513, Pebble Island, West Falkland. Collection A G Bennett, British Graham Land expedition, 1934.

2 *Protoleptostrophia concinna* (Morris and Sharpe, 1846). Internal mould of ventral valve. x 1.5. B15639. Collection C Darwin, HMS Beagle.

3 *Pleurochonetes falklandicus* (Morris and Sharpe, 1846). Internal mould of dorsal valve. x 2. B17792. Collection C Darwin, HMS Beagle.

4 *Pleurochonetes falklandicus* (Morris and Sharpe, 1846). Internal mould of ventral valve. x 2. B17793. Collection C Darwin, HMS Beagle.

5 *Pleurothyrella falklandica* (Clarke, 1913). Internal mould of dorsal valve. x 1. B60290. Presented by R. Baker, 1931.

6 *Schellwienella sulivani* (Morris and Sharpe, 1846). Internal mould of ventral valve. x 1.5. BC 5662. Collection C Darwin, HMS Beagle.

7 *Schellwienella sulivani* (Morris and Sharpe, 1846). Internal mould of dorsal valve. x 1.5. B60286. Presented by R. Baker, 1931.

8 *Australospirifer hawkinsii* (Morris and Sharpe, 1846). Internal mould of ventral valve. x 1. B60275. Presented by R. Baker, 1931.

9 *Australospirifer hawkinsii* (Morris and Sharpe, 1846). Internal mould of dorsal valve. x 1. B60272. Presented by R. Baker, 1931.

10 Criconariids (cf. *Tentaculites*) and bivalve moulds. x 1. FAL60. Falkland Islands Geological Mapping Project.

All specimens except #10 are in the collections of the Natural History Museum, London.

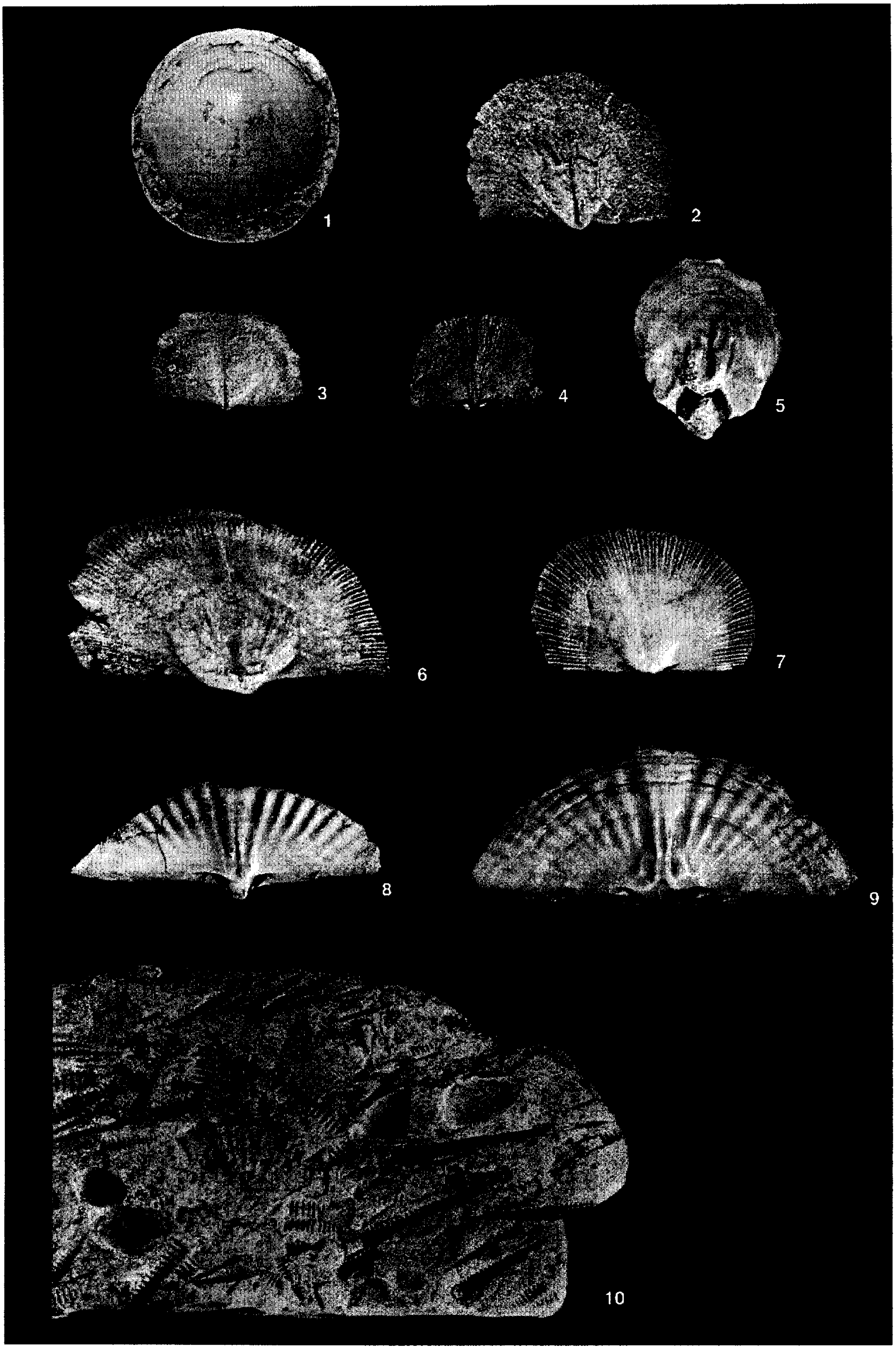

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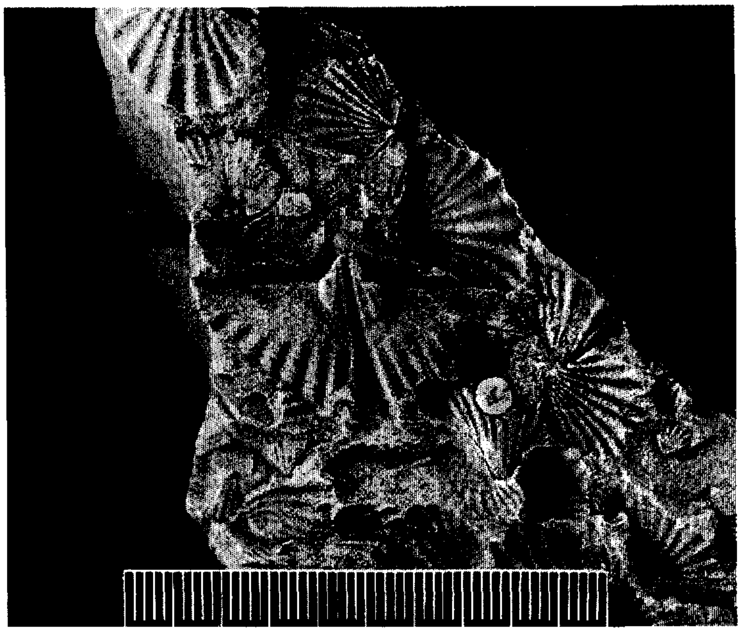
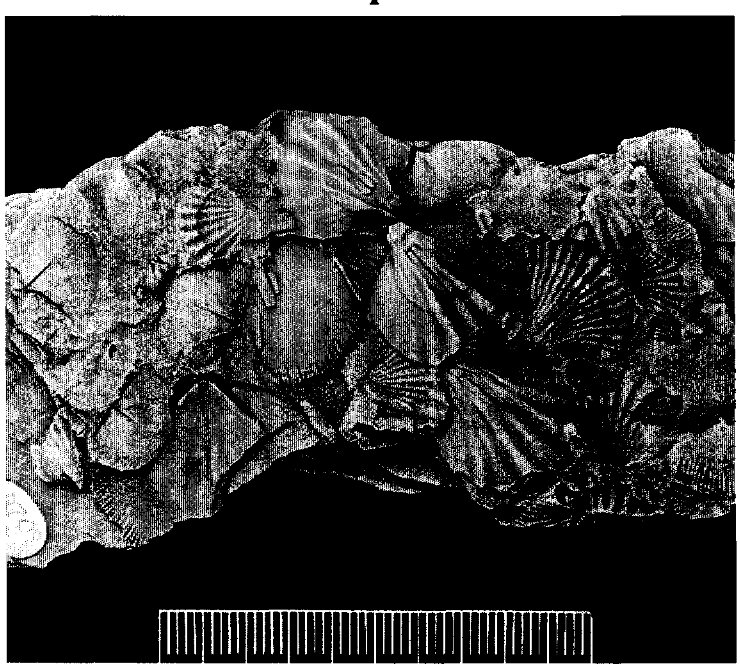
Plate 10b: Common brachiopods of the Falkland Islands



Brachiopods in a shell-drift (coquinite), in this case deformed during D1 folding, as shown by the shape of the internal mould of the ventral valve of *Australospirifer hawkinsii* (Morris and Sharpe, 1846) in centre of picture (compare with Plate 10a, 8 & 9). The other fossils are of *Australocoelia palmata* (Morris and Sharpe, 1846), with an internal mould of a dorsal valve at the upper left side of the picture. x 1.3. B17798. Probably from the Port Louis area. Collection C Darwin, HMS Beagle. Photograph © Natural History Museum, London

Plate 10c: Common brachiopods of the Falkland Islands



Brachiopods in a shell-drift (coquinite). *Australocoelia palmata* (Morris and Sharpe, 1846); internal moulds of ventral valves, together with internal moulds of ventral valves of *Pleurochonetes falklandicus* and *Schellwienella sulivani.* x 1.2. B60284. Manybranch Harbour. Presented by R Baker, 1931. Photograph © Natural History Museum, London

Bay Member (Section 2.3.2.1) is thickest in the west. Trilobites tend to occur only as disarticulated or broken fragments in sandstone. In the Port North to Pebble Island area, by contrast, mudstone units are generally much thicker and some contain well-preserved trilobites in calcareous or pyritic nodules. This increase in mudstone units is also seen in East Falkland, again accompanied by a thickening of the whole sequence. Drifts of brachiopods are proportionately fewer in these thicker sequences, whereas bivalves and gastropods are more common. *Tentaculites* has been found only in the north of the Islands.

Overall, palaeocurrent directions indicate transport from south to north, but seem to have been more variable than in the underlying formations. Transport directions to the NE (Marshall, 1994b), SSE, SW, NW/SE and ENE/WSW are all indicated in various places. At Chartres, asymmetric ripples verging to the south-west overlie symmetrical ripples indicating a palaeocurrent to north-west or south-east. Scasso and Mendia (1985) found a bimodal palaeocurrent distribution at Shag Cove, with a main direction to the north and NNE, and a secondary direction of SSW. Note, however, that their interpretation of the Fox Bay Formation includes strata that are here included in the Fish Creek Member or the Port Philomel Formation.

The type area is around Fox Bay, especially in the coastal sections between Cheek's Creek [TC 9025 3982] and the eastern part of Kelp Point [TC 917 384]. Reference sections for the base occur at: Two Pass Stream at the south-east end of Mile Ridge [TC 8511 7619]; the south-east end of Port North [TC 6555 8900]; the south-west side of The Knob, Port San Carlos [UC 6420 9170], and on the north side of Chabot Creek, Johnsons Harbour [VC 26 94]. Other reference sections are: the East Bay coastline from [TC 7540 5725] to [TC 7515 5819]; Brown Harbour from [TC 7300 5435] to [TC 7260 5460]; and the north-west side of Curlew Creek, north-east from [UC 6621 8935].

The following localities amongst others have been proved to include fossiliferous strata. On West Falkland: Green Hill Pass Pit, near Port Howard [UC 1773 8655]; Saddle Farm road quarry [UC 0251 7852]; Little Chartres Farm road quarry [TC 9579 6118]; Fox Bay East Road quarry [TC 9010 3970]; Rapid Point, South Harbour [TD 9040 0105]; Big Rincon, White Rock Bay [4265 9790]; Port North coast section [TC 65 90]; Pebble Island coastal section SW to SSW of First Mountain [UD 14 13]. The last section, which is accessible only at low tide, appears to have been the source of many excellent specimens collected by Constance Allardyce, wife of Governor Allardyce in the early years of the 20th century, which are housed in the New York State Museum. On East Falkland: the coast of Berkeley Sound between Green Patch and Port Louis, Caneja Creek road quarry, Horseshoe Bay [VC 12 91]; Dans Shanty Creek [VC 16 91], Kings Creek [UD 90 00], and Port San Carlos [UD 62 92].

The argillaceous lower part of the Bokkeveld Group of South Africa (the Ceres Subgroup) contains abundant marine invertebrate fossils of the Malvinokaffric Province (Broquet, 1992) and is here considered broadly equivalent to the Fox Bay Formation. Cooper (1982) argues that the lower, marine portion of the Bokkeveld Group is late Emsian to Eifelian in age, that is latest Early Devonian to earliest Middle Devonian. A marked eustatic sea-level rise occurred in the late Emsian (Cooper, 1982; Broquet, 1992), accounting for the abrupt marine transgression at the base of the Fox Bay Formation. Edgecombe (1994) considered that the calmoniid trilobites in the Falklands confirm a late Emsian age indicating correlation with the Gydo Formation (lowest Bokkeveld Group) of South Africa. Allied assemblages of trilobites occur in the Ponta Grossa Formation, Paraná, Brazil, parts of the Belén Formation, Bolivia, and the Talacasto Formation, Argentina. A similar faunal assemblage to the Fox Bay Formation occurs in the Horlick Formation, Ohio Range, Antarctica, although calmoniid trilobites are absent (Cooper, 1982).

2.3.2.1 East Bay Member

The East Bay Member is mostly composed of fine to medium-grained sandstones, with some siltstones and mudstones. It has been mapped in the area north and north-west of Fox Bay but could be more widespread in West Falkland, especially in the Coast Ridge area.

The East Bay Member is a newly recognised topmost subdivision of the Fox Bay Formation. The name is derived from East Bay, the type area. The unit is conformable with the rest of the Fox Bay Formation, below, and the Port Philomel Formation, above.

The East Bay Member can be regarded as the 'passage beds' between the typical developments of the Fox Bay Formation and of the Port Philomel Formation in Baker's (1924) concept. It displays characteristics of both. It is thought to be present only in the relatively proximal sequences of central West Falkland, where there is probably a continuous gradation between the two formations. Elsewhere the contrast between them is fairly clear cut. On lithological or sedimentological criteria the East Bay Member could be placed with either the Fox Bay or the Port Philomel Formations,

Plate 11: Typical trilobites of the Falkland Islands

1 *Metacryphaeus allardyceae* (Clarke, 1913). x 1.5. syntype NYSM 9697. Dorsal view of cephalon, internal mould with traces of cuticle. Pebble Island, West Falkland.

2 *Metacryphaeus allardyceae* (Clarke, 1913). x 1.5. syntype NYSM 9698. Dorsal view of outstretched thoracopygidium, latex cast from external mould. Pebble Island, West Falkland.

3,4,5 *Bainella nilesi* Edgcombe, 1994. x 1.5. holotype NYSM 5628. Anterior, dorsal and lateral views of partly exfoliated cephalon. Pebble Island, West Falkland.

6 *Bainella nilesi* Edgcombe, 1994. x 1.5. paratype NYSM 16125. Dorsal view of partial thorax, internal mould. Pebble Island, West Falkland.

7 *Bainella nilesi* Edgcombe, 1994. x 1.5. paratype NYSM 16129. Dorsal view of pygidium, largely exfoliated. Pebble Island, West Falkland.

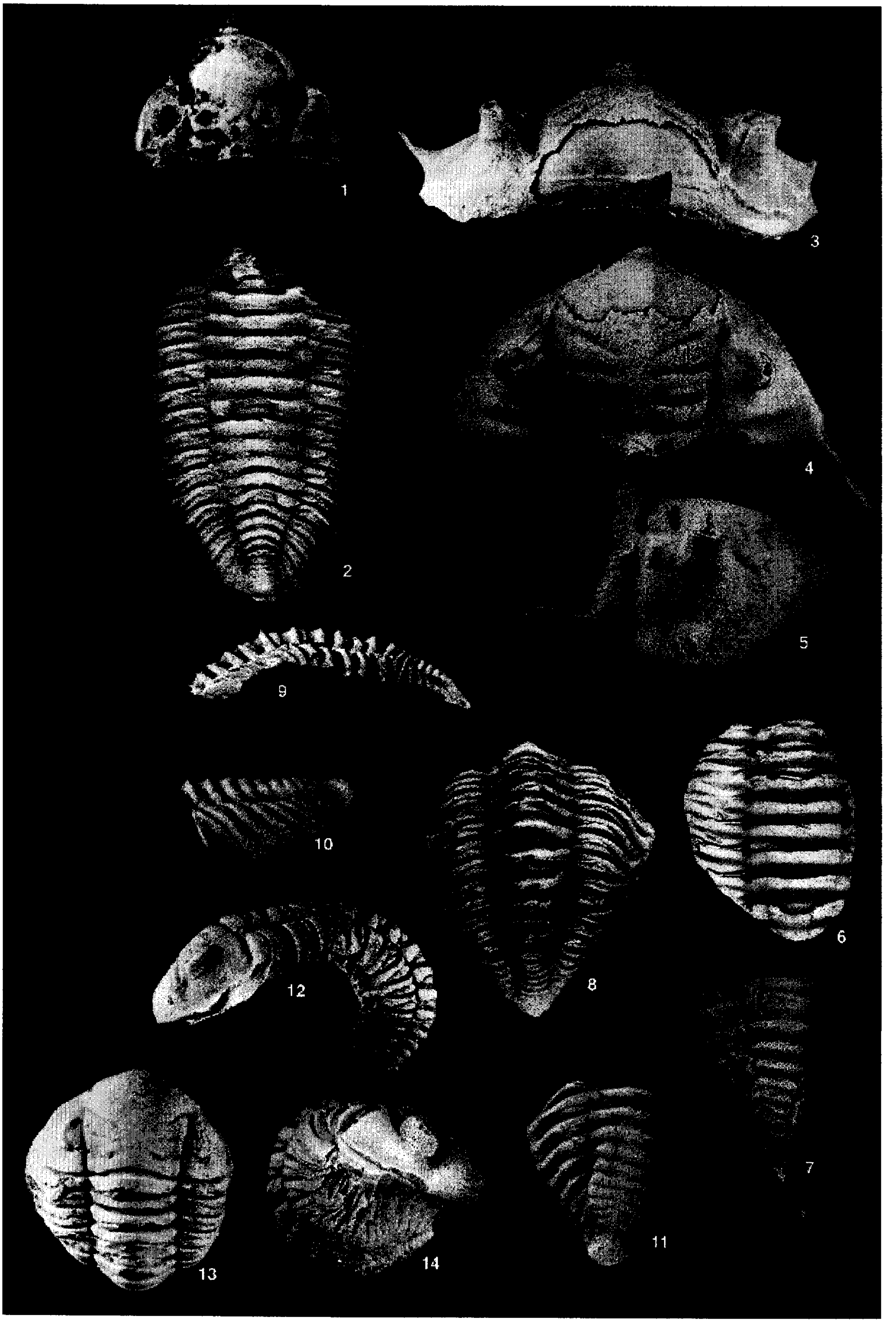
8,9 *Bainella falklandicus* (Clarke, 1913) x 1.5. lectotype NYSM 9722. Dorsal and lateral views of thoracopygidium, internal mould. Fox Bay, West Falkland.

10,11 *Bainella falklandicus* (Clarke, 1913) x 1.5. paralectotype NYSM 9721. Dorsal and lateral views of pygidium, plastic cast from external mould. Fox Bay, West Falkland.

12,13 *Oosthuizenella ocellus* (Lake, 1904) x 2. hypotype NYSM 9646. Lateral and dorsal cephalic views of partly enrolled exoskeleton, internal mould. Pebble Island, West Falkland.

14 *Oosthuizenella ocellus* (Lake, 1904) x 2. hypotype NYSM 9647. Lateral view of enrolled exoskeleton, internal mould. Pebble Island, West Falkland.

All photographs in this plate courtesy of G D Edgecombe from specimens in the New York State Museum. First published in Edgecombe (1994), New York State Museum Bulletin, which includes more illustrations of these and similar specimens.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but the occurrence of marine invertebrate fossils in the East Bay Member links it more closely with the Fox Bay Formation.

The East Bay Member is distinguished from the underlying main development of the Fox Bay Formation by the dominance of fine to medium-grained sandstones over finer-grained lithologies, the common occurrence of cross-lamination and cross-bedding and the absence of hummocky cross-stratification. The East Bay Member tends to form somewhat more hilly ground than the Fox Bay Formation, from which it is separated by a negative break of slope.

Like the overlying Port Philomel Formation, the East Bay Member is characterised by varied bedforms, including lamination in sandstones and finer-grained lithologies, cross-lamination and ripple marks, cross-bedding in trough, wedge-shaped and tabular forms, and channel fills. Trough cross-sets up to three metres across are usual. However, the range of rock-types seen in this member is similar to the underlying part of the Fox Bay Formation. In particular the sandstones tend to resemble those in the Fox Bay Formation more closely in colour, texture and mineral composition than those in the Port Philomel Formation. However, no black shales are seen in the East Bay Member.

There are sporadic occurrences of marine invertebrate body fossils, especially brachiopods, notably in the Spring Point area (Clarke, 1913a), e.g. Letterbox Point, north of Spring Point [TC 64 58]. No trilobites have been noted in this member. Plant debris is absent. Bioturbation is seen in some beds, particularly in muddy and silty lithologies, but also in sandstones.

The East Bay Member is about 250 metres thick.

The type area lies around Strawberry Hill, between East Bay and Brown Harbour. Reference sections are: East Bay coast from [TC 7515 5819] to Rees Harbour [TC 7420 5910]; Brown Harbour coast from Rose Bowl Point [TC 7260 5460] to Reserve Island Point [TC 7080 5545]; Tussac Point [TC 6092 4745] northwards to Spring Point [TC 6065 5265].

2.3.3 Port Philomel Formation

The Port Philomel Formation is a lithologically varied assemblage of medium and fine-grained sandstones, siltstones and mudstones. Its distribution in central and northern West Falkland and in northern East Falkland is similar to that of the Fox Bay Formation and the Port Stanley Formation. Inliers of the Port Philomel Formation in the Wickham Heights may be more numerous and extensive than shown.

Baker (1924) identified an interval of 'soft yellow sandstones, thin-bedded micaceous sandstones and greyish-brown sandy shales' with 'lepidodendroid plant remains', which occur above the fossiliferous Fox Bay Formation and below the quartzites of the Port Stanley Formation. He named these strata the 'Port Philomel Beds', noting that there is no clear line of demarcation with the underlying strata with marine fossils. Borello (1963; 1972) grouped the Port Philomel Formation and the Port Stanley Formation together as the 'Formación Monte Maria', but this was based on a misunderstanding of the geology of the Mount Maria area (Scasso and Mendia, 1985). Greenway (1972) had insufficient ground information to separate this interval from the Fox Bay Formation, and it was not identified by Marshall (1994b). Nevertheless, the Port Philomel Formation (*sensu* Baker 1924) appears to form a distinct unit throughout the Islands, lying conformably between the Fox Bay Formation and the Port Stanley Formation.

The Port Philomel Formation tends to form hills and to underlie steeper slopes than adjacent parts of the Fox Bay Formation. This contrast in associated topography suggests that there is some consistent difference in lithofacies. In many parts of West Falkland, the ground below the base of the Port Stanley Formation is characterised by several well-marked steps in the topography, each comprising a well-marked positive feature with a corresponding negative feature a short distance below it (Plate 14). The positive features are taken to mark resistant sandstones, whereas the negative features are assumed to mark intervals dominated by mudstone or other soft lithology. The appearance of these topographic features changes along strike. In some places only two steps can be distinguished, but in others there are as many as five.

Where exposed, the Port Philomel Formation is distinguished from the Fox Bay Formation by the overall increase in medium-grained sandstones, the greater variety of bed-forms, the absence of marine invertebrate fossils and, generally, by the common occurrence of plant macrofossils. In the Port Philomel Formation, medium and fine-grained, cross-bedded and ripple-marked sandstones predominate over plane-bedded and laminated fine sandstones, siltstones and mudstones, and the colour of fine- to medium-grained sandstones changes from mainly brown and reddish-brown in the Fox Bay Formation to mainly yellowish brown and brownish orange.

Nevertheless, as exposed in West Falkland, there appears to be a complete sedimentological gradation between the Fox Bay Formation and the Port Philomel Formation. Therefore it could be argued that the Fox Bay Formation and the Port Philomel Formation cannot be separated lithostratigraphically and so should be treated as a single unit. However, the typical Fox Bay Formation is quite different to the typical Port Philomel Formation, and so it seems useful to maintain the division proposed by Baker (1924). The two formations are most easily distinguished by the marine invertebrate fossils in the former and the abundant plant macrofossils in the latter. Although given the nature of the sequence it seems possible that the ranges of these fossil groups actually overlap, so far no sections have been found where this is the case. However, in some areas (such as White Rock Bay and Mount Rosalie House), and perhaps in most, the appearance of conspicuous plant fossils does not coincide with a distinctive topographic feature. This suggests that it does not coincide with a significant lithofacies change, and in practice it means that this horizon cannot be mapped.

The base of the Port Philomel Formation is instead taken at the negative feature (that is, most probably at a persistent bed of mudstone) which marks the change from the subdued landscape typical of the Fox Bay Formation to the step-like topography underlain by the Port Philomel Formation. It is assumed that the main mudstones will prove to be the most reliable markers for local and international correlation, which in this case occurs at a significant change in lithofacies. Where the incoming of abundant plant debris can be placed with confidence, it is found to lie a short and fairly consistent interval below the mapped boundary.

The Port Philomel Formation comprises soft yellowish brown and brownish orange to grey or black subarkosic sandstones with greyish-brown to black siltstones, shales and mudstones. The sandstones are typically fine to medium-grained and moderately to well-sorted, but some argillaceous sandstones and coarse sandstones also occur. Detrital mica (both black and white types occur) is abundant in some sandstone and shale beds but is less generally pervasive than in the Fox Bay Formation. Some mudstones are carbonaceous. Isolated pyrite nodules of a few centimetres diameter are characteristic. They are often enclosed by a broader collar of rust-coloured relatively resistant sandstone. Organic-rich mudstone and small lenses of coal have been noted on Saunders Island and Carcass Island.

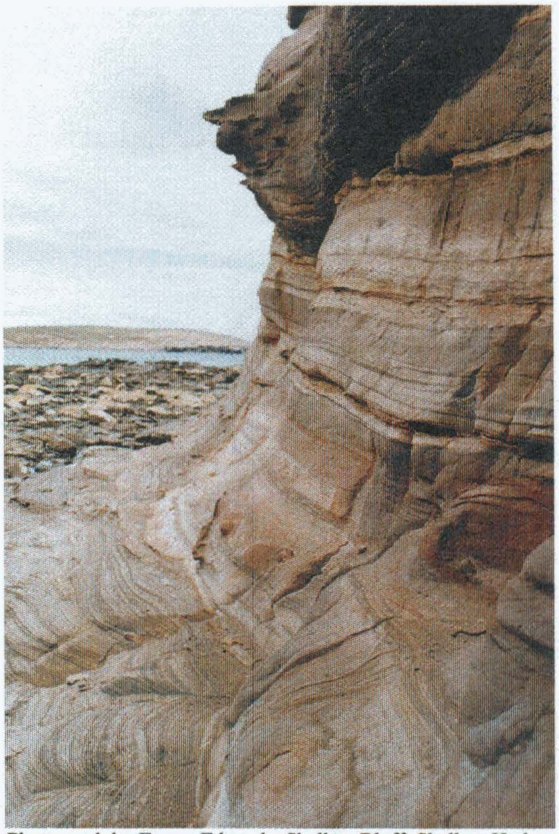
Petrographic examination of two specimens of subarkosic sandstone from the Port Philomel Formation, one fine-grained, the other fine to medium-grained and ferruginous, found that both were moderately well sorted, and composed mostly of monocrystalline quartz, abundant feldspar, micas (biotite and muscovite) and sparse rock fragments (chiefly cherts). One had moderately common zircon and was cemented by clay. The other was cemented largely by pyrite.

The Port Philomel Formation is characterised by very varied bedforms, including medium to thin plane-bedding and lamination (Plate 12), locally with primary current lineation, cross-lamination and ripple marks including climbing ripples, flaser lamination, cross-bedding in trough, wedge-shaped and tabular forms, and small channel fills, including mud-clast conglomerates. Stacked channels occur locally. Cross-sets are generally not as large as in the Port Stanley Formation, being typically in the range 20 centimetres to one metre thick, with trough cross-sets commonly 50 centimetres to one metre wide. Some are truncated at the top, others not, sometimes within the same section. De-watering structures are locally present. The sequence displays fining-up cycles, generally of the order of tens of metres in thickness. Examples seen at Hill Gap commence at an erosive surface with trough cross-bedded sandstones, passing up to ripple cross-laminated and laminated sandstone at the top, although some cycles also include mudstone.

The Port Philomel Formation characteristically contains fragments of lycophyte plants and indeterminate plant debris which is locally very abundant. Some medium to fine-grained sandstones contain abundant fragmentary debris represented by carbonaceous smears of negligible thickness and little more than 10 mm^2 in area. Three-dimensional segments of lycopsid stems of types allied to *Lepidodendron* are widespread and locally abundant (Plate 13). These are up to 30 centimetres long and five centimetres in diameter, although usually less. The stems are characteristically oval in cross-section, with one apex of the oval more rounded than the other. Most show a closely spaced pattern of diamond-shaped leaf scars. Specimens of these plant fossils collected by Baker and by Halle are discussed by Seward and Walton (1923).

Clarke (1913a) records the existence of indeterminate 'fish plates' (presumably parts of the bony carapace commonly found in Devonian fish, rather than anything resembling fish scales) in sandstones from Halfway Cove, near Dunnose Head settlement [TC 64 61]. Short intervals of bioturbated mudstone and siltstone similar to those of the Fox Bay Formation are present, especially in the lower part of the formation where they locally alternate with plant-bearing beds, but marine invertebrate body fossils are not observed.

Plate 12: Laminated sandstones, Port Philomel Formation



Photograph by Emma Edwards, Shallow Bluff, Shallow Harbour/Dunnose Head

Plate 13: Plant fossils, including lycopsid stems, Port Philomel Formation

Photograph by Emma Edwards, Sugar Loaf, Saunders Island

The range of lithological types, the varied sedimentary structures and the abundance of plant debris together suggest deltaic deposition. Subtidal and shoreface deposition may also be represented. Palaeocurrent directions were generally to the north-east but varied between north-west and south-east.

Estimates of the thickness of the Port Philomel Formation vary from 130 metres in Mount Philomel, 200 to 300 metres around Green Hill, Mount Adam, Hill Cove and Port North, and about 350 metres in the Coast Ridge and in East Falkland. Note that strata here described as the Port Philomel Formation have previously been included in either the Fox Bay Formation or the Port Stanley Formation (probably not consistently) either for lack of evidence (Greenway, 1972) , or because not recognised as a distinctive unit (Marshall, 1994). For this reason, previous assessments of the thickness of the Port Philomel Formation and the two adjacent formations should be treated with circumspection.

The type area is the coast between Shallow Harbour Creek [TC 5745 5920] and Isthmus Cove, Port Philomel [TC 7070 6640]. Reference sections can be found at Shallow Bluff [TC 5370 5818] and the coast south to Mare Rincon [TC 5415 5515]; on the north-east side of Port North from Stevelly Bay [TC 6070 9500] to Mount Brown Beach [TC 6422 9132]; and on the west side of Curlew Creek [UC 6626 8895] to [UC 6621 8935]. Sections exposing the upper part only are seen on Settlement Beach, Carcass Island [TD 5290 1160] to [TD 5240 1129]; and on the north-east slope of Sugarloaf Hill, Saunders Island [TD 8660 1091] to [TD 8630 1058].

The age of the Port Philomel Formation is constrained to the Middle and Late Devonian by evidence from the adjacent formations (Sections 2.3.2 and 2.3.4). Seward and Walton (1923) concluded from the 'slender evidence' available to them that the plant fossils in this part of the sequence were probably of Middle Devonian age. Cramer et al. (1972) report the discovery of Lower Givetian (Middle Devonian) palynomorphs in a slab of shaley siltstone containing plant macrofossils, which was collected by Halle (1912) from Halfway Cove, Dunnose Head [TC 648 616].

In the eastern Cape Province, the Traka Subgroup (the more arenaceous upper part of the Bokkeveld Group) contains plant fossils, the remains of freshwater fishes and the remains of marine trace fossils (Broquet, 1992). It thus seems a likely correlative of the Port Philomel Formation, although the succeeding Weltevrede Formation (Witteberg Group) also contains plant fossils. Thus the base of the Witteberg Group might be correlatable with a level within the Port Philomel Formation. As with the Fox Bay Formation, more exact correlation requires a more detailed understanding of the sequence.

2.3.4 Port Stanley Formation

The Port Stanley Formation is mostly composed of medium-grained quartz sandstones, some of them quartzitic, with some fine-grained sandstones, siltstones and mudstones. Some beds of the hard, well-cemented sandstones form characteristic craggy exposures and are known as 'Stanley Quartzite'. The Port Stanley Formation crops out in a broad belt extending the width of East Falkland, with three outliers in the north. It is also widely distributed in the west, forming the ridges which run the length of the east coast of West Falkland, the hills ranging from Mount Rosalie westwards to West Point Island, the high ground on many of the nearby islands, the Dunnose Head peninsula, and outliers on Mount Philomel and Mount Sulivan.

Baker (1924) referred to the 'upper quartzitic sandstone series' which forms the 'heights of the northern part of the Colony' as the Port Stanley Beds. Borello (1963; 1972) grouped the Port Philomel Formation and the Port Stanley Formation together as the 'Formación Monte Maria', but as pointed out by Scasso and Mendia (1985) neither formation occurs in the Mount Maria area. Baker's 'Port Stanley Beds' were named the 'Formación Caleta Shag' by Scasso and Mendia (1985), who based their description on sections exposed at Shag Cove, in which they distinguish a lower and an upper member. Their members approximately correspond to divisions of the Port Stanley Formation seen in the Dunnose Head area, but conform only to the lower cycle of sedimentation seen in the Bold Cove area. Thus the Shag Cove section is less representative of the Port Stanley Formation as a whole than would be sections in the Manybranch Harbour area, for example. The term 'Port Stanley Formation' is preferred here, following the precedent of Baker (1924) and the style of the names 'Port Stephens Formation' and 'Port Philomel Formation'. The Port Stanley Formation conformably overlies the Port Philomel Formation. It is overlain disconformably by the Bluff Cove Formation in East Falkland and probably unconformably by the Fitzroy Tillite Formation in West Falkland. Further work could subdivide the Port Stanley Formation in some areas.

The Port Stanley Formation everywhere underlies higher ground and forms steeper slopes than the Port Philomel Formation, including many of the highest summits in the Islands, but it is generally covered by stone runs and other solifluction deposits. Craggy exposures with visible bedding traces are common, especially on ridge crests. In the Mount Adam range, and particularly to the west of Hill Cove, the Port Stanley Formation can be subdivided by the occurrence of persistently exposed quartzite layers and intervening negative breaks of slope. Some of these topographic features can be traced intermittently on aerial photographs for up to 20 kilometres along strike. Although most eventually disappear laterally, at least two cycles of sedimentation can be distinguished within most parts of the Port Stanley Formation. In some areas of West Falkland, as between Port Howard and Port North, each cycle is capped by a well-defined layer of hard quartzite which forms large concordant sheet-like exposures. Around the head of Bold Cove, the upper of these two layers lies immediately beneath the base of the Fitzroy Tillite Formation. On the west side of Bold Cove, the exposures of the upper quartzite layer gradually diminish in size, and they disappear entirely south of Bold Cove. Similarly, only one thick quartzite layer occurs in the Port Stanley Formation in East Falkland. It is not clear if this is a result of lateral variation in the Port Stanley Formation or of erosion prior to the deposition of the Lafonia Group. In the Dunnose Head peninsula, the strata of the lower cycle can be divided photogeologically into two units, designated 'A' and 'B', but where the Port Stanley Formation forms steep ground, the additional break of slope which separates the units seen on the Dunnose Head peninsula is either absent or hidden beneath solifluction debris.

The Port Stanley Formation can generally be distinguished from the Port Philomel Formation by the more uniform appearance and greater mineralogical maturity of the quartz sandstones. In exposed sections, the base of the Port Stanley Formation is taken at the appearance of abundant conspicuously cross-bedded, generally medium-grained quartz arenites or subarkosic sandstones with only rare shales, plant fossils, fine-grained ripple cross-laminated and laminated sandstones. Pyrite nodules and mudclasts virtually disappear at the top of the Port Philomel Formation. As exposed on the foreshore near Carcass Bay settlement, the base of the Port Stanley Formation is erosive, resting on a 10 metre thick unit of dark grey laminated mudstone with minor siltstone and sparse beds of about 20 centimetres of cross-bedded sandstone.

Elsewhere, the same mudstone seems to coincide with a distinct negative break of slope which has been taken as the base of the Port Stanley Formation for mapping purposes. It lies at the base of the steep convex slopes strewn with quartzite debris and with conspicuous exposures of quartzitic sandstone which are typical of the Port Stanley Formation outcrop in many areas. In most places this negative topographic feature does not have a clearly marked positive feature a short distance above it, as seen in the Port Philomel Formation outcrop (Section 2.3.3). On West Falkland, the mudstone forming the negative feature is exposed in a stream gully on the south side of Mount Robinson [TC 9697 7769] (Plate 14) and on the north-west side of Death's Head [TC 46 98] (seen from West Point Island). It is very probably the same horizon taken as the base of the Port Stanley Formation by Marshall (1994b) in the Port North/Dunbar section. On East Falkland, several metres of laminated mudstone exposed in Curlew Creek [UC 6626 8895] coincide with the negative topographic feature taken as the base of the Port Stanley Formation.

The most characteristic components of the Port Stanley Formation are unfossiliferous, mainly mature, medium-grained, locally coarse sandstones. Many are quartzitic, forming rib-like or sheet-like exposures. However, these resistant rock-types form only a small part of the sequence. The intervening strata are mostly medium-grained sandstones, with some fine-grained sandstones, micaceous muddy siltstones and silty shales at intervals throughout. Most of the sandstones are quartz arenites but subarkoses are also common. They are mostly pale grey, but can be whitish, brownish or reddish. Most are made up of well-sorted, moderately to well-rounded quartz grains, and pebbles are rare. The sandstones typically contain accessory biotite and heavy minerals (such as rutile and zircon), the latter commonly being concentrated on bedding planes. White mica is not usually a conspicuous component. Disseminated pyrite occurs very locally in the sandstones.

Petrographic examination of four fine, medium or coarse-grained sandstones from the Port Stanley Formation found that they were all mostly composed of monocrystalline quartz grains, some with subordinate feldspar and muscovite mica. A few polycrystalline quartz grains and heavy mineral grains occur in the coarsest example. The fine-grained sandstone also included chlorite, a variety of non-opaque heavy minerals and common pyrite aggregates. The sandstones were all well-cemented by silica overgrowths on the primary grains, with patchy clay cement as well in some specimens (Lott, 1997; Lott, 1999).

The Port Stanley Formation sandstones typically display tabular or wedge-shaped planar cross-bedding alternating with medium to thick plane-bedded sandstones (Plate 15). Massive beds up to three metres thick have been recorded, but these are exceptional. There is some trough cross-bedding, which locally predominates. Isolated examples of overturned or

Plate 14: Mount Robinson from the south

Photograph by Don Aldiss, BGS. MN28241

Typical topographic styles associated with the upper West Falkland Group in West Falkland. The smooth convex slope from the skyline down to the upper negative feature is underlain by Port Stanley Formation. The ground including two prominent terraces below that negative feature is underlain by the Port Philomel Formation, the base of which approximately coincides with the lower edge of the large shadow near the centre of the photograph.

Plate 15: Typical cross-bedded quartzites, Port Stanley Formation

Sapper Hill, Stanley. Photograph by Don Aldiss, BGS. MN28242

Plate 16: Channel-fill structure in lower Port Stanley Formation

Death Valley, near Gun Hill, West Falkland. Photograph by Don Aldiss, BGS. MN28243

slumped cross-beds are fairly common. Mud flakes are seen in places. Cross-sets are commonly in the range 0.5 to 1.5 metres thick, but locally are as thin as 10 centimetres, and generally between one metre and 10 metres wide. Co-sets tend to be planar and are from one metre to five metres thick. A channel some 25 metres across and five metres deep, with cross-bedded sandstone infill, is exposed in Death Valley, near Gun Hill [TC 785 686] (Plate 16). Rare straight-crested asymmetrical current ripples occur at the same locality. Plane bedded sandstone with primary current lineation does occur in the Port Stanley Formation, but seems to be rare. Laminated and cross-laminated sandstones are a widespread minor component. In the Dunbar area cross-laminated sandstone with climbing ripples, lenticular and flaser bedding accompanies the usual types of cross-bedded sandstone.

Lycopsid stems and faintly carbonaceous leaf debris very similar to the plant fossils seen in the Port Philomel Formation are locally abundant in sandstones in the lowest part of the Port Stanley Formation, corresponding to the lower division noted in the Dunnose Head peninsula. This lowest, plant-bearing division of the Port Stanley Formation can be seen in other places in the west, for example in the Coast Ridge and on the summit of Green Hill, north-west of Port Howard. Poorly preserved plant debris also occurs in the rarely exposed mudstones elsewhere in the sequence, for example in Mount Pleasant quarry. Bioturbation and trace fossils are rare in the Port Stanley Formation but M. Hunter (oral communication, 1998) found oblique burrows a few millimetres in diameter in the Stanley Quartzite near Cape Pembroke. N. Meadows (oral communication, 1996) observed bioturbation in the Port Stanley Formation near Bluff Cove.

The Port Stanley Formation can be interpreted as the product of marine shoreface environments including tidal channel deposits. As in the rest of the West Falkland Group, palaeocurrent directions are generally to the north but vary between west-south-west and east-north-east. Curtis and Hyam (1998) found that the palaeocurrent distribution in the Port Stanley Formation near Bold Cove varied from unimodal to trimodal within a short distance (Section 4.2.2.1; Figure 4.4). Scasso and Mendia (1985) found northerly palaeocurrents near Stanley but a bimodal distribution (NNE and SE) at Shag Cove. Hiller and Taylor (1992) concluded that the Witpoort Formation of Eastern Cape Province, South Africa, which is closely similar to the Port Stanley Formation, was deposited in a barrier island system on a linear shoreline. The change from deltaic deposition was caused by falling sea-levels, reversing the trend seen since the onset of Bokkeveld Group/Fox Bay Formation deposition.

The thickness of the Port Stanley Formation is estimated as 400 to 500 metres in most of West Falkland, increasing to 700 metres around Mount Rosalie, and up to 1100 metres in East Falkland.

The hills and coastlines around Port Stanley (and therefore also around the town of Stanley), East Falkland, comprise Baker's (1924) type area of the Port Stanley Formation, although neither the base nor the top of the formation are well-exposed in that vicinity. A type section is more likely to be found in the less deformed sequences on West Falkland. Reference sections for the base only can be found on the north-east slope of Sugarloaf Hill, Saunders Island [TD 8630 1058] to [TD 8591 1000]; and on Settlement Beach, Carcass Island [TD 5240 1129] to [TD 5221 1090]. Other reference sections, which include the base, are exposed on the western side of Curlew Creek from [UC 6626 8895] to [UC 6582 8805]; at Hill Gap [UC 1059 5645] to [UC 1085 5615]; and at Shag Cove [UC 1790 6424] to [UC 1830 6420]. The lowest parts of the formation are usually obscured by solifluction debris from the slopes above but good sections are exposed in Death Valley, near Gun Hill [TC 785 686].

Palynological studies by Marshall (1994b) found that the top of the Port Stanley Formation at Dunbar is Fammenian (latest Devonian) in age.

The lower part of the Port Stanley Formation is equivalent to at least part of the Givetian to Frasnian Weltevrede Formation (early Witteberg Group of South Africa) which is a mixed sequence of sandstones, siltstones and mudstones with plant fossils and marine trace fossils (Broquet, 1992; Hiller and Taylor, 1992). The base of the Weltevrede Formation might lie at a level within the Port Philomel Formation. The Witteberg Group is divided by the orthoquartzites of the Witpoort Formation (Witteberg Group), which are Fammenian in age (Hiller and Taylor, 1992), and which are very similar to the Stanley Quartzite, in the upper part of the Port Stanley Formation. Hiller and Taylor (1992) record that overturned and contorted cross-bedding, as seen in the Stanley Quartzite, is a feature of the upper part of the Witpoort Formation. The Early Carboniferous portions of the Witteberg Group (the Lake Menz and Kommadagga subgroups) (Broquet, 1992) appear not to be represented in the Falkland Islands, except, possibly, in the highest parts of the Port Stanley Formation in East Falkland, which are commonly not exposed.

2.3.5 Presumed West Falkland Group strata of Beauchêne Island

Beauchêne Island is the most southerly and isolated island in the Falkland Islands archipelago. Although Lewis Smith and Clymo (1985) noted that it is made of quartzite, the geology of the Island had never been examined in detail prior to the recent survey (Edwards, 1997).

The island is up to four kilometres long and one kilometre wide. The bedrock is all very gently dipping quartzitic sandstones and quartzites, with a total of about 60 metres of strata exposed. Well developed channel structures occur in the top 35 metres of the sequence, overlying a 10 metre interval of planar-bedded quartzites with symmetrical ripple marks. The lowest strata are generally low-angle trough cross-bedded quartzites and arkosic arenites. No trace fossils or body fossils have been observed.

The sandstones are generally mature and well sorted. Field examination indicates that porosity and permeability are low, although secondary porosity may be moderate in some of the more feldspathic sandstone units, due to the dissolution of the feldspar minerals. Petrographic examination of one specimen of medium to coarse-grained quartzite found that it was composed largely of poorly sorted monocrystalline quartz grains. Although there was extensive quartz overgrowth of the primary grains, and some pore-filling clay, a poorly connected secondary porosity resulted from dissolution of unstable grains such as mica or feldspar (Lott, 1999).

The upper strata are very coarse to medium-grained quartz arenites to subarkosic quartzites with a silica cement. The primary grains are well rounded with high sphericity. Wedge-shaped trough cross-bedded units, showing bundled up-building are common in this part of the sequence.

The planar bedded unit in the middle of the section is composed of fine to medium-grained arkosic arenites and quartzites. The grains are well rounded and very well sorted, with high sphericity. Straight-crested symmetrical ripple marks were found in one place.

The lowest strata have a greater percentage of feldspar minerals than the others, and are formed of coarse to medium-grained arkosic sandstones and quartzites. The lithologies in the lowest unit are poorly to moderately sorted with well-rounded grains. Well-developed channel lag deposits with gravel-sized grains were observed in the trough cross-bedded units.

Palaeocurrent directions, taken from ripple marks and trough cross-bedding, show a transport direction of N280° to N040° throughout the sequence.

The nearest land to Beauchêne occurs some 50 kilometres to the north where Permian strata of the Lafonia Group are exposed on Sea Lion Island. However, sedimentary petrography and heavy mineral analysis show that the Devonian and the Permian sandstones in the Falkland Islands are quite different to each other (Lott, 1997; Knox, 1997). The quartzitic sandstones of Beauchêne are coarser and less feldspathic than the Permian sandstones, and they contain none of the volcanic detritus found in the Permian strata.

On the other hand, the rocks on Beauchêne Island are very similar in lithology and sedimentary structure to the quartzites found in typical developments of the Port Stephens Formation and the Port Stanley Formation. The palaeocurrent directions found on Beauchêne Island are similar to those measured in those formations on the main islands. However, it is not clear to which of these two formations the Beauchêne quartzites belong and so they have been designated simply as 'West Falkland Group, undifferentiated' on the geological map. Coarse-grained quartzites are more commonly found in the Port Stephens Formation, as well as the types of sedimentary structure observed on Beauchêne. Moreover, the Port Stephens Formation is considerably thicker than the Port Stanley Formation and is therefore the more likely to be exposed. Heavy mineral analysis confirms that the Beauchêne quartzites are dissimilar to any Permian sandstones, and also that they have different characteristics to the Fox Bay Formation and Port Philomel Formation. However, the closest resemblance in terms of heavy mineral content is with Port Stanley Formation quartzites from West Falkland (Knox and Aldiss, 1999). Thus the Beauchêne quartzites may represent a coarser, more proximal facies of the Port Stanley Formation, which has perhaps been tectonically displaced from its original position relative to West Falkland. Nonetheless, overall it seems more likely that the Beauchêne quartzites belong to the Port Stephens Formation.

2.4 Lafonia Group

The Lafonia Group encompasses a thick and varied sequence of sedimentary strata of Carboniferous and Permian age. Although dominated by medium and fine-grained sandstones, especially in the upper part, it also includes siltstones, mudstones (including some carbonaceous mudstone), rare tuffs and a thick tillite sequence. It is most extensive in Lafonia and the rest of East Falkland south of the Wickham Heights, but the older component formations also occur as outliers within the fold belt of East Falkland. The Lafonia Group is also present in West Falkland, occurring mainly on the east flank of the Coast Ridge but also in Port Purvis and at the east end of Byron Sound. The sequence in West Falkland is much thinner than in the east (Section 4.3), and the oldest (Bluff Cove Formation) and youngest (Bay of Harbours Formation) units of the Lafonia Group are absent there. This variation was possibly partly controlled by growth faulting in the Falkland Sound Fault Zone (Section 4.2.3.2). Note, however, that the estimated formation thicknesses given for East Falkland in the following sections tend to be greater than for their correlatives in South Africa. While this might truly reflect a deepening of the basin, perhaps due to fault movement in what was to become a rift between Africa and East Antarctica, some of the estimates may be too large.

Halle (1912) proposed the term 'Lafonian series' for the Permo-Carboniferous strata found south of the Wickham Heights. Although broadly equating it with the *Glossopteris*-bearing beds of Lafonia, he also included the tillite and other beds overlying the Devonian quartzites. His descriptions indicate that he also would have included the Bluff Cove Formation in the Lafonian series, had he recognised it. Baker (1924) divided the same strata into an 'Upper Lafonian Series' and 'Lower Lafonian Series' taking the base of the Upper Lafonian at the base of his 'Choiseul Sound and Brenton Loch Beds', above his 'Lafonian Sandstone'. He took the base of the lower Lafonian at the base of his 'Bluff Cove (Fitzroy Basin) Beds'. Greenway (1972) referred the Permo-Carboniferous sequence to the 'Lafonian Supergroup' dividing it into the 'Lower Lafonian Group' and the 'Upper Lafonian Group'. Although following Baker (1924) in placing the boundary between the lower and upper divisions at the base of the Lafonian Sandstone, she had insufficient ground information to locate that boundary on aerial photographs.

There are several problems with the terminology used by Baker and Greenway. In the interpretation of the stratigraphy presented here, the boundary between the lower and upper divisions identified by Baker (1924) lies within a formation, the Brenton Loch Formation. A clearer and more significant boundary occurs at the base of that formation, but even so it is considered here that the contrast between the 'Upper Lafonian Group' and the underlying Port Sussex Formation is insufficient to justify placing them in separate groups and so treating the whole Permo-Carboniferous sequence as a Supergroup. Also, the three formations of the 'Lower Lafonian Group' are too dissimilar to put together by themselves in a group. Moreover, according to international conventions, 'Lower' and 'Upper' should not be used in formal names, and nor should an adjective such as 'Lafonian'.

Therefore it is suggested that all of Halle's 'Lafonian Series' should be placed in a 'Lafonia Group', informally divided into a 'lower' and an 'upper' portion. As here defined, the lower Lafonia Group comprises the Bluff Cove Formation, the Fitzroy Tillite Formation and the Port Sussex Formation. The upper Lafonia Group comprises the Brenton Loch Formation and the Bay of Harbours Formation. The alternative, to confine 'Lafonia Group' to the two younger formations, and to exclude the three older formations from any group, would be a confusing departure from Halle's (1912) concept. Note that although the type area of the Lafonia Group is the whole of East Falkland south of the Wickham Heights, Lafonia comprises that part of East Falkland south of a gorse hedge dividing the Goose Green isthmus. (The area was named after an Englishman, S F Lafone, who purchased it in 1846). Borello (1963; 1972) also placed the entire 'Lafonian Series' in a single group, his 'Grupo Isla Soledad', but there seems little value in abandoning Halle's precedent for a name which would be translated as 'East Falkland Group'.

The Lafonia Group everywhere overlies the Port Stanley Formation (West Falkland Group) disconformably or unconformably. No Mesozoic strata are known to occur onshore but the Lafonia Group is intruded by Early Jurassic dolerite dykes. Contrasts between the divisions of the upper Lafonia Group are most clearly seen in the vicinity of Brenton Loch in the central western part of East Falkland.

A major change in palaeogeography occurred between the deposition of the West Falkland Group and the Lafonia Group. The sandstones in the former are dominated by quartz and potassium feldspar, detritus derived from cratonic areas in the interior of Gondwana. Rocks throughout the Lafonia Group mostly include at least a small proportion (in the lower units, only traces) of detritus from a volcanic arc which lay on the margin of Gondwana. This is demonstrated by the admixture of sodic/calcic feldspars in the sandstones, K-bentonite in some of the mudstones, the composition of some lithic fragments in the tillite and in many of the sandstones, the presence of rare volcaniclastic beds, and by the contrasting heavy mineral assemblage (Knox, 1997; Lott, 1997; Lott, 1999; Kemp et al., 1998). Knox (1997) found that the heavy mineral assemblage of 10 sandstones from the upper Lafonia Group included more apatite and less rutile and monazite than the Devonian sandstones. Some of the Permian sandstones also included abundant epidote, garnet and sphene, and one had chrome spinel, an assemblage not found in the Devonian.

The same overall change is found in South Africa, where the Cape Supergroup represents a passive margin sequence with a non-orogenic continental provenance, and the Karoo Supergroup a rear-arc foreland basin setting. The Ecca Group (which includes the South African correlatives of the bulk of the Lafonia Group, above the tillite) has a magmatic arc provenance (Johnson, 1991; Johnson et al., 1996). A palaeoenvironmental model proposed by Kingsley (1981) for the south-eastern outcrop of the Ecca Group (i.e. the portion which lies closest to the inferred position of the Falkland Islands) is broadly applicable to the strata overlying the Fitzroy Tillite Formation. It postulates a sequence of basin plain, turbidity and deltaic deposits. The first phase of folding in the Cape Fold Belt (Section 4.2.1) occurred during deposition of the Dwyka Tillite, of which the Fitzroy Tillite is a lateral equivalent, and continued during Ecca Group times (Cole, 1992; Halbich, 1992). The youngest part of the Lafonia Group appears to be equivalent to the lower Beaufort Group (Section 2.4.5). Equivalents of the Lafonia Group are found in the Victoria Group (Beacon Supergroup) of Antarctica (Barrett, 1991).

The age of the component formations is discussed in the following sections. The oldest is thought to be late Carboniferous. A palynomorph assemblage from the youngest is comparable with material from the Permian Ecca Group of South Africa (Macdonald, 1996). Negative evidence also indicates that the youngest formation does not extend into the Triassic: the characteristic *Dicroidium* flora has not been found in the Falklands, nor is there any sign of the change to fluvial and aeolian sedimentation seen in the Late Permian and Triassic in the South African sequence (Johnson et al., 1996).

2.4.1 Bluff Cove Formation

The Bluff Cove Formation comprises fine-grained sandstones with siltstones and shales. Previously it was thought to be confined to the eastern part of East Falkland, but has now been shown to extend as far west as Falkland Sound and the Inner Verde outlier. It is absent in West Falkland.

Baker (1924) first recognised the 'locally occurring shales and sandstones' which occur beneath the tillite in the Bluff Cove area, naming them the 'Bluff Cove (Fitzroy Basin) Beds'. He found that they are absent in some areas and that shaley beds also occur within the tillite, so concluded that there is 'no definite series of shaley beds at the base of the [tillite] as there is in South Africa'. Nevertheless, Adie (1952a; 1952b) correlated his 'Bluff Cove Beds' with the Lower Dwyka Shales of South Africa. Dawson (1967) divided the Bluff Cove Beds in their type area into a sandstone facies, comprising most of the sequence, with an interfingering and overlying shale facies up to 45 metres thick. The two facies differ mainly in the relative proportions of sandstone and shale. To date this division has been recognised only to the north and north-west of Bluff Cove, but this may owe much to the lack of exposure elsewhere.

The present survey has shown that a unit contrasting with both the underlying Port Stanley Formation and the overlying Fitzroy Tillite Formation can be mapped consistently using photogeological interpretation and field observation. Therefore it is appropriate to treat it as a formation, here named the Bluff Cove Formation. In the Bluff Cove area, Dawson (1967) found that the Bluff Cove Formation rests disconformably on the Port Stanley Formation, and this is supported by photogeological evidence throughout the outcrop.

The Bluff Cove Formation typically crops out in gently sloping ground at the base of steep slopes formed by the Port Stanley Formation, and so is generally covered by Quaternary mass-movement deposits. Nevertheless, it can be separated from the Port Stanley Formation by a persistent negative break of slope. Although there is commonly a parallel negative break of slope within the adjacent Port Stanley Formation outcrop, exposures within the Bluff Cove Formation are rare and never as craggy as those of the Port Stanley Formation. The sections which do expose the Bluff Cove Formation are noted below. The presence of intermontane synclinal outliers east of Colarado Pond [UC 98 68] is inferred from the appearance of the low ground lying between tight anticlines in the higher part of the Port Stanley Formation. In some instances these broad valleys are bounded by weak topographic features similar to those found to mark the base of Bluff Cove Formation elsewhere. The paucity of stone runs in these areas is taken as supporting evidence. The outliers are interpreted as the lateral continuation of the Bluff Cove synclinal outcrop to the east, but as they are likely to be fault-bounded, no coincident fold axis is shown on the map. Other possible outliers of the Bluff Cove

Formation occur in the low-lying peat-covered ground west of Vantan Arroyo [VC 10 67], and in the area of White Point and of The Lagoon, south-east of Bluff Cove [VC 21 64].

The Bluff Cove Formation is mainly composed of fine to very fine-grained sandstones and wackes with interbedded thin laminated muddy siltstones and silty mudstones and shales. Some medium and coarse sandstone occurs locally. Fine-grained white mica is common in the sandstones, which vary in colour from greenish-brown to brown, black and pale grey. The mudstones and shales are mostly pale brown to dark grey, but interbeds up to two centimetres thick of black laminated shale occur sporadically (Dawson, 1967).

Petrographic examination of two specimens of laminated fine-grained sandstone found that they are composed largely of subangular to subrounded monocrystalline quartz, with subordinate feldspar and muscovite mica grains in a ferruginous, micaceous clay matrix. Both potassic and sodic/calcic (plagioclase) feldspars are present, in contrast to the Devonian sandstones which normally contain only potassic feldspar. The plagioclase feldspar is likely to have been derived from a volcanic terrane. In one specimen, pyrite is common as streaks and along laminae (Lott, 1997). Heavy mineral analysis of one sandstone found an assemblage which is more similar to that of fine-grained Devonian marine sandstones than of the sandstones in the rest of the Lafonia Group (Knox, 1997).

The sandstone generally occurs in thin to medium plane beds but small-scale cross bedding, ripple cross-lamination, flaser bedding, parallel lamination and sole marks also occur. Soft-sediment faulting occurs locally and there are rare water escape structures. Where exposed near Bluff Cove, the base of the formation is locally characterised by poorly bedded, slumped and contorted medium-grained sandstone containing stratified intercalations of shale up to 20 centimetres long, and intraformational breccias (Dawson, 1967). Adie (1952a; 1952b) reports that 'lamination becomes very pronounced towards the top [of the Bluff Cove Beds], and elongated boulders up to 4-5 inches [about 10 to 12 cm] in length are frequently observed with their long axes perpendicular to the bedding planes of the laminated shales', interpreting the boulders as ice-rafted debris. However, the presence of glacial debris in the Bluff Cove Formation has not been confirmed by subsequent observations. Bedded glacigene deposits at the top of the Bluff Cove Formation would now probably be included with the overlying formation. Dawson (1967) found that palaeocurrents in the Bluff Cove area were directed towards the ENE on average.

No body fossils have been found in the Bluff Cove Formation. Rare small indeterminate trace fossils and bioturbation both occur in the Bluff Cove area. Plant debris occurs in muddy carbonaceous siltstones and fine sandstones in the Inner Verde outcrop, in the west of East Falkland.

Adie (1952a; 1952b) and Dawson (1967) interpreted the Bluff Cove Formation as glaciomarine or glaciolacustrine deposits. However, other than the dropstones reported by Adie (1952b) no direct evidence for glacial influence is seen. The soft-sediment deformation could conceivably be the result of ice movement or melting but slumping on a depositional slope is more plausible. It is here considered as likely that deposition took place in a shallow marine or deltaic environment without glacial influence.

Dawson (1967) estimated the thickness of the Bluff Cove Formation as ranging from 250 metres west of Fitzroy Bridge [VC 15 64] to 230 metres east of Bluff Cove, thinning north of Fitzroy Bridge to about 150 metres. These estimates for the eastern part of the outcrop are broadly confirmed by the present survey. The Bluff Cove Formation maintains a thickness of about 200 metres elsewhere in East Falkland, except from Port Sussex westwards where estimates range from 70 to 110 metres.

The type area is around Bluff Cove [VC 19 66], where the formation is extensively exposed on the shorelines. Reference sections are exposed on the west coast of Fitz Cove [VC 1819 6523] to [VC 1834 6498]; the west coast of Port Fitzroy [VC 1604 6345] to [VC 1580 6377]; a stream section on the east side of the main road, north side of Fitzroy Ridge [VC 1035 6415]; a stream section in a tributary of L'Antioja Stream [UC 91 60]; a stream section near The Waterfall, Black Rock Arroyo [UC 81 62]; coastal exposures on the south side of Wreck Point [UC 53 80]; a stream section in the Inner Verde near [UC 620 880]; and on the west side of Curlew Creek [UC 6575 8800].

A sample of mudstone from the Bluff Cove Formation yielded scarce, indeterminate spores and bisaccate pollen. The occurrence of the latter confirms a Late Carboniferous or younger age (Warrington, 1996). The Bluff Cove Formation has been suggested to be equivalent to the 'Lower Dwyka Shales' of South Africa (Adie, 1952a; 1952b) (Dawson, 1967), now assigned to the Kommadagga Subgroup of the Witteberg Group (Visser et al., 1990). However, the Kommadagga Subgroup is thought to be Early Carboniferous in age (Broquet, 1992). The presence of bisaccate pollen in the Bluff Cove Formation implies that there is no exact equivalent of the Bluff Cove Formation in South Africa.

2.4.2 Fitzroy Tillite Formation

The Fitzroy Tillite Formation is mainly composed of massive sandy diamictite[1], with minor intercalations of laminated mudstones and small sandstone bodies. It forms a broad outcrop just south of the Wickham Heights but also occurs in synclinal outliers in the Inner Verde and between Port Harriet and the Fitzroy River. The presence of garnet sand and of sporadic cobbles of tillite on beaches between Surf Bay and Cape Pembroke suggests that another outlier occurs offshore in that area. In West Falkland the Fitzroy Tillite Formation occurs along the east coast from Poke Point to Carcass Bay, at Port Purvis and at the eastern end of Byron Sound.

The tillite was probably first noticed by Sulivan (footnote in Darwin, 1846, p. 269) but was first described in detail by Halle (1912), who noted its resemblance to Permian 'glacial boulder-beds' in other parts of Gondwana. Baker (1924) named the unit the 'Lafonian Tillite', following Halle (1912). Frakes and Crowell (1967) preferred the term 'Lafonian Diamictite', which carries no genetic connotation. In due course this became the 'Formación Lafonian' (Scasso and Mendia, 1985; Jalfin and Bellosi, 1983), and the Lafonian Diamictite Formation (Marshall, 1994b).

However, none of these are really suitable as lithostratigraphic names for this unit: 'Lafonia'/'Lafonian' is pre-occupied by the Lafonia Group, and the tillite occurs nowhere in Lafonia. Therefore a new name is proposed here. There is no single section which exposes the whole tillite sequence and so no clearly favoured candidate for the type section. Exposures of the tillite on West Falkland are not good, in general. Although the base and top are both well exposed near Hill Cove, little of the rest of the formation is seen, and the full thickness cannot be estimated there. Although the tillite is little deformed in the west of East Falkland, superficial cover and faulting eliminate the potential for a basal stratotype in that area. Although the tillite occurring around Fitzroy and Port Fitzroy is strongly cleaved, it is relatively well-exposed in what is here regarded as the most useful as a type area. As the dominant rock-type is generally accepted as a tillite (and as this term is widely recognised in the Islands) it seems preferable to use this rather than follow Frakes and Crowell (1967) in using the non-genetic 'diamictite'. Therefore this distinctive unit is here called the Fitzroy Tillite Formation.

In East Falkland, the Fitzroy Tillite Formation overlies the Bluff Cove Formation probably either unconformably or disconformably, while in the west it overlies the Port Stanley Formation unconformably. It is everywhere overlain conformably by the Port Sussex Formation. One member (the Quark Pond Member) is mappable on the Falkland Sound coast of West Falkland (Section 2.4.2.1).

The Fitzroy Tillite Formation has a fairly characteristic appearance on aerial photographs, giving rise to low-lying topography with an uneven or hummocky surface. In the eastern outcrops a linear fabric is visible on aerial photographs, lying parallel to cleavage. The topmost level of the diamictite around Port Pleasant is particularly susceptible to weathering. This has allowed a series of coves to form between headlands composed of the Black Rock Member. On aerial photographs the base of the Fitzroy Tillite Formation is taken at a weak negative break of slope bounding the sloping ground formed by the Bluff Cove Formation or, where the Bluff Cove Formation is absent, a rather stronger break of slope bounding the Port Stanley Formation. These features are locally obscured by superficial deposits.

Good exposures of the base of the tillite are rare. At Hill Cove and Port Purvis, an angular discordance of less than 20° was observed at the slightly irregular basal contact and individual beds of the underlying Port Stanley Formation have been cut out by the tillite (Frakes and Crowell, 1967). This relationship probably also occurs at Port Howard (Section 2.3.4).

The base of the tillite is well exposed on the southern shore of Byron Sound west of Hill Cove [TC 7510 9016], where it rests on a slightly irregular surface of trough cross-bedded quartzites of the Port Stanley Formation. Where the contact is best exposed it can be seen that the diamictite passes down over a few millimetres into a poorly sorted pebbly sandstone, locally with interlaminated muddy sand. The pebbles vary from well rounded quartz pebbles up to two centimetres in diameter to subangular fragments up to five centimetres across, including pieces of granite and quartzite. No pebbles

[1] For most purposes in the context of Falkland Islands geology, the terms 'tillite' and 'diamictite' are virtually synonymous. Strictly speaking, however, 'tillite' refers to a lithified mixture of mud, silt, sand and pebbles deposited by glacier ice. Many tillites are diamictites but some are not. 'Diamictite' is a lithified mixture of unsorted mud, silt, sand and pebbles of unspecified origin. The same mixture in unlithified form is a 'diamicton'. Many diamictons were deposited by glaciers, but some were not. For example, many of the solifluction deposits in the Falkland Islands are diamictons (Section 5.2.1).

were seen in the underlying Stanley Quartzite. The pebbly sandstone probably forms a lensoid body no more than one metre in thickness. Subcylindrical to irregularly-shaped domains of pyritic cement, ranging from a few millimetres to a few centimetres across, occur locally in the lower part of the sandstone, and apparently also in the immediately underlying quartzite. It is usual for tills to be underlain by glaciofluvial sands and gravels. The presence of pyrite suggests anaerobic conditions and perhaps the incorporation of a proportion of organic debris, which is less usual.

Baker (1924) and Adie (1952a) report that glacial striae occur on the surface of the Port Stanley Formation beneath the tillite in Port Purvis and west of Hill Cove, but Frakes and Crowell (1967) repudiate their findings. No evidence of glacially striated surfaces were seen at the base of the tillite during the present survey. Although north to north-easterly plunging lineaments do occur on the exposed surface of the Stanley Quartzite west of Hill Cove, they are lined with smears of vein quartz, and subparallel lineaments also occur on joints well within the subjacent quartzite. These lineaments have originated by tectonic processes, probably by bedding slip during folding.

The base of the tillite is also seen in overturned strata on the east side of the Coast Ridge at Carcass Bay [UC 01 37]. Massive tillite is separated from the stratigraphically underlying Port Stanley Formation by an interval of about five metres of bedded sandstones and mudstones. It seems possible that these bedded sediments are a lateral equivalent of the Bluff Cove Formation but more likely that they are a basal facies of the Fitzroy Tillite Formation.

This unit is mostly composed of massive to very weakly thick-bedded sandy diamictite (Plate 17). Where fresh it is typically dark bluish or greenish grey in colour but it weathers to pale grey and beige. Frakes and Crowell (1967) found that between 40 and 85 per cent of the matrix is composed of sand and silt, which is mainly quartz with some feldspar. Potassic feldspar and sodic/calcic feldspar are present in almost equal proportions, suggesting that debris from both granitic and volcanic sources is present. In the east of East Falkland the matrix is commonly cleaved.

Matrix-supported clasts comprise up to 50 per cent of the tillite (Frakes and Crowell, 1967), with the proportion varying both laterally and vertically. Some clasts are well-rounded and subspherical but most are subangular with a wide range of sphericity. Some clasts are striated (Halle, 1912; Baker, 1924), but these are rare (Frakes and Crowell, 1967). Most are randomly oriented but where they lie on bedding planes they are aligned according to transport direction. They occur in all sizes up to boulders of several metres diameter, although those larger than 50 centimetres diameter are localised. The maximum clast size and the average clast size both decrease overall from west to east (Frakes and Crowell, 1967). The tillite in the Hill Cove area is notable for large numbers of erratic boulders. Although these are generally less than one metre in diameter, Frakes and Crowell (1967) illustrate an angular block of granite on Foot Point some seven metres across.

A wide range of rock-types has been noted forming clasts within the tillite. The most common types are composed of quartzite, sandstone, granite (with or without a tectonic fabric) or granitoid gneiss, vein quartz and shale (Dawson, 1967; Frakes and Crowell, 1967). Dolerite, black slate, garnet gneiss, quartz porphyry, limestone, chert, and deformed clay intraclasts also occur (Halle, 1912; Baker, 1924; Frakes and Crowell, 1967). Matrix-supported oligomict conglomerate (with pebbles up to five centimetres across of grey and pink quartzite and micaceous granite), finely banded phyllite and banded jasper have also been collected from the tillite around Hill Cove (Mrs C McKay, personal communication, 1997).

While much of the debris in the tillite has probably been derived by erosion of the West Falkland Group, the large proportion of granitoids and metamorphic rocks shows that the catchment area of the glaciers included extensive exposures of crystalline basement similar to the Cape Meredith Complex. However, some of the erratics, particularly the garnet gneiss, show significant differences to the rock-types exposed at Cape Meredith and might have been transported from a relatively distant metamorphic terrain (Section 2.2).

The tillite generally appears massive, with no clear stratification, but it does include isolated or clustered intercalations of well-bedded rocks. There are also indistinct partings and joints which appear to lie subparallel to bedding, and bedding-parallel lineaments (some very faint) can be seen on aerial photographs in some parts of the outcrop. The interbedded sediments include linear channel-fills and small fans of sandstone and minor conglomerate around Hill Cove, and thin interbeds of pebbly mudstone, diamictite, wacke, and shale with isolated pebbles (including some pebbles of lithified diamictite) further east. The muddy intercalations can occur as single thin beds, but the bedded intervals are commonly in the range three to 10 metres thick, reaching 30 metres near Port Fitzroy. Rare ripple-marks and primary current lineation indicate easterly transport directions. Sand-filled burrows less than one centimetre long can be found in the sandy mudstone beds: these are the only fossils found in the Fitzroy Tillite Formation (Dawson, 1967;

Frakes and Crowell, 1967). The only interval of well-bedded sediment observed within the tillite which forms a mappable unit is the Quark Pond Member (Section 2.4.2.1).

Linear sandstone bodies enclosed within the tillite occur at Hill Cove and at Port Purvis, with just one example known on East Falkland, one kilometre south-west of Fitzroy Bridge. Some are more than one kilometre long, up to 10 metres wide and five metres thick (Frakes and Crowell, 1967). They are mostly composed of well-sorted fine and very fine-grained quartz sandstone, with some siltstone, medium or coarse quartzose or feldspathic sandstone and clast-supported conglomerate. The sandstones are mostly laminated with small-scale trough cross-beds, between one and five centimetres in thickness, with some plane lamination and ripple cross-lamination in one metre thick cross-sets. Some form lateral accretion structures. Palaeocurrent directions were towards the east or north-east (Frakes and Crowell, 1967). However, bedding attitude within the sandstone bodies can be irregular. This is probably a result of syn-depositional disruption within the glacial environment (including melt-induced collapse) as suggested by the local occurrence of small-scale pre-lithification faults in the laminated sandstones, but in some cases it might simply reflect collapse during weathering of the exposure. Frakes and Crowell (1967, p. 52) state that the sand bodies around Port Purvis are particularly fragmented and contorted, attributing this to penecontemporaneous submarine landsliding.

In two places near Hill Cove, the linear sandstone bodies are closely associated with sandstone sheets up to five metres thick. They are lithologically similar to the linear bodies but contain no conglomerate. Palaeocurrents spread fan-like away from the linear sandstones (Frakes and Crowell, 1967).

In their palaeoenvironmental interpretation of the tillite, Frakes and Crowell (1967) distinguished three facies which they related to a grounded ice-sheet in the west and marine conditions with floating ice in the east. Their western facies around Hill Cove is characterised by the linear sandbodies (interpreted as subglacial or englacial eskers), and sand sheets (interpreted as subglacial deltas formed at a ice-sheet grounding line). Their eastern facies is distinguished by the well-bedded intercalations, which were taken to indicate glaciomarine deposition. Submarine mudflows were thought to have deposited the bedded diamictites, and ice-rafted debris to appear as diamictite pebbles and dropstones in shale. This facies extends as far west as the Coast Ridge of West Falkland. An intermediate facies with disrupted and contorted sand bodies found at Port Purvis and east of Sound Pass was interpreted as the result of remobilization of the western facies of the tillite on a submarine slope. Frakes and Crowell (1967) deduced that the overall transport direction was from west to east or south-west to north-east, as indicated by small-scale sedimentary structures, the orientation of the linear sand bodies, the eastwards decrease in mean and maximum clast size, and clast fabrics. They also believed that the tillite decreased in thickness towards the east, but as noted below this is now in doubt, and would not support their model anyway. Clast orientation in the tillite near Black Rock House is mainly WNW-ESE (Bellosi and Jalfin, 1984b).

Frakes and Crowell's interpretation is consistent with the likelihood that Precambrian crystalline basement was exposed within the glacial catchment to the west or south. It also fits with Visser's (1987) palaeogeographic reconstruction in which the Falklands once lay on the eastern margin of a large interglacial lake (rather than at a marine coastline). When the Falklands are restored to their most probable position within Gondwana (Section 6), it is also consistent with the ice-flow directions inferred by Visser (1987). Volcanic detritus within the tillite may have been derived from a magmatic arc associated with subduction on the Pacific margin of Gondwana (Cole, 1992).

While the main elements of Frakes and Crowell's model seem to remain valid, some aspects are questionable. It seems reasonable to interpret the Hill Cove sequence as massive ice-contact till with glaciofluvial channel deposits. However, the disruption of the sand bodies at Port Purvis is apparently not accompanied by bedded diamictites or wackes and so might be the result of deformation by ice rather than submarine mass-movement. In the east, most of the tillite is massive with randomly aligned clasts and so is more likely to be lodgement till (deposited directly from the sole of an ice-sheet) rather than waterlain till (deposited in water from melting ice). Waterlain till can form massive beds, but any water movement will cause reworking and sorting and so some stratification can be expected. Also, in waterlain tills, the clasts tend to have their long axes arranged vertically rather than randomly. It therefore seems that the eastern facies of the tillite is likely to have been deposited through a combination of grounded and floating ice. Visser (1997) interprets the alternations of massive tillite and bedded sediments found in the Dwyka Group of South Africa as 'deglaciation sequences' deposited during the recessional phases of a marine ice margin. It seems that a detailed reassessment of the Fitzroy Tillite in the light of the present understanding of glacial processes would be worthwhile.

Frakes and Crowell (1967) estimated the thickness of the tillite as 850 metres at Hill Cove, decreasing to 450 metres at Port Purvis and 500 metres at Port Fitzroy. Following the recent survey, the Fitzroy Tillite Formation is estimated as about 900 metres near Fitzroy, decreasing to 700 - 750 metres near Port Sussex and 610 metres on the east coast

Plate 17: Fitzroy Tillite near Sound Pass

Photograph by Don Aldiss, BGS. MN28244

Plate 18: Bedded diamictites and mudstones, Quark Pond Member

Near Quark Pond, West Falkland. Photograph by Don Aldiss, BGS. MN28245

of West Falkland. Bellosi (1982, quoted by Scasso and Mendia, 1985) made a similar estimate of 650 metres in the Black Rock area, although Scasso and Mendia (1985) found only 160 metres at Shag Harbour. At Hill Cove, the base of the Fitzroy Tillite Formation dips at 15° to 20°: if that continues north uniformly, then the tillite sequence could be as thick as 850 metres in that area. However, it is quite possible that the dip decreases towards Byron Sound and also that the sequence is partly repeated by faulting. This estimate should therefore be treated as a maximum. Thus there is no clear evidence that the thickness of the tillite increases westwards on West Falkland and the tillite otherwise seems to follow the general trend of decreasing thickness from east to west.

The type area of the Fitzroy Tillite Formation is around Fitzroy settlement [VC 15 61] and the west end of Port Fitzroy [VC 15 64]. Exposures of both the base and the top of the formation occur in this area although not in the same sections. There is no evidence that the sequence between Fitzroy Creek and Shallow Cove has been repeated by folding, although some faulting may occur. Reference sections can be found at Comoda Ditch [UC 96 56]; L'Antioja Stream [UC 92 58]; Port Sussex/Hells Kitchen [UC 62 75]; Hill Gap [UC 11 56]; Carcass Bay [UC 01 37]; and around Hill Cove [TC 75 90].

The Fitzroy Tillite Formation is a close correlative of the Dwyka Group of South Africa which is thought to date from the Late Carboniferous (late Westphalian or Stephanian) to the Early Permian (Artinskian) (Visser, 1987; Cole, 1992). Visser (1990) noted that the base of the Dwyka Group is diachronous, and that in the east of the Karoo Basin (i.e. closest to the inferred past position of the Falkland Islands; Section 6), deposition commenced in the late Stephanian. Visser (1997) identified a series of 'deglaciation sequences' in the Dwyka Group, proposing them as a framework for interbasinal correlation of the glacigene deposits in southern Africa. The Fitzroy Tillite Formation can be tentatively identified with his deglaciation sequences 3 and 4, with sequence 2 possibly represented by the bedded sediments seen at the base of the formation at Carcass Bay. If this correlation is correct, those basal beds would have been deposited during the late Stephanian marine transgression, the Quark Pond Member (Section 2.4.2.1) during the early Sakmarian (Early Permian) transgression and deposition of the rest of the Fitzroy Tillite Formation would have continued into the Artinskian.

2.4.2.1 Quark Pond Member

The Quark Pond Member is composed of diamictites and mudstones, some of which include dropstones. It has been mapped only on the east flank of the Coast Ridge, where it can be traced from the Narrows at Port Howard to between Hill Gap and Carcass Bay. It also occurs in coastal exposures in Bold Cove. It might occur in the Black Rock area of East Falkland but has not been identified with certainty elsewhere in the Falkland Islands.

The Quark Pond Member is a newly named unit within the top half of the Fitzroy Tillite Formation. The name was first proposed by N Meadows (Geochem Group Ltd, 1996). It is taken from Quark Pond, a feature close to the type section, which is about nine kilometres south-south-west of Port Howard. The unit was observed by Frakes and Crowell (1967, p. 51) but not named by them. It was described as 'facies B' of the 'Formación Lafonian' by Scasso and Mendia (1985).

The unit coincides with a distinct negative topographic feature marking a change in the angle of slope of the eastern flank of the Coast Ridge. Although it is thin and is only exposed at the coastline, it is thus visible on aerial photographs and also to some extent on satellite images.

The Quark Pond Member is made up of thin to medium bedded and laminated diamictites, fine sandstones and mudstones, with numerous dropstones (Plate 18). The dropstones are mostly less than five centimetres in diameter but include at least one subangular block of granitic gneiss about one metre across. Mudstone intraclasts occur in the upper part of the unit but it is capped by about two metres of mudstone in which dropstones are rare. The lowermost layers are laterally discontinuous. They rest abruptly on the underlying massive tillite, which is conspicuously rich in large erratics up to one metre across. In one place the basal layers drape over a 40 centimetre-wide erratic which sticks out of the tillite. Frakes and Crowell (1967, pl. 4, fig. 3) illustrate soft-sediment deformation in the Quark Pond Member in its type area. The overlying tillite is also massive but is relatively poor in erratics. It rests abruptly on the Quark Pond Member, truncating some of the underlying beds.

Although some individual beds in the Quark Pond Member near Quark Pond are grey, most of the unit is reddish-brown or reddish-purple in colour, as is the topmost one to 1.5 metres of the underlying tillite. The overlying tillite is grey in colour. At Shag Harbour, the reddening is confined to the topmost three metres of the unit (Scasso and Mendia, 1985).

Bellosi and Jalfin (1984b) describe an interval (their 'facies pelitica') within the upper third of the tillite near Black Rock House, East Falkland, in which thick or very thick beds of diamictite are intercalated with laminated mudstones containing dropstones. The lower contact of each intercalation of laminated mudstones is sharp and planar, but the upper contact is gradational. This is the only distinct interval of well-bedded rocks described in that area by Bellosi and Jalfin (1984b) and its position within the tillite suggests a correlation with the Quark Pond Member.

As with the other intervals of bedded mudstone and diamictite in the Fitzroy Tillite Formation (Section 2.4.2) the Quark Pond Member can be attributed to glaciomarine or glaciolacustrine deposition. The reddish colouration indicates a period of weathering during subaerial exposure, presumably in an interval of falling sea-level or glacial lake drainage before the encroachment of the glacier which deposited the overlying tillite.

Near Quark Pond, the Quark Pond Member is between five and six metres in thickness. Scasso and Mendia (1985) measured it as seven metres at Shag Harbour. The bedded unit near Black Rock House is 15 metres thick (Bellosi and Jalfin, 1984b).

The type area is the low cliff and foreshore on the western side of the cove immediately south-west of Quark Pond [UC 2150 6985]. Reference sections can be found at Hill Gap [UC 1185 5683], Bold Cove [UC 3110 8190] and Shag Harbour [UC 18 63].

The Quark Pond Member can be tentatively correlated with bedded mudstones which mark the top of Visser's (1997) third 'deglaciation sequence' in the Dwyka Group of South Africa (Section 2.4.2). If this correlation is correct, the member was deposited during a glacial regression in Early Sakmarian (Early Permian) times. A short interval of bedded sediments very similar to the Quark Pond Member is described from Swart Umfolozi in northern Natal (South Africa) (Savage, 1971).

2.4.3 Port Sussex Formation

The Port Sussex Formation is dominated by dark grey or black mudstones but has minor diamictites and wackes at the base and some fine sandstones and siltstones in its upper part. In East Falkland it crops out in a narrow belt between Pleasant Roads and Port Sussex, with a synclinal outlier intermittently exposed around Port Pleasant. In West Falkland it crops out in the low ground east of the Coast Ridge, occurring discontinuously from Carcass Bay to Poke Point. It also occurs at the east end of Byron Sound near West Lagoons and in Skip Rock. It is reported to overlie the tillite in Port Purvis (Frakes and Crowell, 1967, fig. 8), although this was not confirmed during the present survey.

As originally defined by Frakes and Crowell (1967, p. 42), the Port Sussex Formation comprised two members: the Black Rock Member and the Shepherds Brook Member. A thin basal facies recognised by Frakes and Crowell (1967) in their Black Rock Member is here named the Hells Kitchen Member. The type section for the formation was originally defined as the 'north shore of Port Sussex from Hells Kitchen to the head of the bay'. This definition should be superseded by definitions of the type sections for the three component members, which together occupy the whole of the formation. In most areas, the Port Sussex Formation overlies the Fitzroy Tillite Formation conformably and is overlain conformably by the Brenton Loch Formation. Curtis and Hyam (1998) observed a localised onlapping unconformity between the Port Sussex Formation and the Fitzroy Tillite Formation on the northern side of Port Pleasant [VC 133 603]. In common with the other formations, the Port Sussex Formation diminishes in thickness westwards, being about 55 per cent thinner on the east coast of West Falkland than on East Falkland.

As described in the following sections, mineralogical evidence shows that some volcanic detritus was incorporated in each member in the Port Sussex Formation, as in the Fitzroy Tillite Formation. Palaeogeographic reconstructions and palaeoenvironmental evidence for the Fitzroy Tillite Formation (Section 2.4.2) and the Brenton Loch Formation (Section 2.4.4) suggest lacustrine deposition, so it is likely that these conditions prevailed during Port Sussex Formation times as well.

By correlation with its South African equivalents in the Ecca Group (the Prince Albert Formation, the Whitehill Formation and the Collingham Formation), the deposition of the Port Sussex Formation can be placed late in the Early Permian to early in the Late Permian (Sections 2.4.3.1, 2.4.3.2 and 2.4.3.3).

2.4.3.1 Hells Kitchen Member

The Hells Kitchen Member is a thin sequence of interbedded mudstones, wackes and diamictites which forms the base of the Port Sussex Formation on both East and West Falkland. The outcrop is everywhere too narrow to map separately from the Black Rock Member.

Frakes and Crowell (1967) observed that the contact between the tillite and the overlying Port Sussex Formation is sharp but that the 'lower metre or so of the latter unit contains abundant diamictite layers which average about two centimetres in thickness and which are interbedded with dark shale'. The occurrence of erratics in the 'basal part of the stratified series' [of the Lafonia Group] was also noted by Halle (1912, p. 157). This basal facies has been treated as a separate member by recent workers (Storey et al., 1995; Macdonald et al., 1996) and during the present survey it was found to be consistently distinguishable from the Black Rock Member, which overlies it with an abrupt but conformable contact. Hells Kitchen is a narrow steep-sided river valley to the east of Port Sussex House. Before the construction of the all-weather road to San Carlos, the main motor track crossed this valley near the river mouth, a test of drivers' skill and of passengers' nerve. Exposures on the south side of the inlet west of that track clearly expose both the base and the top of the unit, as well as most of the beds within it. In this locality the Port Sussex Formation is not cleaved and is only mildly disrupted by faulting, whereas at Fitzroy it is strongly cleaved. This section has the further virtue that it can be considered as part of the original 'type section' of the Port Sussex Formation (Section 2.4.3). The Hells Kitchen Member rests on massive tillite of the Fitzroy Tillite Formation conformably or disconformably and is overlain conformably by the Black Rock Member of the Port Sussex Formation.

The base of the Hells Kitchen Member is very clear on aerial photographs. It lies at the stratigraphically lower limit of the bedding traces usually visible on the outcrop of the Lafonia Group above the tillite, typically at a weak positive break of slope part-way up the flank of the ridge formed by the Black Rock Member. The tillite outcrop on the lower part of the same slope is typically eroded by closely-spaced, shallow gullies perpendicular to strike. On the east coast of West Falkland, the delicate bedding traces marking the Port Sussex Formation are seen only on foreshore exposures.

The Hells Kitchen Member is made up entirely of laminated wacke or thinly bedded diamictite alternating with thin beds of laminated mudstone. These rest at a sharp contact on the underlying tillite (Plate 19). All lithologies contain sporadic dropstones, most being granule-sized but with some up to 12 centimetres in diameter. Mud or silt intraclasts up to three millimetres across are locally abundant. Some individual beds up to four centimetres thick are graded from diamictite or wacke to mudstone. Slump structures, sole marks and rip-up clasts have also been recorded. No fossils are known. XRD analysis of a mudstone from the Hells Kitchen Member near Hill Cove found K-bentonite, indicating that the unit includes weathered volcanic detritus (Kemp et al., 1998).

These strata are essentially similar to many of the interbedded sediments in the Fitzroy Tillite Formation, including the Quark Pond Member. They are likewise interpreted as glaciolacustrine or glaciomarine deposits, with sporadic subaqueous slumping. The small mud and silt intraclasts are interpreted as fragments of frozen unlithified sediment.

In most sections the Hells Kitchen Member is five to seven metres thick. It is possibly somewhat thicker at Shell Point [VC 1479 6045], where it is repeated by two NW-SE normal faults, but does not exceed 10 metres.

The type section lies on the north side of the narrow peninsula which bounds the southern side of the short tidal inlet separating the river of Hells Kitchen from Port Sussex [UC 6155 7506]. Reference sections are found at Memorial Cove[2], Fitzroy [VC 1655 6080]; Shell Point [VC 1479 6045]; the north side of Black Rincon [VC 1325 6032]; a cutting on Mare Harbour Road, MPA [VC 001 561]; the coast near the shearing shed at Port Sussex House [UC 6100 7522]; Hill Gap [UC 1220 5696]; and on the West Lagoons coast, near Hill Cove [TC 8642 9532].

The latest glacigene deposition in South Africa occurred in the Artinskian-Kungurian (latest Early Permian). In the eastern Karoo Basin, the base of the Prince Albert Formation was probably deposited during the Artinskian (Visser, 1989; Visser, 1990; Visser, 1997) and this seems to be the most probable age for the Hells Kitchen Member.

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[2] 'Memorial Cove' is the name given by M L Curtis of BAS to a small cove on the north side of Port Pleasant, south-east of Fitzroy. It is the site of a group of monuments to the memory of lives lost in the vicinity during 1982. The name does not appear on existing topographic maps, but seems highly appropriate.

2.4.3.2 Black Rock Member

The Black Rock Member comprises black and dark grey mudstones, some of which are markedly carbonaceous, with minor siltstones and cherts. It is the most commonly exposed part of the Port Sussex Formation. In West Falkland it is not mapped separately from the Shepherds Brook Member.

Baker (1924) identified the black to greyish-black cherty slates forming the persistent ridge at the southern limit of the tillite outcrop as the 'Black Rock Slates'. These are named after Black Rock House, which stands on the ridge near Swan Inlet, and which was evidently named after the rock of which the ridge is composed. Frakes and Crowell (1967) renamed this unit the Black Rock Member. It conformably overlies the Hells Kitchen Member and conformably underlies the Shepherds Brook Member. It is distinguished from the former by the absence of sand-sized and coarser material.

In most places, the Black Rock Member forms a low but prominent ridge (part of which near MPA is named March Ridge), which appears dark-toned on the aerial photographs, usually with fine bedding traces visible. The basal layers, just above the Hells Kitchen Member, are typically marked by a grassy strip (appearing pale on aerial photographs).

The Black Rock Member is composed of black or grey, thinly bedded, laminated or massive mudstones with some siltstones, together grouped in medium to thick beds (Plate 20). The mudstones are typically carbonaceous, some are pyritic, and some cherty. Medium to thinly bedded and laminated cherts and cherty mudstones are more abundant in West Falkland. In eastern East Falkland the Black Rock Member is folded and cleaved, and the cleavage can obscure the bedding. Mostly, bedding is planar, but cross-lamination occurs locally. Small-scale bioturbation lying parallel to bedding, typically of *Planolites* type, occurs near the top of the unit.

XRD analysis of mudstones from the Black Rock Member near Hill Gap and Carcass Bay found K-bentonite, indicating that the unit includes weathered volcanic detritus (Kemp et al., 1998).

The Black Rock Member can contain up to 41.3 per cent total organic carbon, which seems to be most abundant in the Port Sussex area (Marshall, 1994a; Macdonald et al., 1996). This local concentration of carbon might reflect an increase in water depth due to growth faulting on the Falkland Sound Fault Zone, or beneath the Hornby Mountains Anticline (Sections 4.2.2.1 and 4.2.3.2) (Macdonald et al., 1996). This would represent the opposite tendency to that seen in the Brenton Loch Formation and Bay of Harbours Formation, in which the more proximal facies seem to occur between the Falkland Sound Fault Zone and the Goose Green Axis (Figure 4.1). At some localities (most notoriously at the head of Port Sussex), tectonic shear surfaces in the carbonaceous mudstones are shiny and the rock has then been mistaken for coal. However, no part of the Black Rock Member contains enough carbon for the rock to be combustible.

The Black Rock Member was deposited under post-glacial anoxic conditions in a very low energy marine or lacustrine environment, as postulated for the Whitehill Formation by Visser (1992a; Visser, 1992b). Its thickness is estimated at between 150 and 250 metres in East Falkland. The whole Port Sussex Formation is between 100 and 170 metres thick in West Falkland.

The type area is Port Sussex, particularly from the mouth of the Hells Kitchen inlet [UC 6155 7505] eastwards for about 1.5 kilometres on the north coast. Reference sections are found in the cutting on the Mare Harbour road, approx. 200 metres south of the junction with the Stanley road, MPA [VC 001 561]; streams at the head of Swan Inlet [UC 90 57] and [92 57]; Hill Gap [UC 1220 5696]; Carcass Bay [UC 0350 3886]; and on the West Lagoons coast, near Hill Cove [TC 8642 9532].

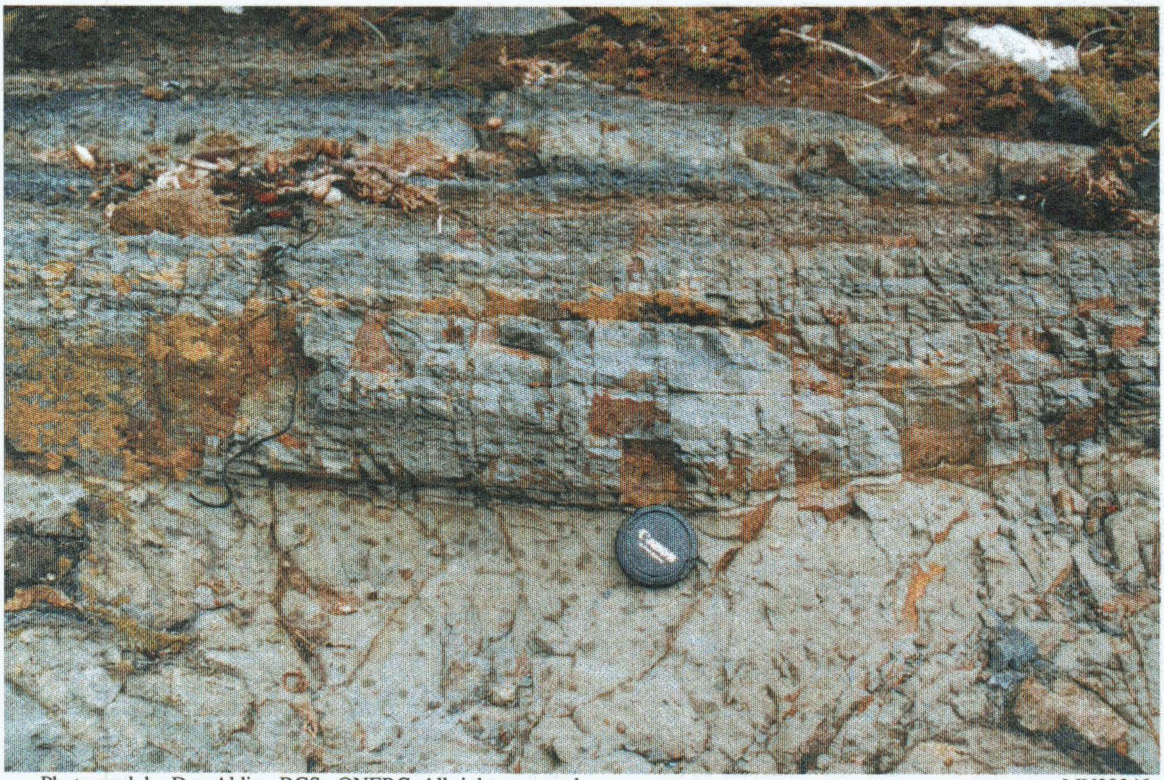
The Black Rock Member can be correlated with the Prince Albert Formation and Whitehill Formation of South Africa, which are thought to be Early Permian and Early or Late Permian in age (Visser, 1990; Visser, 1992b; Johnson et al., 1996).

2.4.3.3 Shepherds Brook Member

The Shepherds Brook Member consists of mudstones and siltstones which are very susceptible to weathering. It probably occurs throughout the Port Sussex Formation outcrop, although possibly absent east of MPA and near Hill Cove. It is not differentiated from the Black Rock Member in West Falkland.

Frakes and Crowell (1967) coined the name 'Shepherds Brook Member' for the sequence of grey mudstones and sandstones which lies between the typical Black Rock mudstones and the 'Lafonian Sandstone'. Baker (1924) had included these strata with his Lafonian Sandstone as 'passage beds' from the Black Rock Slates. The name is taken from

Plate 19: Base of Hells Kitchen Member, Port Sussex



Photograph by Don Aldiss, BGS. MN28246

Plate 20: Black Rock Member, L'Antioja Stream

Photograph by Don Aldiss, BGS. MN28247

the stream which enters the head of Port Sussex at the type section. The Shepherds Brook Member lies conformably between the Black Rock Member and the Terra Motas Member of the Brenton Loch Formation.

The Shepherds Brook Member everywhere underlies low-lying ground between the ridges formed by the Black Rock Member and by the Terra Motas Member. Much of the outcrop is covered by water or by alluvium and exposures are very rare. The base is taken at the incoming of silt and sand at the top of the Black Rock Member. Although this change is exposed in very few places, around the head of Swan Inlet it can be seen that the strongest negative break of slope on the south side of the ridge formed by the Black Rock Member actually lies within the mudstone sequence: the siltstones and fine sandstones which characterise the Shepherds Brook Member appear at a weaker negative feature somewhat further south, higher in the sequence.

The Shepherds Brook Member is typically made up of grey mudstones, grey or brownish grey siltstones and poorly sorted fine-grained sandstones, although at Port Sussex the top of the unit returns to mud-dominated sediments with only a few sandstone beds. Some beds have developed pervasive rusty alteration, taken to indicate weathering of volcanic detritus. The unit is mostly medium to thinly plane-bedded, with some lamination, and a few thick sandstone beds. A few sandstones contain mudflakes like those of the overlying Brenton Loch Formation. Burrows with spreite occur in places. Frakes and Crowell (1967) report a claystone with plant fossils near the top of the unit at Port Sussex, and during the present survey a siltstone in the lower half of the unit (exposed on the north side of Port Sussex [UC 6310 7380]) was found to contain probable plant debris.

The Shepherds Brook Member was probably deposited in a similar marine or lacustrine environment as the Black Rock Member but with slightly more energetic water circulation. It is approximately the same thickness as the Black Rock Member, ranging from about 150 to 200 metres on East Falkland. In Port Sussex, the outcrop of the Shepherds Brook Member gets narrower towards the west, suggesting that part of the unit may have been removed by strike-parallel faulting. Nevertheless, this is the only suitable type area.

The type section of the Shepherds Brook Member is in Port Sussex. The lower to middle parts of the member are seen about 500 metres north-west of the head of the inlet [UC 6329 7377] to [6320 7362] and require a low tide to be fully exposed. The middle to upper parts are exposed at the head of the inlet from [6361 7338] to [6351 7308], although there is a break in the section at the mouth of Shepherds Brook. Reference sections are exposed near the head of Swan Inlet on shorelines between [UC 9154 5630] and [9050 5665].

The age of the Shepherds Brook Member is constrained to the Late Permian by the probable ages of the Black Rock Member and of the Brenton Loch Formation. It is probably equivalent to the Collingham Formation (Ecca Group) of South Africa, as described by Visser (1992b).

2.4.4 Brenton Loch Formation

The Brenton Loch Formation is made up of mudstones, siltstones and fine sandstones. Parts of the formation are notable for impressively well-preserved trace fossils. It crops out in a broad belt across East Falkland either side of Brenton Loch and Choiseul Sound, in the Sea Lion Islands and in a narrow strip on the east coast of West Falkland.

This formation includes both the 'Choiseul Sound and Brenton Loch Beds' and the 'Lafonian Sandstone' of Baker (1924). Neither name is appropriate to a modern lithostratigraphic scheme. A formation name should include only one place name. Recognising this, Jalfin and Bellosi (1983) attempted to define a 'Formación Bahia Choiseul' to replace the 'Choiseul Sound and Brenton Loch Beds'. However, not only does their 'type section' not include the lower part of the formation, it overlaps with what is here regarded as the lower part of the succeeding Bay of Harbours Formation. Moreover, their section is locally disturbed by faulting on the margins of the Goose Green Graben. Strata exposed on the Grantham Sound coastline either side of Brenton Loch are considerably less deformed and better exposed than around Choiseul Sound. Both the base and the top of the formation can be seen there and it is likely that a complete composite type section can be described. Therefore the name 'Brenton Loch Formation' is adopted here. Scasso and Mendia (1985) assigned the portion of the Brenton Loch Formation which crops out on the east coast of West Falkland to the informal 'Formación Punta Larga', but there seems to be no need to place these outcrops in a different formation. The Brenton Loch Formation lies conformably between the Port Sussex Formation and the Bay of Harbours Formation.

Baker's (1924) 'Lafonian Sandstone' is now regarded as the basal Terra Motas Member of the Brenton Loch Formation (Section 2.4.4.1). It cannot be traced into eastern East Falkland with certainty. Two other members, the Cantera Member and the Saladero Member, are recognised in the type area. They cannot be traced into or west of the Goose Green Graben because of lateral variation in the sequence and increasing deformation. Thus the greater part of the Brenton Loch Formation outcrop remains undifferentiated. Nevertheless, the description of the Cantera Member can also be applied to the undifferentiated outcrops north of Choiseul Sound and on West Falkland. The outcrop south of Choiseul Sound, including the Sea Lion Islands, can be approximately correlated with the Saladero Member, although it also has similarities to the succeeding Bay of Harbours Formation. Lateral changes in the sequence from west to east in East Falkland are attributed to tectonic control of basin depth by the Goose Green Axis. This possibly represents a tectonic reversal compared with Port Sussex Formation deposition.

The Brenton Loch Formation outcrop is characterised by persistent linear ridges of uniform height which mark short sequences dominated by sandstone. These alternate with valleys of similar width, underlain by softer lithologies (mainly siltstone and mudstone). Photogeology shows clear lateral continuity of the larger sedimentary units in this 'ridge-and-valley' terrain in the area north-west of Camilla Creek. Between Darwin and Pleasant Roads the terrain is disturbed by gentle folding and strike-parallel faulting, and individual sandstone ridges cannot be correlated across the Goose Green Graben with confidence. It is clear that in general the ridge-and-valley topography is not a result of folding or faulting. Indeed, as it passes into areas where folding and faulting is more pronounced, the continuity of the ridges is obscured, not enhanced.

Nevertheless, some individual beds in the middle of the formation can be traced between the southern part of Swan Inlet and Darwin Harbour. This demonstrates that the north shore of Choiseul Sound lies approximately parallel to regional strike and so that it is unlikely to be controlled by a discordant tectonic structure, such as a thrust. Choiseul Sound presumably coincides with the part of the Brenton Loch Formation which is least resistant to weathering. This has been offset north-eastwards within the Goose Green Graben to underlie the low ground between Burntside and Laguna Babas, but otherwise it continues in Brenton Loch. The eastwards increase in the outcrop width of the Brenton Loch Formation, especially on the south side of Choiseul Sound, could be due to stratigraphic thickening, tectonic repetition, or perhaps just a decrease in regional dip.

The age of the Brenton Loch Formation is taken to be Late Permian. It is equivalent to the Ripon Formation and part of the Fort Brown Formation (Ecca Group) of South Africa, according to descriptions by Kingsley (1981).

2.4.4.1 Terra Motas Member

The Terra Motas Member consists mainly of medium and fine-grained lithic sandstones, with minor interbeds of laminated siltstones and mudstones. It can be traced from its type area, Terra Motas Point (on the south side of Port Sussex), almost as far east as Swan Inlet. In the rest of the outcrop it cannot be distinguished from other parts of the Brenton Loch Formation with any confidence.

Baker (1924) named the 'fairly fine-grained, uniform, soft, rather thin bedded brown sandstone' which succeeds the tillite as the Lafonian Sandstone. He estimated the thickness as 'about 300 feet' [about 90 metres] and described the upper limit as being at the lowest horizon at which the transformation from sandstone to laminated siltstone and mudstone takes place (i.e. on the south side of Terra Motas Point). Frakes and Crowell (1967) renamed this unit the Terra Motas Sandstone, apart from what Baker (1924) identified as fine-grained 'passage beds' at the base, which they placed in the Shepherds Brook Member (Section 2.4.3.3). Greenway (1972) was unable to differentiate the Lafonian Sandstone from the overlying beds.

Sandstones of the same type as those exposed in Terra Motas Point occur throughout the Brenton Loch Formation and the top of the Terra Motas Sandstone is difficult to trace consistently, especially in the east of East Falkland. Therefore there seems little justification for treating it as a separate formation and it is here regarded as a basal member of the Brenton Loch Formation. The Terra Motas Member is conformable between the Shepherds Brook Member of the Port Sussex Formation and the Cantera Member of the Brenton Loch Formation.

The Terra Motas Member forms a broad low ridge or series of closely spaced ridges without any persistent intervening valleys. Bedding traces are commonly visible on aerial photographs and locally are very distinct. At Port Sussex the lower boundary is rapidly gradational, with a few isolated beds of sandstone of 'Terra Motas' type up to 40 centimetres thick present in the topmost part of the Shepherds Brook Member. The base of the Terra Motas Member is defined as the first appearance of massive fine to medium-grained sandstone more than one metre thick or comprising more than 30 per cent of the section. At the head of Port Sussex the base coincides with an inflection in the

Plate 21: Sandstones and laminites, Brenton Loch Formation, Swan Inlet

 MN28249

A thin, brown-coloured tuff bed at the base is overlain by medium to thinly bedded, laminated and cross-laminated mudstones and siltstones. The base of a thin (about 25 cm) turbidite lies just above the hammer head. The lower part is in massive fine-grained sandstone, passing up into thin faintly plane-laminated sandstone, then climbing ripple cross-laminated sandstone. A thin mudstone bed drapes the current-rippled surface preserved at the top of the turbidite. That is overlain by the base of the succeeding turbidite, which is in massive to faintly laminated fine-grained sandstone with sparse mudstone intraclasts.

coastline. Elsewhere it is locally expressed by a weak negative break of slope on the northern side of the ridge formed by the Terra Motas Member.

The Terra Motas Member mainly comprises medium to very thickly bedded (some beds are up to about four metres thick), rather uniform medium and fine to very fine-grained feldspathic lithic sandstone, some silty or muddy. Generally it is massive but some beds show an upwards passage into faintly laminated or cross-bedded sandstone, and some are graded to siltstone or mudstone. Small subangular intraclasts of dark grey mudstone or siltstone are common, locally displaying diagenetic halos. The base of each sandstone unit is generally sharp and commonly erosive, locally with sole marks, ball-and-pillow structure or mudstone rip-ups. Syn-sedimentary faulting occurs locally. Baker (1924) states that 'in its lower part it [the Lafonian Sandstone] occasionally contains included fragments and small boulders of igneous and other erratics, such are found in abundance in the underlying tillite'. No such erratics have been described by subsequent workers and this seems to have been a rare error by Baker.

Petrographic examination of five specimens of medium to very fine grained sandstone found that they tend to be poorly sorted, with angular to subrounded grains. The major component is monocrystalline quartz, occurring with abundant feldspar, mica and sparse rock fragments, in a muddy, ferruginous matrix. The framework grains tend to have a ferruginous coating. The feldspars include both potassic and sodic/calcic varieties, and many are extensively altered. The rock fragments include chert, pyritised mudstone, sparse micaceous siltstone, and basic to intermediate volcanic rocks. In some specimens, pyrite is finely disseminated throughout the sandstone and also occurs as larger aggregates. One specimen of siltstone was found to be composed of angular to subrounded silt-grade quartz and mica in a laminated, micromicaceous and pyritic muddy matrix (Lott, 1997).

The sandstones alternate with thin to medium interbeds of dark grey to buff laminated mudstones and siltstones, typically with a characteristic striped appearance, with some thin bedded fine sandstone. There is commonly a upwards gradation from sandstones to laminated rocks and some of the laminites display graded bedding or small sand injection structures. These intervals are less than 10 metres in thickness and are subordinate to the sandstones.

Rare indeterminate plant debris occurs locally. Trace fossils occur near the top of the unit at Canada Runde Quarry [UC 838 571]. The most common is an irregularly meandering horizontal burrow up to two millimetres wide with an intermittent sand fill. Also found were a striated burrow (cf. *Scoyenia*), *Undichnia* sp. (*U. bina* and *U.* cf *insolentia*) and poorly preserved examples of *Umfolozia* (N H Trewin, written communication, 1998). These types all also occur in the Cantera Member (Section 2.4.4.2). No body fossils are seen.

The sandstones in the Terra Motas Member are interpreted as basin floor turbidites, with the intervening laminites representing periods of low energy deposition. The poor sorting in the sandstones is consistent with deposition by turbidity currents, but the uniform grainsize distribution suggests that any coarser sand had already been separated, for example by sedimentary sorting in a deltaic system.

The Terra Motas Member is estimated to be 250 metres thick.

The type area is Terra Motas Point [UC 586 755] and the ridge extending east back to the mouth of Shepherds Brook. The type section for the base of the member (and the base of the Brenton Loch Formation) is in the cliffs south-west of the mouth of Shepherds Brook [UC 6347 7310]. A well-exposed reference section within the Terra Motas Member is found at Canada Runde Quarry [UC 838 571]. Elsewhere the unit is very poorly exposed.

2.4.4.2 Cantera Member

The Cantera Member is composed of fine-grained feldspathic lithic sandstones and finely laminated siltstones and mudstones. This member forms the greater part of the Brenton Loch Formation and represents its most typical development. It has been differentiated from the other members only in the area between Grantham Sound and the Goose Green Graben.

The name proposed for this newly recognised division of the Brenton Loch Formation is taken from Cantera camp, which is the type area. In Spanish, 'cantera' is a quarry of the kind opened for rock or some industrial mineral. Presumably Cantera camp was once the source of some building stone, possibly that used in the construction of Saladero settlement. The Cantera Member is conformable between the Terra Motas Member and the Saladero Member of the Brenton Loch Formation.

The Cantera Member is composed of fine-grained sandstones similar to those of the Terra Motas Sandstone together with intervals dominated by distinctly striped laminated siltstones and mudstones with minor very fine sandstones. These intervals of laminated sediments vary from about ten centimetres to some tens of metres in thickness. The thickest are regular enough to form persistent broad valleys of uniform width, although interrupted at intervals by joints, folds and faults. Some sandstones do occur within these broad valleys, in addition to the laminites. This alternation of sandstones and laminites gives rise to the 'ridge-and-valley' terrain between Grantham Sound, Choiseul Sound and Port Pleasant. This topography distinguishes the Cantera Member from the Terra Motas Member, which lacks intervals of laminites thick enough to form broad valleys.

The base of the member is taken at the base of first thick unit of laminated siltstone above the Terra Motas Sandstone. For mapping purposes this is taken to be the negative break of slope on the north side of the first valley of the ridge-and-valley terrain (that is, on the up-sequence side of the ridge or ridges formed by the Terra Motas Member).

The sandstones are mostly fine-grained, varying to medium or to very fine-grained. As in the Terra Motas Member, they are feldspathic lithic arenites, with a proportion of volcaniclastic debris. Beds of volcaniclastic sandstone or of tuff up to about five centimetres in thickness occur rarely (Plate 21). Scasso and Mendia (1985) record intervals of up to four metres of interbedded black silty mudstones and tuffs near Shag Harbour. Where seen in weathered exposures the sandstones are a monotonous greyish brown colour, although where unweathered they are dark grey or greenish grey. They vary from thinly to very thickly bedded. In many exposures they appear to be massive, but a regular series of internal structures can be seen in places. The base of individual sandstone units is abrupt and commonly erosive, with tool marks and flutes in places. The lowermost parts of the sandstone units commonly contain concentrations of mudstone intraclasts up to 20 centimetres long which diminish in size and number upwards. Basal load structures are common. Massive sandstone in the lower part of such units gives way upwards to thin bedded or plane laminated sandstone, in some places with intraformational small-scale slump structures, then to cross-laminated sandstone, often with climbing ripples. Some units are graded. The massive sandstones are prone to spheroidal weathering.

Petrographic examination of four sandstones from the Brenton Loch Formation found that they are essentially similar to those in the Terra Motas Member (Lott, 1997; Lott, 1999).

Most of the laminites are composed of very regular planar alternations (couplets) of very thinly bedded to very thinly laminated, very fine sandstone or siltstone, and mudstone, which gives rise to a characteristic 'stripy' appearance (Plate 22). The average thickness of 50 randomly selected couplets near Camilla Creek was 0.35 millimetres and the thickest sandstone bed was 15 millimetres (Macdonald et al., 1996). Some of the thicker couplets are graded from silt to mud. Rare small mudstone intraclasts occur in the thicker sandstone units. Some individual beds of sand or silt are cross-laminated but there are no significant intervals of ripple cross-laminated sandstones. Small-scale prod marks and other sole marks, and some de-watering and other soft-sediment deformation has occurred locally.

Trace fossils on bedding planes in the laminites are common at some horizons in all parts of the Cantera Member. One fairly common type conforms to *Umfolozia longula*, an arthropod trail (Plate 23), but other species of *Umfolozia* are possibly present. Other types known to occur include *Kouphichnium* and *Undichnia* (trails left by the fins of swimming fish), with both *U. bina* and *U. insolentia* present in places. Some localities towards the top of the member show burrows in sand-mud couplets, including *Planolites, Skolithos* and *Diplocraterion*. There also are several unidentified (and possibly undescribed) trails and burrows (trace fossil determinations by N H Trewin, written communication, 1998). Pervasive bioturbation occurs rarely at the top of the member. Examples of *Umfolozia* are illustrated by Savage (1971) and Anderson (1981), *Umfolozia, Kouphichnium* and associated traces by Acenolaza and Buatois (1993) and *Undichnia* by Anderson (1976) and Buatois and Mangano (1993). The *Umfolozia* assemblage occurs within Permian fluvio-lacustrine deposits in Argentina, and in the Carboniferous to Permian Dwyka Group and Ecca Group lacustrine deposits in South Africa (Anderson, 1981; Acenolaza and Buatois, 1993). *Undichnia* occurs in glaciolacustrine settings in the Late Carboniferous in Argentina (Buatois and Mangano, 1993; Buatois and Mangano, 1995). Thus the trace fossil association found in the Brenton Loch Formation can be interpreted as an indicator of lacustrine, rather than marine, deposition.

Poorly preserved plant remains occur in some Brenton Loch Formation sandstones exposed on the east coast of West Falkland. These include sphenopsid stems and glossopterid leaves (Scasso and Mendia, 1985).

The sedimentary assemblage in the Cantera Member can be interpreted as turbidites deposited at the foot of advancing delta slopes, alternating with basinal fines. Although some of the laminated sediments were deposited under stagnant

Plate 22: Typical laminites, Cantera Member, Ceritos Arroyo

 MN28248

Plate 23: *Umfolozia longula* and other arthropod trails, Swan Inlet

 MN28250

conditions, others do show evidence of very low energy traction currents (isolated cross-laminae and very thin cross-beds) and oxic conditions (trace fossils). Halle (1912) and Baker (1924) identified the laminated rocks as varvites, implying that each couplet represents an annual cycle of deposition. This does not necessarily imply the proximity of glacier ice, only that the basin underwent annual freezing. The incursion of turbidites into the sedimentary basin seems to have followed a broad cyclicity of a longer period, which has given rise to the fairly regular alternations of sandstone-rich and sandstone-poor intervals underlying the ridge-and-valley terrain. This might be related to episodic volcanicity or tectonic uplift in adjacent terranes. Apart from the appearance of minor bioturbation near the top of the unit, there seems to be no systematic change in facies within the Cantera Member. In particular, planar laminites undisturbed by minor sedimentary structures (which might be thought of as the most distal facies) can be found throughout the Member, as can laminites which were deposited under slightly more energetic and probably more oxic conditions.

The Cantera Member is estimated to be 1750 metres in thickness.

The type area is the coastline of Cantera camp between Terra Motas Point [UC 586 755] and the north-facing bay north-north-west of Cantera House [UC 597 704]. The *Umfolozia / Undichnia* trace fossil assemblage has been found at the following localities: in the lower half of the Cantera Member, on the Cantera coast; [UC 6168 7330] (close to the base of Cantera Member), [UC 6222 7232], [UC 6232 7202], [UC 6167 7160]; near the middle of the Cantera Member, Swan Inlet [UC 9430 5178]; near the top of the Cantera Member, Camilla Creek [UC 6498 6192].

2.4.4.3 Saladero Member

The Saladero Member is composed of fine-grained sandstones and siltstones. This member has been differentiated from the rest of the Brenton Loch Formation only in the area between Grantham Sound and the Goose Green Graben.

The name proposed for this newly recognised division of the Brenton Loch Formation is taken from Saladero camp, which is the type area. The ruins of the 'saladero', a small establishment for the production of salt beef, can be seen on the west side of the Brenton Loch narrows [UC 587 690]. The Saladero Member is conformable between the Cantera Member of the Brenton Loch Formation and the Bay of Harbours Formation.

The Saladero Member forms ridges and valleys of similar general appearance as those of the Cantera Member, but which are less laterally persistent. This is taken to reflect a change in the underlying sedimentary sequence: the turbidite sandstones may be more channelized and so less laterally extensive, and the contrast between the turbidites and the intervening basinal sediments is less than in the Cantera Member. Furthermore, the terrain which lies along strike to the south-east of the Goose Green Graben is more subdued than near Brenton Loch, and this is thought to indicate an overall decrease in the proportion and grade of sand in the sequence. In comparing terrain formed by the Cantera and the Saladero Members, care must be taken to distinguish the topographic expression of a laterally variable sedimentary sequence from the effects of the folding and faulting seen in some parts of the Brenton Loch Formation. The base of the Saladero Member is taken at the eastern side of the Rat's Piece, a narrow but persistent ridge which can be traced into a chain of elongate reefs and islands down the east side of Brenton Loch. This ridge includes the first massive siltstone and mudstone in the sequence, and seems to mark a change in the texture of the landscape. This horizon cannot be identified with confidence to the east of the Goose Green Graben.

The Saladero Member is composed of fine-grained sandstones similar to those of the rest of the Brenton Loch Formation but the intervening laminites are more sandy and silty, and many more are cross-laminated. There are rare intervals of small-scale cross-bedded and current ripple cross-laminated fine sandstone, possibly confined to the basal part of the Saladero Member. No 'stripy' laminites, like those of the Cantera Member, are seen. There is also generally more bioturbation. This is often seen as individual burrows, but pervasive bioturbation locally gives rise to massive medium to thick siltstone beds (mostly at the top of the unit). A fine trace fossil assemblage, including *Umfolozia longula,* two species of *Undichnia,* 10 other named species and some un-named ones, occurs on Sea Lion Island (N H Trewin, written communication, 1998) but so far have been found nowhere else in the Saladero Member. No body fossils are seen. Halle (1912) and Baker (1924) recorded silicified wood at Arrow Harbour and at Walker Creek, respectively (Section 2.4.5), but other plant fossils are absent.

On the east shore of Camilla Creek, a few hundred metres below Low Pass [UC 6492 6180], there is an splendid set of south-westerly verging metre-scale syn-sedimentary recumbent slump folds, with a small overthrust bounding them 'upslope' to the north. Essentially similar structures are exposed on the east side of the entrance to Brenton Loch [UC 5922 7045], close to the base of the Saladero Member, but were seen nowhere else on the Grantham

Sound shoreline. This suggests that the two occurrences mark a single event horizon. The lateral displacement of this horizon within the Goose Green Graben is approximately the same as inferred for the base of the Bay of Harbours Formation (Section 4.2.5.1). Slump folds exposed on Sea Lion Island (Macdonald et al., 1996) might have formed during the same event, although corroborative evidence is required. While local slope instability is to be expected in a turbidite sequence, a regional slumping event might have been caused by tectonic disturbance in the Gondwanide fold-belt.

Compared with the Cantera Member, the Saladero Member represents an upwards increase in depositional energy in the intervals between turbidite units, with an associated increase in sand content and in biological activity. The disappearance from the sequence of varvites suggests a climatic amelioration. The lateral change to a more subdued topography south-east of the Goose Green Graben indicates more distal deposition, suggesting some tectonic control of basin depth by the Goose Green Axis. The trace fossil assemblage on Sea Lion Island is typical of deep-water lacustrine environments and resemble assemblages assigned to the *Mermia* ichnofacies described by Buatois and Mangano (1993; 1995) from the Late Carboniferous of Argentina (N H Trewin, written communication, 1998). The persistence of *Umfolozia* and associated trace fossils in the Saladero Member on Sea Lion Island is consistent with this being in a more distal part of the basin in which low energy conditions prevailed for longer. Jalfin and Bellosi (1983) interpreted their 'Formación Bahia Choiseul' as intertidal deposits, but this fails to account for the presence of the turbidite sandstones found throughout the Brenton Loch Formation. The difference between their interpretation and that presented here is in part due to their having included part of the Bay of Harbours Formation in their analysis.

The Saladero Member is estimated to be 1050 metres in thickness.

The type area is the coastline of Saladero Camp between Brenton Loch [UC 5875 6878] and Black Island [UC 5660 7100].

2.4.5 Bay of Harbours Formation

The Bay of Harbours Formation is composed of coarse to fine-grained sandstones, with siltstones and minor mudstones. Plant fossils are fairly common. The unit forms the greater part of Lafonia and the adjacent islands.

Baker (1924) referred to the 'thick series of alternating sandstones and shaly rocks' with the *Glossopteris* flora underlying most of Lafonia as the 'Bay of Harbours Beds'. This unit is here renamed the Bay of Harbours Formation. One facies member, the Egg Harbour Member, (equivalent to Baker's 'West Lafonian Beds') is recognised in the upper part of the formation in the west of Lafonia. The distribution of this facies is attributed to proximity to the relatively stable area of West Falkland, together with tectonic control on the Goose Green Axis. The 'Bay of Harbours Beds' and the 'West Lafonian Beds' were combined in the 'Formación Estrecho de San Carlos' by Jalfin and Bellosi (1983) but the subdivision is worth retaining (Section 2.4.5.1) and the units described here are close to those of Baker's (1924) concept. The Bay of Harbours Formation conformably overlies the Brenton Loch Formation. No overlying formation occurs onshore.

Compared with the Brenton Loch Formation, the Bay of Harbours Formation lacks the 'stripy' plane-laminated siltstones and mudstones, these being replaced upwards by more massive, medium to thick-bedded, generally bioturbated siltstones and very fine sandstones. Where silty laminites do occur, they lack the interleaved mudstone layers which confer the strong colour contrast in the stripy laminites seen in the Cantera Member. The Bay of Harbours Formation is also characterised by the common appearance of ripple cross-laminated sandstones in units from one metre to several tens of metres in thickness, with sporadic ripple-marked bedding planes, plus large-scale cross-bedded sandstones and plane-bedded sandstones, commonly associated with deposition in channels. Massive turbiditic sandstones do occur in the Bay of Harbours Formation, but are a less characteristic component than in the Brenton Loch Formation. No syn-sedimentary slump folds have been seen in the Bay of Harbours Formation, but as this phenomenon is very localised within the Brenton Loch Formation, it may not be diagnostic. *Glossopteris* is common and widespread in the Bay of Harbours Formation but absent in the Brenton Loch Formation.

The base of the Bay of Harbours Formation is gradational over several hundred metres in the Black Island area, on the Saladero coast. It is expressed by several changes in lithofacies, together representing an upwards increase in sand content, traction current strength and biological activity. However, for mapping purposes it is necessary to assign the boundary to a horizon which can be traced across unexposed ground on aerial photographs - and which can be correlated across the Goose Green Graben. Most of the Bay of Harbours Formation forms very subdued topography in which bedding traces are fairly distinct and regular, as seen on aerial photographs. Only a few low narrow ridges occur, in marked contrast to the distinct ridge-and-valley topography of the Brenton Loch Formation. On the foreshore, bedding is distinct and regular, appearing coarser than the very fine lineations visible on foreshore exposures of the Brenton Loch Formation. The lithofacies change that appears to correspond most closely to this change in topographical texture seen on the aerial photographs is the incoming of common indistinctly laminated to massive siltstones. This marks the start of a sequence in which there is less contrast between adjacent lithofacies than in the Brenton Loch Formation, and hence a less strongly 'striped' appearance on imagery. In the Black Island area, the western limit of the ridge-and-valley terrain occurs close to a relatively prominent ridge (or group of closely spaced ridges) which appears to be formed by one or more massive sandstone beds. While this type of sandstone is more characteristic of the Brenton Loch Formation than of the Bay of Harbours Formation, the base of this ridge forms the best mappable contact in this part of the sequence. It is assumed that a thick turbiditic sandstone would have been deposited rapidly over a wide area and so marks a time plane.

The sandstone marking the base of the Bay of Harbours Formation also appears between Darwin and Goose Green, where it forms a bed between five and 10 m thick, once quarried for building stone. This coincides with a subtle change in topography apparent both on satellite images and aerial photographs, and with the same change in the nature of the siltstones as seen near Black Island. Note also that Halle (1912) records *Glossopteris* plant fossils in the sequence between Darwin and Goose Green, which also indicates that the Bay of Harbours Formation is exposed there.

The boundary between the Bay of Harbours Formation and the Brenton Loch Formation becomes very indistinct to the south-east of the Goose Green Graben. In particular, the ridge-and-valley terrain of the Saladero Member is much more irregular than to the north-west. This is consistent with field evidence that the lithofacies contrasts between the Brenton Loch Formation and the Bay of Harbours Formation become less to the east. The boundary is again taken at what is interpreted as the base of a persistent sandstone layer between a more ridged terrain to the north and a smoother terrain to the south. Pyramid Point and Enderby Point appear to have been formed by this resistant sandstone, and so the formation boundary can be traced across Lively Sound.

The sandstones in the Bay of Harbours Formation are typically brown and grey feldspathic lithic arenites, commonly with small mud clasts but otherwise well-sorted. Most are fine or very fine-grained but medium to coarse-grained varieties are also present, especially in the north and west (Plate 24). Turbidites like those of the Brenton Loch Formation are present and these tend to form the thicker, more massive beds. De-watering structures, load-cast bases, slumps and syn-sedimentary faults are fairly common. In addition, large eroded channels appear in the sequence for the first time. These are typically several decametres metres across and up to five metres deep, infilled with fine, medium and coarse-grained sandstone with large-scale cross-bedding and plane bedding with primary current lineation. Current and wave ripple cross-laminated silts and fine sandstones in intervals from 10 centimetres up to some tens of metres in thickness are common. Sporadic ripple-marked surfaces are also characteristic (Plate 25).

Baker (1924, p. 21) notes the presence of a conglomerate near Dos Lomas creek (presumed to refer to New Haven [UC 47 63]) comprising 'many rounded boulders, up to and exceeding a foot in diameter, of intensely hard siliceous sandstone or quartzite, ... found imbedded in greenish-grey claystones'. No such conglomerate was found in that area during subsequent fieldwork (N H Trewin, written communication, 1997), or elsewhere in Lafonia. However, the small mud clasts which occur in some of the sandstones are locally enclosed in diagenetic sandstone nodules. Clusters of oblate spheroids of hard sandstone up to 15 centimetres in diameter, each with a mud clast at its core, have been observed lying on the ground. If these were seen in a section of deeply weathered bedrock, it is possible that they would be interpreted as water-worn boulders in a conglomerate. Also some massive sandstone beds are very hard and weather spheroidally, so that in some localities there are accumulations of large rounded boulders. These can bear a misleading resemblance to spheroids of weathered dolerite.

Petrographic examination of three specimens of moderately well sorted to well sorted fine to medium-grained sandstone from the Bay of Harbours Formation found that they were composed mainly of subangular to subrounded grains of monocrystalline quartz with abundant feldspar, mica and sparse rock fragments. Potassic and sodic/calcic feldspars are both abundant, although the potassic varieties were dominant in one specimen. The rock fragments were mainly polycrystalline quartz in one specimen, but include chert, pyritised mudstone, micaceous siltstone and rare volcanic grains in the others. The micas tend to be deformed and altered. Non-opaque heavy minerals such as epidote and zircon are present in one specimen. The cement or matrix is ferruginous and chloritic, in some cases with pyrite aggregates (Lott, 1999; Lott, 1997).

Massive siltstones, in beds ranging from thin to more than one metre thick, are common, with some compact mudstone and laminated shales. These finer sediments vary in colour from yellow or grey to green. Plane laminated and cross-laminated to massive siltstones and mudstones do occur, together with minor fine sandstones, but no sharply laminated ('stripy') silts or alternating silts and mudstones (as seen in the Brenton Loch Formation) occur. Intervals of some decametres dominated by siltstone and very fine sandstone are frequent.

Around North Arm, the Bay of Harbours Formation is predominantly made up of siltstones and very fine to fine-grained sandstones, with some muddy siltstones and rare medium to fine-grained sandstones. The sequence is generally less sandy than in the north. Most bedding and lamination is planar. Although cross-lamination, ripple-marked surfaces, cross-bedding, primary current lineations and isolated shallow channels do occur, they are less common than to the north. Channels as small as three metres in width and 30 centimetres deep occur near Fanny Cove. Ball and pillow structures are more conspicuous, especially near the base of the thicker units of sandstone.

The siltstones in the Bay of Harbours Formation are commonly bioturbated but distinct trace fossils are not usual. Those which do occur are burrows rather than trails. Halle (1912) illustrates an insect wing, some three centimetres long, which was found at 'Bodie Creek Head'. Leaf impressions of *Glossopteris,* usually marked by rusty staining, are widespread and common in the massive siltstones. Imprints of sphenopsid stems occur sporadically. Plant fossils also occur in other rock types but are then less well preserved. Some sandstones contain comminuted plant debris. Fragments of silicified wood are also widespread.

Plant fossils were first recorded in Lafonia by Andersson (1907) at Seal Cove, Bull Cove and on Speedwell Island. Nathorst (1906) examined the specimens from Speedwell Island, assigning them to the sphenopsid *Phyllotheca* and so revealing the occurrence of the Permian *Glossopteris* flora on the Islands for the first time. He also described coniferous branches, which he compared to *Voltzia heterophylla,* and elongate leaves which he ascribed to *Desmiophyllum.* Halle (1912) records that 'determinable plant remains' were collected along the shores from a place between Darwin and Goose Green to Bodie Creek, at 'Bodie Creek Head', Dos Lomas (New Haven), east of Low Bay, at North Arm and on Speedwell Island. Halle (1912) assigned these plant fossils to several species of *Phyllotheca* and of *Glossopteris.* In places these two genera were associated with each other. He also records two specimens of *Gangamopteris.* Seward and Walton (1923; Baker, 1924) discuss specimens of *Phyllotheca, Neocalamites* and *Glossopteris* found at Dos Lomas (New Haven), Egg Harbour, Cygnet Harbour, Speedwell Island, George Island, and North Arm. They draw comparisons with specimens collected from other parts of Gondwana, including India, Australia and Antarctica[3]. They expressed doubt over several of Halle's (1912) identifications, in particular concluding that *Gangamopteris* does not occur. During the present survey, plant fossils were found at various localities throughout the Bay of Harbours Formation, including many of those named.

D. Cantrill (written communication, 1996) notes that the specimen compared by Seward and Walton (1923) to *Neocalamites carrerei* could equally belong to *Asterocalamites* or *Paracalamites* as *Neocalamites.* He considers that the poorly preserved sphenopsid material so far collected from the Falklands is not age diagnostic. He points out that in order for *Glossopteris* leaves to be of any use in refining the age of the Lafonia Group within the Permian, large collections need to be made, preferably supported by reproductive structures. He believes Halle's (1912) identification of *Voltzia heterophylla* is erroneous, and that the leaves identified as *Desmiophyllum* could actually belong to any of a number of Permian or Triassic plant genera. He concludes that the plant assemblage collected from the Bay of Harbours Formation is typical of Permian floras and considers that the widespread distribution of *Glossopteris* coupled with the absence of *Dicroidium* indicates that no part of the sequence is Triassic in age. The potential uses and limitations of glossopterid plant remains in Permian stratigraphy in Gondwana are discussed more generally by McLoughlin (1993).

Halle (1912) noted that fossil wood (which he assigned to *Dadoxylon*) was found near Arrow Harbour and on Bodie Creek. Baker (1924) recorded it from Walker Creek and Fanny Cove. The localities at Arrow Harbour and Walker Creek may be in the distal facies of the Brenton Loch Formation. During the recent survey, fossil wood was found, or has been reported to occur at Saladero, on Barren Island, on the section of coast opposite Tiny Island, Bleaker Island, on Cattle Point, on Kelp Island (North Arm) and near Fanny Cove House. One verbal description suggests that a tree trunk some 40 centimetres in diameter and perhaps as much as four metres in length has been preserved.

[3] The Antarctic specimens examined by Seward and Walton (1923) include some collected, astonishingly, only about 480 kilometres from the South Pole by Dr E A Wilson during the 1910-12 British Polar Expedition. Wilson's commitment to science remains an inspiration.

Plate 24: Sandstones in the Bay of Harbours Formation

Photograph by Don Aldiss, BGS. MN28252

Plate 25: Wave ripple-marks, Bay of Harbours Formation

Photograph by Don Aldiss, BGS. MN28251

The Bay of Harbours Formation is thought to have been deposited in a delta advancing into the Permian lacustrine basin. The delta-top facies occurs in the west, shallowing upwards to wave-base at the top. Palaeocurrent directions vary from north-west to south-east. Variations in the petrographic composition of the sandstones, particularly in the composition of the feldspars and the rock fragments, suggests that detritus from at least two source terranes was accumulating in the basin.

The thickness of the Bay of Harbours Formation in northern Lafonia is estimated as 3000 metres. However, the number of faults in that area could have caused significant repetition within the sequence, so this is likely to be an overestimate. The proximity of Egg Harbour Member strata in the Swan Islands to the outcrop of the Brenton Loch Formation on the coast of West Falkland indicates either that a large part of the sequence has been removed by faulting, or that the Bay of Harbours Formation is considerably thinner in that area, with the Egg Harbour Member replacing most of the rest of the formation (Section 4.2.3.2).

The type area is the coastline on the northern side of the Bay of Harbours [UC 32, UC 42 and UC 41]. The best reference section is the coast in the Praltos area, particularly between New Haven [UC 47 63] and Black Island [UC 56 71]. The type section for the base is suggested as the coast adjacent to Black Island [UC 5620 7123] to [UC 5662 7100].

The Bay of Harbours Formation is of Late Permian age. It can be no older than the Port Sussex Formation, which probably extends into the Late Permian (Section 2.4.3), while the absence of *Dicroidium* noted above implies that none of the sequence is Triassic. A palynomorph assemblage from the Egg Harbour Member is comparable to material from the Permian Ecca Group of South Africa (Macdonald, 1996). According to the descriptions by Kingsley (1981), the Bay of Harbours Formation is approximately equivalent to the upper part of the Fort Brown Formation (Ecca Group) and the Koonap Formation (lower Beaufort Group) of South Africa.

2.4.5.1 Egg Harbour Member

The Egg Harbour Member comprises sandstones and shales with *Glossopteris* flora. It occurs only in western Lafonia and the adjacent islands in Falkland Sound. It is equivalent to Baker's (1924) 'West Lafonian Beds', a sand-dominated sequence above the less sandy 'Bay of Harbours Beds'. Present work suggests that this is a lateral facies member in the upper part of the Lafonia sequence west of the Goose Green tectonic axis. Moreover, no such place as West Lafonia exists, and the place name 'Lafonia' is pre-occupied, so a new name is proposed for this division. The name is taken from the type area, Egg Harbour.

The Egg Harbour Member overlies the rest of the Bay of Harbours Formation conformably, although the base is probably diachronous. It is likely that contemporary strata are present in eastern Lafonia but these are in a more distal facies and have not been separated from the underlying portions of the Bay of Harbours Formation. Indeed, strata close to the synclinal axis west of Adventure Sound could be younger than any part of the Egg Harbour Member. The top of the Egg Harbour Member is not seen, or is not yet recognised.

Qualitative comparisons of the sedimentary sequences in the Speedwell Island, Kelp Harbour, Egg Harbour and Praltos areas show that similar lithofacies are present throughout, suggesting that similar depositional environments prevailed and that the sequences should all be assigned to the same formation, namely the Bay of Harbours Formation. However, the appearance of the terrain on aerial photographs and satellite images varies across the western side of Lafonia. The higher parts of the sequence present a coarser texture, reflecting a slightly more rugged topography. It seems most likely that this is caused by the sequence in that area becoming more arenitic upwards, both in having more channel sandstone deposits, but also by having more sand in the background sequence between the channels. The general impression of the sequences as seen during fieldwork lends some support, but quantitative analysis of the sequence would be required to confirm this interpretation, or to suggest if there is some other explanation for the upwards change in the appearance of the sequences on remote images.

Therefore, the Egg Harbour Member is here defined only by its photogeological appearance. It is characterised by a topography of distinctly 'rough' texture, with numerous rather irregular sandstone ridges (where dipping) or steps (where subhorizontal). Lines of diddle-dee (*Empetrum rubrum*) growing on the ridges appear dark on the aerial photographs, enhancing the texture. The base of the unit can be traced with confidence only in the uniformly dipping strata in the north. It appears to step down sequence between the latitude of Cygnet Harbour and New Haven, indicating that it is diachronous over short distances. However, the interpretation of the boundary is complicated by the presence of strike-parallel faults of unknown but probably large displacement (Section 4.2.3.4). Moreover, at about the latitude of Glorious Hill, regional strike changes to east-west in an area of very gently dipping to subhorizontal strata. Thereafter to the south the boundary shown on the map is highly generalised and must be treated only as a general guide.

Some of the ground around North Arm, for example east of Mappa Ponds, and between North Arm and Northwest Arm, has a similar combination of rough, ridged, topography and lines of diddle-dee marking sandstone beds. These might indicate easterly extensions of the Egg Harbour Member but they are rarely as 'rough-textured' or prominent as in the west of Lafonia. It therefore seems that there is a facies change across Lafonia. Similarly the topography of George Island is markedly subdued compared with that of Speedwell, suggesting that whereas the latter is within the Egg Harbour Member, the former is not. Swan Island and West Swan Island also display series of sandstone ridges like those seen around Egg Harbour, rather than a clear division into the ridge-and-valley topography typical of the Brenton Loch Formation. This is supported by field observations. North Swan has relatively subdued topography and so remains as undifferentiated Bay of Harbours Formation.

The composition of the Egg Harbour Member is essentially similar to the more proximal, western portions of the underlying Bay of Harbours Formation (Section 2.4.5). As seen at Speedwell Island, it comprises two component lithofacies. The background sequence is mostly composed of thinly bedded to laminated mudstones and siltstones with medium interbeds of very fine sandstones. The sandstones display wave ripples, small-scale trough cross-bedding, climbing ripples, and dewatering structures. This lithofacies contains locally abundant *Glossopteris* flora and rare indistinct trace fossils, including isolated examples of *Skolithos.* Channels within the background sequence are marked by sporadic thick to very thick sandstone beds (up to seven metres) some of which can be seen to wedge out laterally. The channel lithofacies is made up of uniform medium to fine-grained sandstones, which varies from plane-bedded (with primary current lineation) to large-scale trough cross-bedded with lateral accretion surfaces. Slumps, load casts, and some small-scale cross-beds are also present.

Petrographic examination of four fine to medium-grained sandstones from the Egg Harbour Member showed that they are similar to those in the rest of the Bay of Harbours Formation. One contained only sodic/calcic varieties of feldspar. A ferruginous laminated siltstone was found to be composed dominantly of angular to subrounded monocrystalline quartz and feldspar grains in a micromicaceous matrix. The laminations are formed by different proportions of mica (Lott, 1999; Lott, 1997).

The Egg Harbour Member represents more proximal deltaic deposition than the rest of the Bay of Harbours Formation, with sporadic fluvial channels or tidal channels with high-energy flow regimes.

Assuming an average regional dip of 12 degrees between Egg Harbour and Port King, the thickness of the Egg Harbour Member can be estimated as about 2300 metres, if the effects of major faulting in that area are ignored. If the faulting has a net westerly downthrow, then this thickness is an underestimate. If the net downthrow is to the east, then it is an overestimate. The proportion of the Bay of Harbours Formation formed by the Egg Harbour Member seems to increase westwards across Falkland Sound.

The type area is Egg Harbour [UC 34 56] to [UC 36 52]. Reference sections are exposed on the southern shores of Kelp Harbour [UC 38 60] to [UC 43 58].

A palynomorph assemblage from the Egg Harbour Member is comparable to material from the Permian Ecca Group of South Africa (Macdonald, 1996). Like the rest of the Bay of Harbours Formation, the Egg Harbour Member is thought to be of Late Permian age.

3. MESOZOIC IGNEOUS ROCKS

3.1 Igneous Intrusions

Dolerite and basalt dykes are widespread and locally very numerous in West Falkland and some adjacent islands. Greenway (1972) estimated that between 300 and 400 are visible on aerial photographs. A few occur on East Falkland, with one on Lively Island. Igneous dykes of intermediate or acidic compositions also occur, but are mainly confined to the south-west of West Falkland.

Previously, the Mesozoic dykes have been described in three categories: east-west trending, north-south trending, and the Cape Orford dyke swarm (Greenway, 1972; Taylor and Shaw, 1989). Here the dykes are divided into seven groups according to their orientation, distribution and field characteristics. These are: the Cape Orford Swarm; the westerly Fox Bay Swarm, South Harbour Swarm and New Island Swarm; the northerly Sulivan Swarm and Saunders Swarm, and all dykes east of Falkland Sound (Figure 3.1). Not all individual dykes can be confidently assigned to a particular swarm. Baker (1924) considered that the dykes have a radial arrangement but this has not been confirmed by subsequent work. In addition, at least two thin sills are known to be associated with the dykes in the west of West Falkland, while South Fur Island appears to be part of a large sill.

Geochemical and petrographic analysis (Mitchell, 1988) (Cox et al. in prep; M J Hole, oral communication, 1997) shows there are at least four compositional types of basalt or dolerite (microgabbro) intrusion in the Falklands. Some geochemical data are given by Thistlewood et al. (1997). It is likely that when comprehensive data are available for the dykes then some overlap in composition of dykes from apparently different swarms will be found, or that some of the swarms described here can be further subdivided.

Most of these minor intrusions are thought to be of Early Jurassic age (c. 190 Ma) but the South Fur Sill is likely to be Middle to Late Jurassic in age, and some of the dykes could be as well. Nearly all the mafic dykes which cross-cut the Cape Meredith Complex pre-date the Port Stephens Formation and are described in Section 2.2.

Most of the dykes are marked by a photogeological lineament, reflecting a contrast in vegetation and, commonly, topography. The lineaments are mostly straight or slightly sinuous, indicating that the dykes are typically vertical or steeply dipping. They post-date all structures except for some major faults (Sections 4.2.4 and 4.2.5). Fieldwork shows that the overall distribution of dykes as determined by photogeology is generally correct, with many individual dykes being traceable on aerial photographs with considerable accuracy. However, a small proportion of dykes is not marked by lineaments and so cannot be mapped across unexposed ground without recourse to geophysical methods.

When unaltered the dolerite is typically dark bluish grey and can be extremely difficult to break. However, the dykes tend to be deeply weathered and most are unexposed, or are seen only as a scattering of brown-coloured spheroidal blocks ('corestones'). The style of weathering is well illustrated in a low shoreline cliff on the southern shore of Whisky Creek, near Port Stephens [TC 3860 2515]. Here a 25 metre-wide dolerite dyke has been entirely altered to ochreous clay minerals, which tend to mimic the original igneous texture of the dolerite. In the cliffs on New Island it can be observed that some dykes have been weathered to a depth of some tens of metres.

In contrast, in most places the rock adjoining a dyke has been recrystallised through contact metamorphism. Thus many dykes are marked by a semi-continuous ridge formed of hardened sandstone which stands up from the surrounding country by as much as 25 metres. Grandfathers Rocks, near Albemarle Station [TC 623 191], are a particularly spectacular example of this phenomenon, but usually the hardened sandstone is not exposed. In some cases the top of the ridge is divided by a shallow grassy gully containing the unexposed dyke. The extent of this hardening, and the size of the ridge, varies both with the size of the dyke and the composition of the local bedrock. In coastal exposures, many dykes are represented only by breaks in exposure in the cliffs or foreshore, in some cases with the baked sediments of the wall rock protruding on either side of a gully or recess. This is seen extremely well on Rocky Inlet about one kilometre north of Chartres [TC 878 677] (Luxton, 1994). Inland, deep weathering of dolerite dykes has in a few instances given rise to closed linear depressions a few metres wide and of the order of 30 metres long. Examples between Lake Orissa [TC 44 29] and Watchful Valley [TC 44 27] occur on relatively high ground more than a kilometre from the nearest shoreline.

Age relationships demonstrated by one dyke cross-cutting another are surprising difficult to determine, partly because inland exposures are so poor but also because there is limited overlap between the various dyke swarms. The few examples noted suggest that the Cape Orford Swarm is the oldest, followed by the westerly-trending dykes, then by the northerly-trending dykes. However, it is possible that dykes of distinctly different orientation could have been emplaced as co-eval conjugate sets, and quite likely that dykes of similar trend were intruded at different times. This inferred sequence should therefore be regarded with caution.

Amygdaloidal dykes of several different orientations occur from South Harbour westwards (Figure 3.1). The presence of amygdales indicates that the level at which the dyke is now exposed was relatively close to the surface when the dyke was emplaced.

3.1.1.1 Cape Orford Swarm

Although the existence of a distinct group of dykes in the Cape Orford area was noted by Mitchell (1988) and by Taylor and Shaw (1989), the recent survey has shown that this is more widespread and varied than previously suspected. Northerly-trending dykes occurring in the far south-west of West Falkland and adjacent islands, ranging in composition from basalt and dolerite to rhyolite, are here assigned to the Cape Orford dyke swarm (Figure 3.1).

The Cape Orford Swarm is marked on aerial photographs by numerous very fine lineaments on the foreshore, and to a lesser extent crossing rocky summits or ridges. Most are aligned NNE-SSW, but they vary through more northerly directions to NNW-SSE. Photogeological interpretation suggests that ENE-WSW dykes which occur on Weddell Island and Beaver Island might also be part of the Cape Orford Swarm. The fine photo-lineaments are too numerous to trace individually and many are too narrow or too short to be mapped, even at the scale of the aerial photographs (1:25 000). Field investigation of foreshore exposures shows that dykes in the Cape Orford Swarm are even more numerous than suspected from aerial photograph interpretation. Many of those seen on the foreshore are expressed as sand-filled gaps in otherwise continuous sandstone exposures. Some of these gaps possibly just represent open joints or sedimentary dykes (Section 4.4) but where these phenomena are known to occur elsewhere in the region, for example on New Island and east of South Harbour, the distinctive photogeological texture formed by clusters of fine lineaments is absent. Due to the deep weathering, the absence of dolerite at the surface on these lineaments does not mean it is absent at depth.

Examples of the Cape Orford dykes observed on the coast north of Cape Orford House [TC 26 26] range in thickness from 30 centimetres to three metres thick, commonly spaced less than one metre apart. Within a small area their orientation varies from N005° to N045°, with a lesser number between N330° and N360°. Some of the thinnest dykes are sinuous. Most dykes are subvertical, but westerly dips of as little as 60° also occur. M J Hole (oral communication, 1998) reports that on Dyke Island[4] [TC 3 3], dykes oriented NNW-SSE and up to five metres thick dip 35° to 50° in westerly or northerly directions. No consistent cross-cutting relationships between dykes of different composition or orientation were detected amongst dykes assigned to the Cape Orford Swarm. Observations during the present survey and by M J Hole (oral communication, 1998) suggest that the Cape Orford dykes pre-date the South Harbour dyke swarm. This relationship can be seen on the western shore of Anthony Creek [TC 4485 3229].

Although many dykes in the Cape Orford Swarm are basaltic or doleritic, some are more evolved, including examples of spherulitic rhyolite. Many of the dolerites are feldspar-phyric, in some cases with abundant zoned euhedral phenocrysts a centimetre or more across. Pale grey dykes of probable rhyolitic composition contain small phenocrysts of euhedral alkali feldspar, rounded quartz and euhedral pyrite or magnetite. Petrographic examination of a conspicuously plagioclase and clinopyroxene-phyric dolerite found textural evidence for low pressure crystallisation (Hards, 1997c). Olivine was found in two dykes tentatively assigned to the Cape Orford Swarm. These are an NNW-SSE amygdaloidal basalt dyke from Weddell Island, which contains olivine microphenocrysts, and a NNE-SSW olivine and plagioclase-phyric basalt dyke from near South Harbour (Hards, 1997a).

The extent of the characteristic lineaments shows that the Cape Orford dyke swarm is at least 20 kilometres wide (Figure 3.1), and it could extend even further west on Weddell Island. Only a few dykes of Cape Orford type are seen east of Shag Point [TC 37 31], but do occur in the South Harbour area. Narrow north-south trending

[4] Dyke Island appears to have been named for the local abundance of this kind of geological structure. The name is shown on a map of the Islands published in 1884 made by Fitzroy (Captain of HMS Beagle during the 1830's) and Sulivan (then his lieutenant). It was Sulivan who first observed the presence of dolerite dykes in the Falkland Islands (Darwin, 1846).

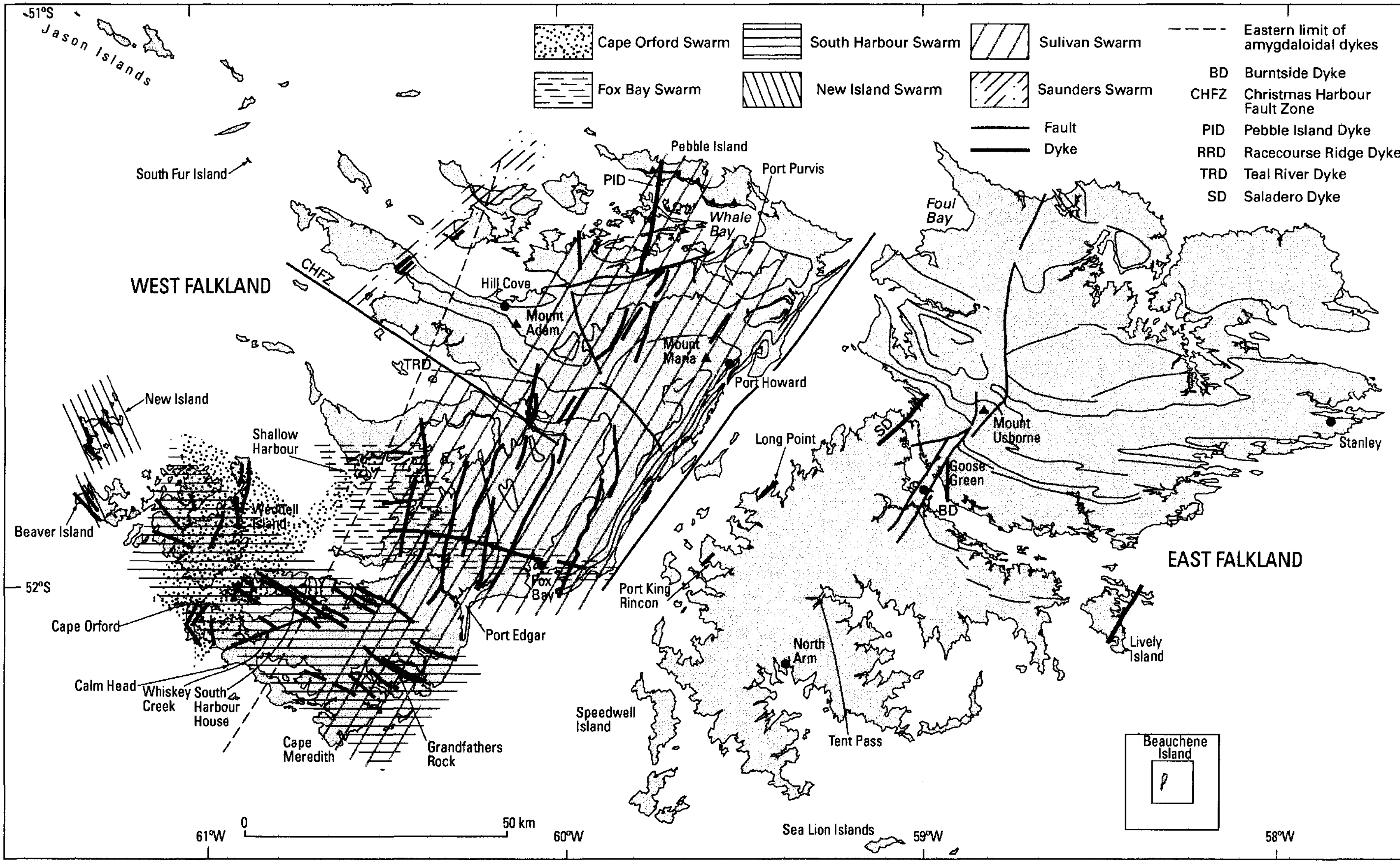


Figure 3.1 Mesozoic minor intrusions

plagioclase-phyric dykes at Mare Rincon (near Shallow Harbour) [TC 57 55] are probably also part of the Cape Orford swarm but there is no evidence that it extends as far as the Passage Islands.

The eastern limit of the Cape Orford Swarm thus approximately corresponds with the eastern edge of a complex positive gravity anomaly. Confidential aeromagnetic data shows a series of very high amplitude anomalies in the Weddell Island area which can be attributed to the Cape Orford dyke swarm. M J Hole (oral communication, 1997) notes that the moderate westerly dip of many of the dykes on Dyke Island would be consistent with an origin as cone sheets intruded from the west. Narrow dykes, and those of highly evolved composition, can be expected not to have been intruded far from their source. The sparse geophysical evidence, the evolved compositions, the abundance and small size of many of the dykes, together with the restricted extent of the Cape Orford swarm, suggest that a remnant volcanic centre lies a short distance offshore to the south-west of Weddell Island (Section 3.2).

3.1.1.2 Fox Bay Swarm

The Fox Bay Swarm is composed of a small number of broad, straight dykes oriented approximately east-west (Figure 3.1). All are rarely exposed inland, even as weathered corestones. The largest forms Racecourse Ridge, west of Fox Bay [TC 73 44], while smaller east-west dykes occur as far north as Dunnose Head settlement. At Fox Bay, the Racecourse Ridge Dyke is seven metres wide but thickens westwards to ten metres. One of the subparallel dykes at Fox Bay is five metres wide. Although the straight trace of these dykes suggests they are mostly subvertical, one just south of Dunnose Head settlement dips north at 50°.

Where examined, these dykes are all composed of olivine dolerite (Andersson, 1907; Mitchell et al., 1986; Hards, 1997a). Mitchell (1988) found that their compositional variation was controlled by fractionation of olivine with minor chrome spinel. They are thus fundamentally different to the hypersthene-bearing northerly dykes in the Sulivan Swarm.

Rare east-west dykes near South Harbour might also form part of this swarm. One a few metres wide on the high ground just east of Lake Orissa [TC 4581 2923] is marked by a train of cobbles composed of dark green-grey magnetite-rich aphyric dolerite. This train is interrupted by the trace of an unexposed WNW-ESE dyke, inferred to post-date it.

The Racecourse Ridge Dyke is seen on the ground only as a broad linear hollow along the crest of a ridge formed by recrystallised sandstones. Just east of Arroyo Malo [TC 8415 4171], this hollow is crossed by a train of corestone boulders marking a NNE-SSW dyke, implying that the NNE dyke is the later of the two.

3.1.1.3 South Harbour Swarm

South Harbour House [TC 43 31] lies in the main concentration of a locally dense swarm of westerly dolerite dykes extending from Beaver Island to Cape Meredith (Figure 3.1). Like the Fox Bay dykes, they are rarely exposed inland, even as weathered corestones.

Typically the dykes in the South Harbour Swarm are shorter than the Fox Bay dykes, and also narrower (mostly between two centimetres and four metres in thickness). Most are straight but a few are curved. In general, the adjacent sandstones have been baked over only a narrow zone although in some cases sufficiently to give rise to a topographic ridge. Most trend WNW-ESE, but their orientations vary between ENE-WSW and NNW-SSE. Although photogeological lineaments attributable to these dykes are very numerous near South Harbour (only a selection are shown on the map face), dykes seen in foreshore exposures are locally even more abundant than aerial photograph interpretation would suggest. For example, on the eastern shore of Whalebone Bay [TC 45 31], seven dolerite dykes, ranging in width from two centimetres to 2.2 metres, occur in a zone only 5.4 metres wide. In some places the thin dolerite sheets are separated by sandstone selvages less than 10 centimetres wide. Nine dykes on Ram Point [TC46 31] range from 32 centimetres to 3.5 metres in thickness, striking between N266° and N334°. Although mostly subvertical their margins dip between 56° north and 65° south.

Most of the South Harbour dykes are aphyric and uniformly medium to fine-grained, but some examples include plagioclase phenocrysts and display chilled margins. Some of those occurring from South Harbour westwards contain sparse amygdales up to eight millimetres in diameter, typically radially infilled by zeolite. Petrographic examination of a narrow (0.6 m) east-west dyke from Ram Point [TC 46 31] found it was a sparsely plagioclase-phyric amygdaloidal basalt, composed largely of clinopyroxene and olivine, with patches of cryptocrystalline material, probably after a glassy mesostasis (Hards, 1997a). In contrast to the broad north-south dykes of East and West Falkland, the South Harbour dykes only ever contain one pyroxene (augite) (M J Hole, written communication, 1999).

Many of the WNW-ESE dykes in the South Harbour area (at least as far south as Whisky Creek) were apparently intruded along minor faults. Although the strata on the southern side of each of these dykes retain the gentle northerly regional dip, a zone a few metres wide on the northern side has been tilted to dip north at up to 40°. Similar tilting is observed on the north side of an east-west trending dyke south of Dunnose Head settlement, in the Fox Bay Swarm. It is attributed to extensional faulting during dyke emplacement, with downthrow on the northern side.

Field observation and photogeological interpretation show a marked concentration of dykes close to a line passing through South Harbour House. In contrast, very few dykes crop out on the coast between the bay south of South Harbour House [TC 43 30] and the South Harbour farm boundary [TC 39 28] (M J Hole, oral communication, 1997). However, the apparent paucity of dolerite dykes in the high ground between Calm Head [TC 31 18] and Port Edgar (approximately corresponding to the outcrop of the Mount Alice Member) is probably due to the extensive cover of solifluction debris, including stone runs. Photogeological lineaments attributable to dykes are more frequent on the outcrop of the Albemarle Member and M. Hunter (oral communication, 1997) reports that dolerite dykes occur on the coast between Cape Meredith and Albemarle Station. There is also weak photogeological evidence that dykes of both WNW and NNE orientations are much more numerous in 'Ten Shilling Bay Rincon' (south-east of Port Stephens) than shown on the map.

3.1.1.4 New Island Swarm

About a dozen separate dolerite dykes occur on New Island, although some are marked only by photogeological lineaments and open gullies several metres wide in the cliffs. Where their attitude can be observed, the dykes are subvertical or dip steeply north-east. Although the New Island dykes are approximately aligned with the South Harbour Swarm, there are several differences. The New Island dykes are generally less closely spaced and they are broader, ranging from about two metres to 30 metres in thickness Most are oriented NW-SE. Although at least two NNW-SSE dykes transect the cliffs west of the settlement, exposures in 'Rookery Cove' [TC 0241 6160] show two very closely spaced dykes changing strike direction from N155° to N127° over the course of a few metres (Plate 30). Two of the dykes in Rookery Cove are zoned with weakly porphyritic and amygdaloidal centres, suggesting the effects of flow differentiation during emplacement.

In further contrast to the South Harbour dykes, a 4.5 metre-wide dyke on New Island is composed of olivine dolerite. It is similar to a sparsely amygdaloidal NW-SE dyke from near Shallow Harbour (Hards, 1997c), which might also belong to this swarm.

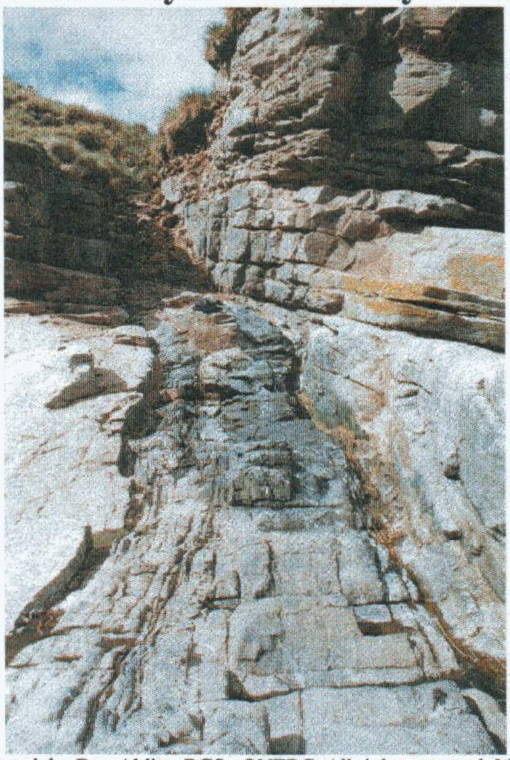
The WNW-ESE dyke exposed at Grandfathers Rock, near Albemarle Station [TC 62 19], is about 20 metres thick. It contains a few per cent of both olivine and orthopyroxene (Geochem Group, 1996). It is aligned with the South Harbour Swarm, but is more similar to the New Island dykes.

3.1.1.5 Sulivan Swarm

These northerly trending, thick dolerite dykes are widespread in West Falkland, forming a loose swarm which passes either side of Mount Sulivan, close to Lake Sulivan and Sulivan House (Figure 3.1). It possibly includes at least some of the dykes in East Falkland. The dykes are quite different to those at Cape Orford. They are relatively few in number, occurring either singly or in closely spaced groups. Some are sinuous, some bifurcate. Many can be traced individually for more than 10 kilometres. Most are between eight and 50 metres thick, although some are as thin as 1.8 metres. The adjoining zones of contact metamorphosed sandstone are correspondingly broad, giving rise to conspicuous linear ridges.

Brown (1967) made a detailed petrographic description of samples from the northerly-trending dyke near Black Hill House [UC 03 54], finding a mildly altered, coarse-grained, hypersthene-bearing dolerite with no olivine but abundant micrographic mesostasis, typical of quartz tholeiites. She compared it with a dolerite specimen from Port Sussex illustrated by Baker (1924). Mitchell et al. (1986) report that the northerly dykes between Fox Bay and East Bay are also hypersthene-bearing, contrasting them with the olivine-bearing Racecourse Ridge Dyke. Dykes at Teal River [TC8770 7455] and Top Hog Ground [TC 9515 6902] also contain orthopyroxene, probably as xenocrysts (Hards, 1997a). Thistlewood et al. (1997) found that northerly-trending Jurassic dykes at Cape Meredith contain orthopyroxene, sometimes mantled by clinopyroxene. Mitchell (1988) deduced that compositional variation in the northerly dykes is controlled by fractionation of pyroxenes. However, northerly trending dykes near Main Point [TD 9520 0250], near Chartres [TC 8708 6602], at Three Mile Ridge [UC 0846 8368] and at Crates Point [UC 0723 9605] contain small amounts (up to five per cent) of olivine and no orthopyroxene (Hards, 1997a).

Plate 30: Dolerite dykes in Rookery Cove, New Island



Photograph by Don Aldiss, BGS. MN28257

Plate 31: Aerial view of dolerite dykes, Chartres

Photograph by Don Aldiss, BGS. MN28258

The longest cluster of dykes extends almost 90 kilometres from the northern end of Port Edgar to Pebble Island (Figure 3.1) although none of the individual dykes in the cluster can be traced continuously over that distance. In places, the dykes have clearly been offset by faults (Section 4.2.5), most notably just south of River Harbour. Near Mount Arthur it is likely that the dykes have been obscured by solifluction deposits. Elsewhere, it appears that the dykes are simply discontinuous, at least at the present level of erosion.

It is possible that this elongated dyke cluster marks a deep structural discontinuity of some kind. It marks the eastern extent of displacement on the Christmas Harbour Fault Zone (Section 4.2.3.1; Figures 3.1 and 4.6) and it coincides with a break in a major NW-SE fault zone between Sound Bridge [TC 94 93] and Mount Moody [UC 12 63]. It also coincides with a major deflection in regional strike on the north-west side of the Hornby Mountains. To the east, northerly dykes are fewer and those near Blue Mountain have more complex outcrop patterns.

Other elements of the Sulivan Swarm show a loose association with D2 folds and faults, perhaps because their position is controlled by the same underlying structures (Section 4.2.2). Note that those in the west lie subparallel to the Symonds Harbour Anticline (Figure 4.2), whereas those to the east are more closely aligned with the Coast Ridge. Furthermore, as Greenway (1972) noted, the northerly trending dykes mostly occur north of latitude 52° S. This could also be a consequence of some deep structural discontinuity, perhaps that which controls the extent of the D2 structures between Port Edgar and Port Richards (Section 4.2.2.2). Some northerly trending dykes do occur further south, including some at Cape Meredith (Thistlewood et al., 1997). As with the South Harbour dykes, some may be obscured by superficial deposits.

There is uncertainty about the northern extent of the Sulivan Swarm. The Pebble Island Dyke (Figure 3.1) is clearly seen on the south side of Pebble Island [UD 10 14], where it is some 20 to 30 metres wide, lying within a low ridge. However, it cannot be traced northwards across the base of the Pebble Island thrust sheet (Section 4.2.4.1) although there is insufficient exposure to be sure that it does terminate there. Although the presence of dolerite boulders on the north shore of Pebble Island has been taken to indicate the northerly extension of the Pebble Island Dyke, it remains possible that the Pebble Island Thrust post-dates the dyke. The dykes near Blue Mountain [UC 04 56] possibly continue through the Hornby Mountains (hidden by solifluction debris), reappearing about five kilometres north-west of Port Howard, and extending to MacDonalds Rock [UC 23 86]. A small fault-bounded exposure of dolerite at the eastern end of Port Purvis [UC 3400 9995] is likely to be part of the same dyke (Baker, 1924), but there is no evidence that this dyke ever intruded the Pebble Island thrust sheet lying just to the north.

3.1.1.6 Saunders Swarm

A previously unrecognised compact swarm of NE-SW trending dolerite dykes lies between Saunders Island and King George Bay. It is here named the Saunders Swarm (Figure 3.1). These dykes are apparently offset across Byron Sound and across Brett Harbour, presumably by faulting. A vertical dyke about four metres wide, trending NNE, is exposed at the Neck on Saunders Island [TD 75 10]. Three photogeological lineaments of similar trend on Penharrow Point [TD 71 00] clearly mark dykes: they represent narrow gullies of uniform width which do not displace the steeply dipping strata which they cross-cut. Along strike to the south-west, at least four similar traces lie across Stevelly Hill, over a cross-strike distance of about 1.2 kilometre (about the same width as the cluster of lineaments at Penharrow Point). M. Hunter (oral communication, 1997) confirms the presence of at least four dolerite dykes on the north shore of Port North, along strike from the lineaments in Stevelly Hill. One NE-SW dyke lineament passes just west of Mount Pickthorne [TC 57 88].

The dyke at Saunders Island Neck is a heavily altered amygdaloidal dolerite composed of plagioclase, clinopyroxene and rare pseudomorphs after olivine (Hards, 1997b).

3.1.1.7 Dykes in East Falkland

Of the few dykes which occur on East Falkland most follow NE-SW trends. They might be assumed to represent a continuation of the Sulivan Swarm but this requires verification by geochemical analysis.

Baker (1924) noted the presence of at least two dolerite dykes at Port Sussex. One is represented only by a very faint lineament and a cluster of boulders on the shoreline close to the head of Port Sussex [UC 6358 7369]. The other occurs 700 m to the west [UC 6278 7393] and is more clearly defined. It can be traced south-westwards for more than 12 kilometres. At Saladero [UC 58 69] it coincides with the entrance to an old cattle corral and so is here named the Saladero Dyke. It is subvertical and is seen to vary from seven metres to 11 metres in thickness. It contains rare olivine and no orthopyroxene (Baker, 1924; Hards, 1997c).

Frakes and Crowell (1967, p. 42) note two small NE-SW dykes which intruded the tillite 'east of Ponds Rincon, Port Sussex'. It is not clear that this description refers to the two dykes found by Baker (1924).

The NNE-SSW Lively Island dyke is more than 30 metres wide (Macdonald et al., 1996) and can be traced for about 12 kilometres. In addition to plagioclase and clinopyroxene it contains about seven per cent orthopyroxene (Geochem Group, 1996).

At least two NE-SW dykes some eight to 10 metres thick occur on Long Point [UC 34 56], one of which was noted by Baker (1924). These are petrographically similar to the dyke at Three Mile Ridge, containing about five per cent olivine (Hards, 1997a). A NNE-trending dyke of somewhat greater width is exposed on the coast of Port King Rincon (Figure 3.1) but its inland extent is not known (Macdonald et al., 1996).

No other dykes are known to occur on Lafonia although a single piece of weathered dolerite was collected by the stream near Tent Pass [UC 41 34]. Note that many of the massive fine-grained sandstones in the Bay of Harbours Formation in Lafonia develop spheroidal weathering. They then resemble weathered spheroids of dolerite, requiring close inspection to distinguish.

A northerly-trending dolerite dyke crops out near Burntside House [UC 68 60] but it is very poorly exposed and its continuation to the north and south cannot be traced. It contains no olivine or orthopyroxene (Hards, 1997a). The occurrence of dolerite at the west end of Teal Creek Pond [UC 67 55] could be a continuation of this dyke, and if so, it seems most probable that it has been offset by the fault extending NNE from Darwin Narrows. If a northern continuation of the dyke is present, it is evidently hidden in the poorly exposed ground in the northern end of the Goose Green Graben, and possibly beneath the thick solifluction debris in the mountains.

It is very likely that igneous dykes occur in the northern part of East Falkland. Wyville Thomson (quoted in Andersson, 1907, p.2) reports a 'big block of diorite containing some crystals of augite' found in a stone run of unspecified location. Loose pieces of dolerite were collected during road building near Romford Brook and east of Cerro Montevideo (R. Hancox, oral communication, 1996). The apparent absence of dolerite dykes in the north may be a consequence of their being hidden by the widespread superficial deposits. On the other hand, there is no reason to suppose that dolerite dykes are any more common there than to the south of the Wickham Heights.

Interpretation of ship-borne magnetometer profiles suggests that a group of five to ten near-vertical dolerite dykes occurs some 20 kilometres offshore, south-east of East Falkland between Choiseul Sound and Port William. They are oriented north-south and are reversely magnetised (Barker, in press).

3.1.1.8 Minor sills

Two subhorizontal sills, both about 50 centimetres thick, occur in association with dykes in the west of West Falkland. One crops out on the foreshore just west of South Harbour House [TC 43 31] (Macdonald et al., 1996). The other occurs on the southern shore of Mare Rincon, Shallow Harbour [TC 550 549]. Both are assumed to represent local deviations by dykes along bedding planes.

3.1.1.9 South Fur Sill

I J Strange (oral communication, 1996) first observed that South Fur Island is composed largely of dark-coloured rock and so is quite different to any of the nearby islands. A published photograph (Strange, 1987, p. 91) shows joint-bounded blocks of dolerite several metres across. R W Woods (oral communication, 1998) found that some of the dolerite beach boulders on South Fur are cross-cut by white veins (about one centimetre thick) here interpreted as traces of microgranite evolved from the dolerite by magmatic differentiation (see also Section 3.2).

Petrographic examination shows that the dolerite boulders on South Fur Island contain as much as 55 per cent olivine, some poikilitically enclosed in clinopyroxene (Hards, 1997b), and so are very probably cumulates. M J Hole (oral communication, 1998) confirms suggestions that the South Fur dolerite is different from other olivine dolerites in the Falklands: it is very fresh and contains only one pyroxene (augite), whereas the others contain both augite and pigeonite, many together with orthopyroxene.

The apparent extent of the dolerite on South Fur Island indicates that it is part of a large sill. This conclusion is supported by the occurrence of cumulate textures and probable veins of granitic differentiate.

3.1.1.10 Age and palaeomagnetic studies

A small number of radiometric age determinations of the Falklands dolerite dykes have been published, some in conjunction with palaeomagnetic studies. These indicate that most of the dykes are about 190 Ma old, in the Early Jurassic, but that some may date from the Middle Jurassic.

Whole rock K-Ar analysis of a dolerite from near Cape Meredith by Cingolani and Varela (1976) yielded an age of 192 ± 10 Ma.

Using the more reliable ^{40}Ar-^{39}Ar whole rock analysis, Mussett and Taylor (1994) estimated a maximum age of 193 ± 4 Ma for two north-south dykes from the Sulivan Swarm (one from Spring Point and one from Fox Bay). They determined the age of two WNW-ESE dykes of the South Harbour Swarm (from near Mount Alice) as 188 ± 2 Ma, and one from the Cape Orford Swarm (from near Cape Orford itself) as 190 ± 4 Ma. As they point out, these determinations indicate that all the dykes are about the same age of roughly 190 Ma. No conclusions about the relative age of the dyke swarms can be drawn from these radiometric data.

Two northerly-trending dolerite dykes from Cape Meredith, apparently part of the Sulivan Swarm, yielded Middle Jurassic ages of 176 ± 7 and 162 ± 6 Ma (K-Ar, whole rock) (Thistlewood et al., 1997). These authors point out that the discrepancy with the previously published data could simply be due to argon loss from their samples, but also that the determinations could indicate the real age of these dykes. They would then be of similar age to that of the Ferrar Group of Antarctica and of Karoo basalts in southern Africa.

In Section 3.2 it is argued that the South Fur Sill intruded a sequence of Middle to Late Jurassic volcanic rocks, implying that it is no older than the Middle Jurassic.

Mitchell et al. (1986) reported palaeomagnetic evidence from West Falkland that suggests the Islands had been rotated by some 120° during the early stages of break-up of the Gondwana supercontinent. They sampled four dykes west of Fox Bay, in the Sulivan Swarm, and metamorphosed shales adjacent to the Racecourse Ridge Dyke. They found that a geomagnetic polarity change had occurred between the emplacement of the northerly and westerly dykes.

In a follow-up study, Taylor and Shaw (1989) carried out palaeomagnetic determinations on thirteen dolerite dykes from West Falkland. They concluded that prior to the opening of the South Atlantic, the crustal block containing the Islands had undergone a clockwise rotation of 100° and that it had moved some 500 kilometres southwards from its position within Gondwana, close to the south-eastern coast of South Africa. They suggest that this occurred between about 200 and 125 Ma years ago (Section 6).

3.2 Volcanic Rocks

No deposits of volcanic rocks have been found on the Islands. Nevertheless, the occurrence of agate and of a variety of acidic volcanic rocks as beach pebbles indicates that elements of a bimodal volcanic suite are exposed on the sea floor near the Islands. This is corroborated by the occurrence of hornfelsed acid volcaniclastic rocks in close association with the large dolerite sill forming South Fur Island.

Pebble Island has been famous for the occurrence of agate pebbles ('Falklands Pebbles') since the earliest days of settlement. Moody (1842, reprinted 1969), noting that 'many [of the pebbles from Pebble Island] being beautiful specimens of onyx and bloodstone', was probably referring to agate and jasper. The agate (banded chalcedony) typically varies from pale to mid bluish grey, but very dark grey and black varieties also occur, together with carnelian. The pebbles are mostly very well-rounded and subspherical, commonly with chitter-marks as a consequence of high-energy abrasion on beaches, and all with a very thin cloudy patina. They are typically up to about five centimetres in diameter, but an occurrence 'the size of a beer can' was reported during the survey.

Certain beaches are known as reliable sources of these pebbles, particularly some of those in the west of Pebble Island, and on Pebble Islet. However, agates can be found on other beaches on Pebble Island and elsewhere in the north of the Islands. During the survey agate pebbles were found or reported to have been found at The Neck on

Saunders Island, Pebble Beach and adjacent parts of White Rock Bay, Paloma Beach, and the small beaches between Cow Bay and Cape Carysfort.

Agates form in cavities in volcanic rock, particularly in basalts. Basalts tend to be susceptible to weathering, leaving loose fragments of the more resistant agate. Where agates occur on the Falklands they are usually associated with other erratic pebbles and cobbles, chiefly acid volcanic rocks. Indeed, such erratic volcanic pebbles appear to be more common and widespread than the agates, presumably reflecting the relative abundance of the two types at source. They are also all very well-rounded and occur up to about 15 centimetres across. The most common erratic volcanic rocks are rhyolites, some of which are aphyric although many contain alkali feldspar phenocrysts, commonly together with quartz phenocrysts. Spherulitic texture occurs in some specimens. Rhyolitic volcaniclastic rocks also occur, including ignimbrite, lapilli tuff and fine-grained bedded tuff. Pebbles of fine-grained amygdaloidal metabasalt and porphyritic basalt have also been collected, in addition to some non-volcanic rock types including hornblende metagabbro and hornblende schist, but these are much less abundant than the rhyolitic rocks. Although these mafic rocks do crop out elsewhere in the Islands, they have all been found as pebbles in places with no known immediately local source. Petrographic examination shows that the erratic volcanic rocks are heavily altered, with growth of zoisite, epidote and chlorite, and probable secondary silicification, typical of hydrothermal alteration in volcanic sequences (Hards, 1998).

Like the agates, pebbles of acid volcanic rock seem most common on beaches in the north of the Islands, having been found in localities from Steeple Jason to Cape Carysfort, but they also occur widely elsewhere close to the open sea, including New Island, Shallow Harbour, Bold Cove, Poke Point, the North West Islands, and on Lafonia adjacent to Blind Island. They occur both on modern beaches and in raised beach deposits.

The origin of the agate pebbles has long posed a problem. Although some of the dolerite or basalt dykes exposed in the Islands are amygdaloidal (Section 3.1), none were seen to include agate, and there are no exposures of basaltic volcanic rock. Likewise, there are no known outcrops of rhyolitic volcanic rock on the Falklands. Moreover, agate and rhyolite are always found as well-rounded pebbles at or close to exposed coasts, never as angular weathered debris in soil, or anywhere inland, or even in deep tidal inlets.

Erratic pebbles and boulders of a variety of compositions occur in many places on or close to outcrops of the Fitzroy Tillite Formation and it might be supposed that this is the derivation of the volcanic pebbles. However, volcanic rocks are not common as clasts in the tillite (Section 2.4.2), the rhyolite and agate pebbles are not concentrated close to tillite outcrops, and they are not usually associated with an abundance of other types of erratic, such as granite, that *are* common in the tillite.

One author (EJE) considers that the agates have been released by weathering from pebble lags in quartzites and sandstones of the Port Stanley Formation and Port Stephens Formation, and that they are not derived directly from local outcrops of basalt. However, pebbly quartzites are not particularly voluminous in the Falklands and many of the agates are larger than any intraformational pebbles. So far, loose agate pebbles have not been found inland on outcrops of the Devonian sandstones.

A volcanic unit, taken to be equivalent to elements of the Jurassic to earliest Cretaceous Chon Aike volcanic province (including the Tobifera Series) of southern South America (Gust et al., 1985; Pankhurst et al., 1998), has been encountered in a number of offshore boreholes between the Falklands and Argentina, and possibly occurs within 60 kilometres south-west of West Falkland, on the eastern margin of the Malvinas Basin (Richards et al., 1996b, figs. 2, 8). Richards et al. (1996b) also postulate that the North Falkland Basin contains Middle to Late Jurassic volcanic rocks, as seen in the San Jorge Basin, north-west of the Falklands (Fitzgerald et al., 1990). The Chon Aike province includes both acid and basic volcanism associated with crustal extension, although acid volcanism predominates. The rhyolitic pebbles found in the Falklands are similar to rock-types which occur in the Chon Aike province, both mineralogically and chemically (Hards, 1998). Although this does not of itself prove they are from related outcrops or even that they are of the same age, the extent and inferred mode of formation of the Chon Aike Province suggests the correlation is extremely probable.

Transport of pebbles and cobbles across the sea floor from these distant occurrences seems unlikely. Carriage by floating ice during periods of glaciation might seem to be a possibility, but is unlikely as a major source of erratic material in the Falklands. Human activity can also be dismissed as a major cause. Some rock was brought to the Islands as ships' ballast or as weights in fishing nets (Hards, 1998), but the erratic volcanic pebbles are not associated

Plate 32: Hydrothermal alteration in sandstones, Steeple Jason

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with any obvious ballast material, nor are they concentrated at ports or near sites of known wrecks. Human activity would not explain their occurrence on raised beaches.

Strangely enough, it is far more likely that pebbles and small cobbles have been carried to the Islands in the digestive tracts of penguins or seals. For example, diet studies of Magellanic and Gentoo Penguins reveal that their stomachs generally carry a small number of rounded pebbles between two and 15 millimetres in diameter (Jeremy Smith, oral communication, 1997; Stonehouse, 1967). Pebbles of this size are commonly seen around colonies of these species, presumably where they have been regurgitated. Likewise, Southern Sea Lion (which were once much more numerous in the Falklands) are known to swallow pebbles up to at least 10 centimetres across. The reasons for this have not been demonstrated, but theories include buoyancy adjustment, digestion and internal parasite control (Hamilton, 1934; Taylor, 1993; Fedak et al., 1994). All these species are wide-ranging, and it is within their capabilities to dive to the sea floor between the Falklands and the Argentine, for example (e.g. Thompson et al., 1998).

Nevertheless, if all the erratic agate and rhyolite pebbles in the Falklands had been carried in by wildlife then one would expect them to be randomly distributed around the Islands, with local concentrations at past colonies. While pebble spreads do occur in areas of burnt and eroded tussac, many of which may have once been used by Sea Lion, the volcanic pebbles and agates appear to be concentrated in the north, with scattered examples elsewhere. Furthermore, it is said that on beaches where agates are collected regularly, their abundance increases after storms, suggesting that fresh material has been brought up from lower in the tidal zone or from offshore deposits.

At times during the Quaternary, sea level at the Falklands has been as much as 50 metres below its present level (Section 5.2.10). If storm wave-base is taken at, say, 30 metres, it can be seen that the sea floor in present water depths of up to 80 metres could have been liable to erosion. This encompasses a considerable area around the Islands from which, given favourable near-shore currents and seabed topography, eroded material could be carried to the shoreline and incorporated in beach deposits. The possibility of this occurring is shown by the occurrence of tillite cobbles up to 20 centimetres long on beaches near Cape Pembroke [VC 50 73], more than 20 kilometres from the nearest known exposures of the Fitzroy Tillite Formation (although sea floor outcrops might be closer). On balance it seems most likely that the bulk of the erratic volcanic pebbles has been carried to the shore by wave and tidal currents from sea floor exposures of volcanic rock close to the Falklands. One author (DTA) considers it most likely that agate-bearing basalts occur offshore together with rhyolitic volcanic rocks, and that these are the source of the 'Falklands pebbles'. These sea floor exposures are most probably in the north but possibly occur elsewhere. A proportion of the erratic pebbles (including the more scattered examples) could have been brought in by seals and penguins, and these might include material carried for long distances.

In the absence of aeromagnetic surveys in the relevant areas, there is very little direct evidence for sea floor exposures of volcanic rock close to the Islands. Well-rounded pebbles of a distinctive fine-grained dark grey rock are locally abundant amongst the dolerite boulders on the shores of South Fur Island (R W Woods, personal communication, 1997). Petrographic examination shows that these pebbles include spotted hornfels formed by contact metamorphism from volcanic rocks, including a fine-grained bedded volcaniclastic rock (Hards, 1998). The abundance of these pebbles suggests that a source outcrop lies nearby, and it is reasonable to suppose that the hornfels is part of the host rock of the South Fur Sill (Section 3.1.1.9). In other words, the sill was intruded into a sequence of volcanic rocks which at that point lies only 15 kilometres from the nearest shore of West Falkland. Pieces of apparently similar hornfels occur as beach pebbles on The Twins and, rarely, on the south coast of Pebble Island.

The local occurrence of spectacular ferruginous staining along joints in sandstone on the south-west shore of Steeple Jason [TD 0520 3735] (Plate 32) is possible circumstantial evidence for the presence of volcanic rocks on the sea floor near the Jason Islands. Staining of this kind is known from nowhere else in the Islands, and it is here thought to result from the oxidation of iron sulphide minerals originally introduced into the sandstone by percolation of mineralising fluids along a pre-existing joint system. Such fluids could be of hydrothermal origin, plausibly reflecting proximity to a volcanic centre, although they might instead have arisen during de-watering of a nearby sedimentary basin, perhaps one marked by the large negative gravity anomaly to the south and west of the Jason Islands.

In Section 3.1, it was argued that a remnant volcanic centre related to the Cape Orford Swarm possibly lies a short distance offshore to the west of Weddell Island. If so, the sparse evidence for the age of the Cape Orford dyke swarm suggests that this would date from the Early Jurassic, prior to the eruption of the Chon Aike volcanic deposits.

Pieces of pale grey pumice are quite commonly found on Falklands beaches. Although these pieces range in size up to about 30 centimetres across, most are smaller and spreads of well-rounded pumice gravel occur in places. Although pumice is a volcanic rock, it floats in water, and so can be carried great distances by ocean currents. It is assumed that the pumice found on the Falklands originated at volcanic centres in the Scotia Sea. It is reported locally that the abundance of pumice on Falklands beaches increased greatly following the Deception Island eruption of 1967.

4. PERMIAN TO MESOZOIC STRUCTURE AND METAMORPHISM

4.1 Overview

Permo-Triassic and later Mesozoic deformation in the Falkland Islands can be divided into at least five phases of folding and faulting, or of faulting with negligible folding. The earliest, D1, is represented by an approximately east-west belt of folding and subparallel faulting which extends throughout the northern part of East Falkland, except for the furthest north, and through most of the north of West Falkland (Figure 4.1). It is likely that some of the structures in the far north of the Islands, although aligned subparallel to the D1 structural grain, were formed during later (D4 and D5) phases of deformation.

The second phase of deformation, D2, was responsible for the formation of a series of large-amplitude folds trending between north-south and ENE-WSW and found at intervals throughout the Falklands, except in the eastern part of East Falkland (Figure 4.2). These folds all seem to be associated with blind faulting at depth and can be regarded as drape folds or, in one case, a hanging wall fold. In some places the faulting associated with D2 structures appears at the surface.

The third phase, D3, appears to have occurred during a dextral rotation of the Islands but its main effects are confined to displacement, partly by strike-slip faulting, of older structures in a zone adjacent to Falkland Sound. (Figure 4.6).

The fourth phase, D4, is represented by brittle deformation in a high-level compressional regime, generating at least one major thrust sheet in the north of the Islands (Figure 4.8).

The final phase, D5, gave rise to a series of WNW-ESE through to ENE-WSW-trending extensional faults in the north of the Islands, at least in part reactivating D4 thrusts and probably also some older structures (Figure 4.10). Faults attributed to this phase include many that displace dolerite dykes inferred to be of Early Jurassic age (c. 190 Ma) (Section 3.1). However, late high-angle brittle faulting may have occurred in several phases over a period of time, perhaps including the Cenozoic.

These five phases of deformation are discussed in more detail in the following sections.

Permian to Mesozoic regional metamorphism in the Falklands generally attained only very low metamorphic grades in the late diagenetic zone and anchizone, only locally reaching the epizone (greenschist facies) (Figure 4.11). Field observation and conventional petrographic methods yield only a very general appreciation of variations in metamorphic grade, so the regional metamorphic history has been investigated through the techniques of clay mineral maturity and vitrinite reflectance. The results are summarised in Section 4.3. Examples of contact metamorphism and possible hydrothermal metamorphism have been described in Section 3.

Tectonic and metamorphic events older than the Permian are represented only in the Cape Meredith Complex (and in erratic pebbles derived from allied terranes and found in the Fitzroy Tillite). They are described in Sections 2.2 and 2.4.2. Quaternary earth movements are mentioned briefly in Sections 5.2.8 and 5.2.10.

The D1 fold belt in the Falkland Islands can be correlated with the eastern part of the Cape Fold Belt, forming one line of evidence for the previous position of the Islands within Gondwana (Adie, 1952a; Marshall, 1994b; Curtis and Hyam, 1998) (Section 6). However, in the eastern Cape Province, the exposed width of the fold belt is in excess of 130 kilometres, whereas in the Falklands it is less than 80 kilometres. It may be that part of the fold belt lies offshore to the north of the Islands, or that it was displaced or obscured by younger structures (such as D4 thrusting) during Gondwana break-up. Folding and thrusting in the Cape Fold Belt is thought to have occurred during regional deformation commencing during the Permian and continuing into Triassic times. There were at least four phases of compressional movement (Halbich, 1992). This deformation, described as the Gondwanide Orogeny, is attributed to convergent movement of tectonic plates at what was then the south-western margin of the Gondwana continent (Section 6). A volcanic arc formed at this convergent plate margin, some distance to the south-west of the inferred position of the Falkland Islands at that time, and contributed much detritus to the Permian sedimentary sequences of the region (Johnson, 1991; Johnson et al., 1996; Dalziel and Grunow, 1992; Cole, 1992). Remnants of the once-continuous Gondwanide fold belt are found in parts of Antarctica and South America (de Wit and Ransome, 1992; Curtis and Storey, 1996; Rossello et al., 1998) (Figure 6.2).

4.2 Folding and faulting

4.2.1 First phase of deformation (D1)

Folds and coeval faults formed during the first phase of Phanerozoic deformation (D1) are found in a broad area of northern East Falkland, although not in the most northerly parts of the Island, and in the northern part of West Falkland and most of the adjacent islands (Figure 4.1). The regional strike of D1 structures is east-west, modified to NW-SE by later deformation in some areas. To the north, D1 structures were over-ridden by D4 thrust sheets.

The areas where D1 structures are seen can be divided into four zones, separated by three major tectonic lineaments: the Goose Green Axis (following sections), the Falkland Sound Fault Zone (Section 4.2.3.2), and the South Jason Line (Section 4.2.2.3) (Figure 4.1). Structural style and intensity vary from one zone to the next (Plates 26, 27) but contrary to previous suppositions (Marshall, 1994b; Curtis and Hyam, 1998), the change in intensity of regional deformation across the Islands is not gradual, but stepwise. Also, although *in general* the metamorphic grade decreases westwards, within the eastern zone the metamorphic grade increases towards the centre of East Falkland. Furthermore, the Jason Islands have the highest grade of metamorphism outside the Cape Meredith Complex (Figure 4.11, Section 4.3). Except in the Jason Islands, D1 deformation is seen to cease to the south at a poorly defined 'deformation front': D1 folds are not found in the south of the Islands (Figure 4.1). However, for convenience some comments on the structure of the areas south of the D1 deformation front are included in Sections 4.2.1.2, 4.2.1.3 and 4.2.1.7.

4.2.1.1 Eastern zone of D1 deformation

The main zone of D1 deformation, lying east of the Goose Green Axis, forms the landscape of the east and central part of East Falkland, including the Malo Hills [UC 9 7] and the Wickham Heights from Mount Usborne eastwards. The structural trend is predominantly east-west, although deflected to WNW-ESE in the west of the zone.

Folding is pervasive. Fold closures generally range from open to tight, but very tight to isoclinal folds occur locally. Fold symmetry varies from typically upright in the north (as around Port Salvador) to weakly asymmetrical, southerly-facing in the Wickham Heights and near Fitzroy and Port Pleasant (Curtis and Hyam, 1998). Fold axes are subhorizontal or gently plunging to either east or west. In their studies of parts of eastern East Falkland, Curtis and Hyam (1998) found mostly easterly plunges of up to 10°, but the 'pod-shaped' outcrop patterns apparent on the geological maps indicate that the predominant plunge direction varies.

The largest folds have wavelengths of several kilometres (Figure 4.1), but folding occurs on all scales down to a few metres from peak to trough. The intensity of folding, expressed both as frequency and tightness of closure, varies across the area. In part this reflects the massiveness and competence of the rock strata, so for example the Fitzroy Tillite and the sandstones of the Port Stephens Formation are less intensely folded than the softer, well-bedded strata of the Fox Bay Formation in the southern part of Port Salvador [VC 1 7]. However, variation also occurs independently of lithofacies, both along strike and across strike. For example, folds in the Fox Bay Formation around Port Salvador appear to tighten as they pass west of Paso Grande Creek [VC 13 78] and Rumford Creek [VC 12 75], so that as seen on aerial photographs the fold noses become more elongate and closures less easily visible. This is presumed to be a consequence of the westwards constriction of the Fox Bay Formation outcrop in the upper Arroyo Malo area [UC 9 7]. Irrespective of rock type the fold intensity seems to be greatest in a zone between Fitzroy and Port Harriet, presumably passing along strike west into the Wickham Heights. Many of the tightest folds are seen in exposures of the Port Stanley Formation in this zone, for example at Bluff Cove [VC 20 65] and in Vantan Arroyo [VC 12 67].

As noted by Baker (1924), Adie (1952a), Greenway (1972) and Curtis and Hyam (1998) the intensity of folding decreases south of the Wickham Heights and D1 folding is absent from most of Lafonia (Figure 4.1). This southern limit of folding is analogous to the 'deformation front' seen between the Cape Fold Belt and the Karoo Basin of South Africa (Halbich, 1992). However, the exact nature of the deformation front and its position in eastern Lafonia are not well-defined, in part because of the generally very poor exposure in the interior. Baker (1924) concluded that in eastern Lafonia the deformation front lies between Choiseul Sound and Adventure Sound. No evidence of folding is seen on aerial photographs of the area south of Choiseul Sound, but this does not preclude the presence of gentle, large amplitude structures. However, although Macdonald et al. (1996) record minor folding in the north of Lively Island, no northerly dipping strata, or other evidence for D1 folding, have been recorded on the ground between Lively Island and the Goose Green Axis. As described further below (Section 4.2.1.4), strata of the Bay of Harbours

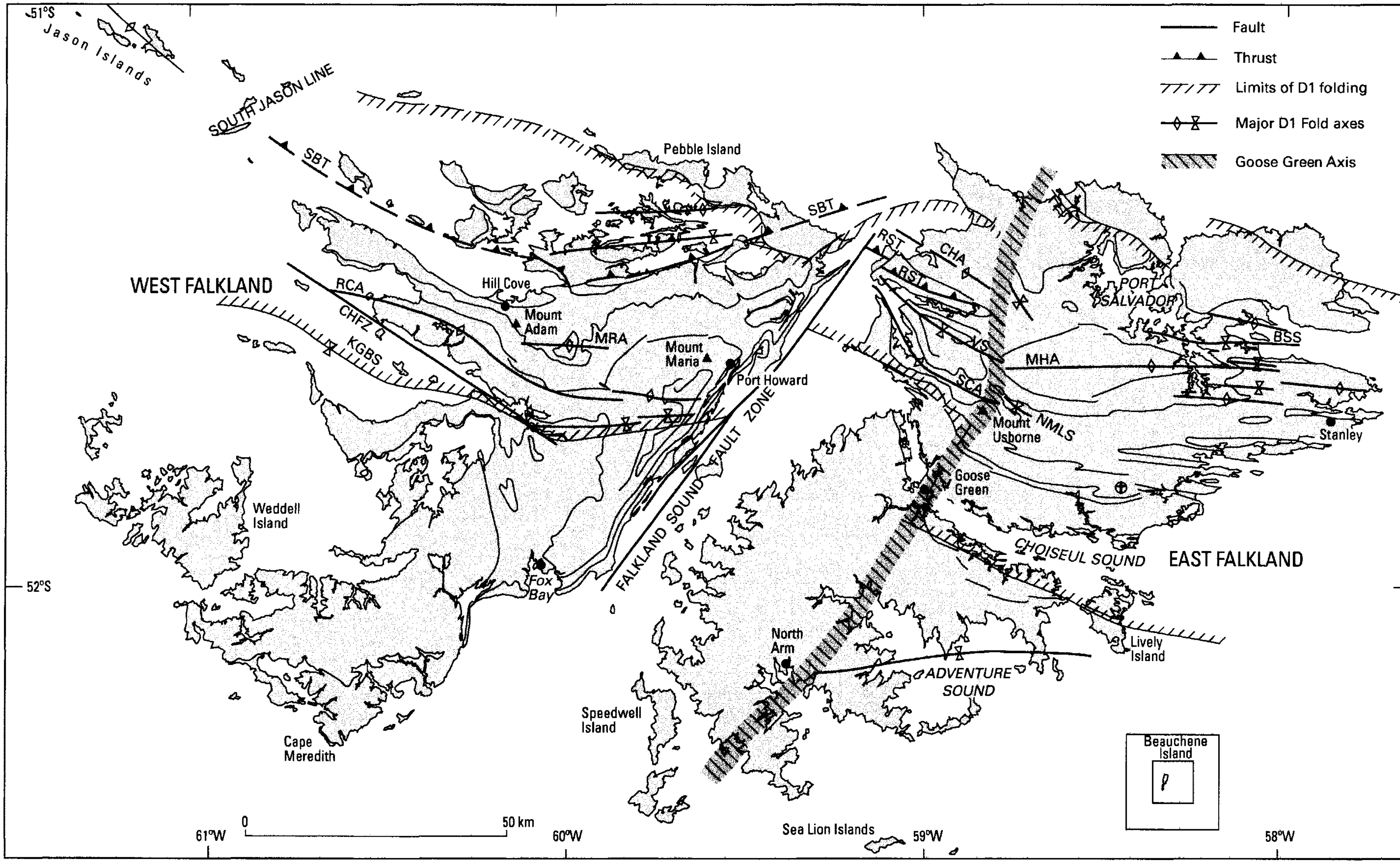


Figure 4.1 D1 structural elements

BSS	Berkeley Sound Syncline	MHA	Malo Hills Anticline	RST	Rookery Sands Thrust
CHA	Coutts Hill Anticline (approximate)	MRA	Mount Robinson Anticline	SCA	San Carlos Anticline
CHFZ	Christmas Harbour Fault Zone (D3)	NMLS	No Man's Land Syncline	SBT	Sound Bridge Thrust (inferred trace of hidden thrust)
KGBS	King George Bay Syncline	RCA	Roy Cove Anticline	VS	Verde Sycline

Formation near Bodie Creek Bridge [UC 61 53] have been disrupted by several isolated folds and small thrusts here attributed to D1.

Fabrics (S1) formed parallel to the axial planes of D1 folds also vary considerably in intensity across the area. For example, the Port Louis Careenage [VC 22 90] lies in a zone of strong east-west trending, north-dipping fracture cleavage in siltstones and fine sandstones, in places almost completely obscuring bedding in the Fox Bay Formation. Fossil brachiopods are distinctly deformed within S1 (Plate 10b). Although an S1 fabric also occurs in the Fox Bay Formation in Chabot Creek [VC 26 94] and in the Johnsons Harbour area it diminishes in intensity in the intervening ground (for example in Fish Creek [VC 23 92]). Also, it is very faint and inconspicuous in exposures of similar rock-types at the west end of Berkeley Sound [VC 19 88] and in Monty Dean's Creek [VC 20 86].

The character of the S1 fabric also varies according to lithology, both in the Fox Bay Formation and elsewhere in the sequence, as described by Curtis and Hyam (1998). The Stanley Quartzites display only a weak pressure solution cleavage localised within fold hinges, although a well-spaced fracture set subparallel to the axial plane of minor folds is seen in places. The Fitzroy Tillite Formation and Black Rock Member display a conspicuous disjunctive cleavage in most places in the east of East Falkland. Curtis and Hyam (1998) also note that competent elongate clasts within the tillite have been aligned and extended parallel to the bulk stretching direction, arguing that this took place under the influence of tectonic stress before the tillite was fully lithified. They observe that a localised onlapping unconformity between the Port Sussex Formation and the Fitzroy Tillite Formation on the northern flank of the Port Pleasant Anticline (Section 2.4.3) also suggests the early onset of regional deformation.

The dip of the S1 fabric varies between 45° to the north and 75° to the south, largely reflecting local variations in fold attitude, but possibly also regional tilting during later deformation. In the north of the eastern zone of D1 deformation the regional attitude of cleavage is symmetrical about the vertical, but south of the Wickham Heights its average regional dip is to the north, agreeing with southerly facing direction of some of the folds, and with the southwards sense of early thrusting (Curtis and Hyam, 1998).

An S1 fabric can be found beyond the apparent southern limit of folding in both eastern and western East Falkland, although it is very weak and sporadic. For example, a weak fracture set parallel to the regional trend of S1 occurs locally in fine sandstones along the eastern shore of Grantham Sound [UC 6 7], although no fabric is seen in silts and mudstones exposed nearby. This localised fabric development seems to be typical of the region close to the deformation front, but localised cleavage has been noted as far south as the northern tip of Bleaker Island [UC 7 2].

While steeply dipping faults lying oblique to bedding are not uncommon in eastern East Falkland, and can be readily detected where they cause lateral offsets in bedding, faults lying parallel to bedding traces (strike-parallel faults) are also important structural elements, probably more so than those lying across strike. While strike-parallel faults are more difficult to demonstrate than strike-oblique faults, either in the field or on aerial photographs, there are several lines of evidence for their occurrence in the eastern zone of D1 deformation. Curtis and Hyam (1998) observed exposures of both reverse and thrust faults with a mean north-south slip direction lying parallel to regional strike. They found that thrusting occurred before and synchronous with D1 folding, citing examples of overturned thrusts in the Port Salvador region, and anticlinal crests breached by small thrust faults around Tumbledown Mountain [VC 35 72]. The present survey confirmed these findings, observing a small thrust duplex overturned during D1 folding at the western end of Ponys Pass Quarry [VC 340 697], for example.

Although exposures of Stanley Quartzite are very common throughout the Wickham Heights, the present survey has shown that these all represent a single interval of resistant strata within the Port Stanley Formation, which has been repeated by folding and faulting. For example, this can be demonstrated in the area between Strike Off Point [VC 33 84], the upper part of Turner's Stream [VC 34 79] and Twelve O'Clock Mountain [VC 39 80], where the interval of resistant quartzite is repeated by an east-west fault lying between an anticline and a syncline. Only one ridge of Stanley Quartzite is seen between that syncline and the base of the Port Stanley Formation near Strike Off Point to the north, whereas two such ridges occur between the syncline and the anticline (which also exposes the base of the Port Stanley Formation), one either side of the fault. The repetition of north-dipping strata by this fault implies relative downthrow to the south, but in this weakly exposed ground there is no good evidence for the direction and amount of dip of the fault plane.

Larger-scale evidence for strike-parallel faults is seen near Port Harriet and Port Fitzroy. For example, an east-west fault cuts out the Bluff Cove Formation outcrop west of Port Harriet [VC 27 68], juxtaposing overturned Stanley Quartzite against Fitzroy Tillite. Superficial cover at the foot of the Wickham Heights obscures the line of faulting but it seems to truncate a series of folds between Vantan Arroyo [VC 12 67] (which exposes three fold pairs), Bluff Cove [VC 20 67] (two pairs) and Port Harriet (only one).

Unfortunately, in much of East Falkland and especially in the more hilly areas, the extent to which faults can be mapped with any confidence, either on aerial photographs or on the ground, is severely restricted by the lack of exposure and the extent of the superficial deposits. In photogeological interpretation of the Port Stanley Formation outcrop in much of the Wickham Heights (especially west of Mount Kent) it is possible to determine the bedding dip direction and to see fold axes only sporadically. Anticlinal cores tend to be well-exposed over a few hundred metres to a couple of kilometres at most, and generally only on the higher ground. The intervening synclines are very rarely exposed and it is possible that most of them have been replaced by strike-parallel faulting. In some places lateral displacement of fold axes by faulting can be demonstrated, but usually it is difficult to know if an apparent displacement represents an offset due to faulting or a *en échelon* offset of the fold axis itself. However, the lack of lateral continuity of fold traces and of topographic features in general suggests that many more faults exist in the Wickham Heights than are shown on the maps, both parallel and oblique to strike. Similarly, some folds in the area of Port Salvador are difficult to trace along strike on aerial photographs, even though bedding is normally clear in outcrops of the Fox Bay Formation. In some instances this may be due to strike-parallel faulting. Interpretation of aerial photographs and satellite images shows that the Brenton Loch Formation has been significantly deformed throughout its outcrop north of Choiseul Sound, between Pleasant Roads and the Goose Green Axis, including medium-scale folding, strike-oblique and strike-parallel faulting. Only those faults and folds which can be identified with reasonable confidence are shown on the map: there are many other possible instances of faulting, either parallel or oblique to bedding.

Similarities in structural style with the Cape Fold Belt of South Africa suggests that at least one large-scale thrust of the type found there might also be present in the D1 fold belt in the east of East Falkland. If such a thrust is present it is most likely to be within the Port Stanley Formation outcrop of the Wickham Heights. Indeed, the presence of a zone of tight folding in the competent strata of the Port Stanley Formation near the southern margin of the Wickham Heights could be explained as deformation associated with a large-scale D1 thrust, possibly blind, as found in the Cape Fold Belt (e.g. Booth, 1998). In contrast, the continuity of outcrop patterns and bedding traces in the Lafonia Group from Port Fitzroy southwards and westwards indicates that no large thrusts are exposed at the surface in that area, although Curtis and Hyam (1998) note possible evidence near Whale Point [VC 14 52] for a blind thrust tip at depth in the Brenton Loch Formation.

Folds are only very faintly visible on aerial photographs within the Fitzroy Tillite, presumably outlined by intervals of thinly bedded strata, or by bedding-parallel joints or low-angle shears. Strong folding is conspicuous at some localities in the Black Rock Member, for example where it is exposed on the east side of L'Antioja Stream [UC 927 574]. However, the north-verging fold pair (30 metres from peak to trough) in the Black Rock Member at this locality reflects its proximity to the outcrop-scale fold pair just to the east, not to some regional structure such as a decollement horizon associated with D1 thrusting. Where folding is seen in most other exposures of the Black Rock Member, as on the Swan Inlet River [UC 84 58], it is similar to the gentle to open, upright north-vergent fold pairs in the Brenton Loch Formation, as exposed in Swan Inlet [UC 92 55], and as seen on aerial photographs in the adjacent outcrops of the Fitzroy Tillite Formation. Photogeology also shows that most of the outcrop of the Black Rock Member has been only locally disrupted by faulting or folding: no evidence is seen for truncation by thrusting.

Similarly, photogeological interpretation of the Brenton Loch Formation shows no evidence for the presence of a major tectonic structure within Choiseul Sound, which is, overall, concordant with regional strike. The apparently gradual southwards decrease in folding and cleavage development, and the lack of any evidence for laterally-extensive strike-parallel faults in the Lafonia Group outcrop, indicates that the D1 deformation front is not controlled by a frontal thrust at or near surface.

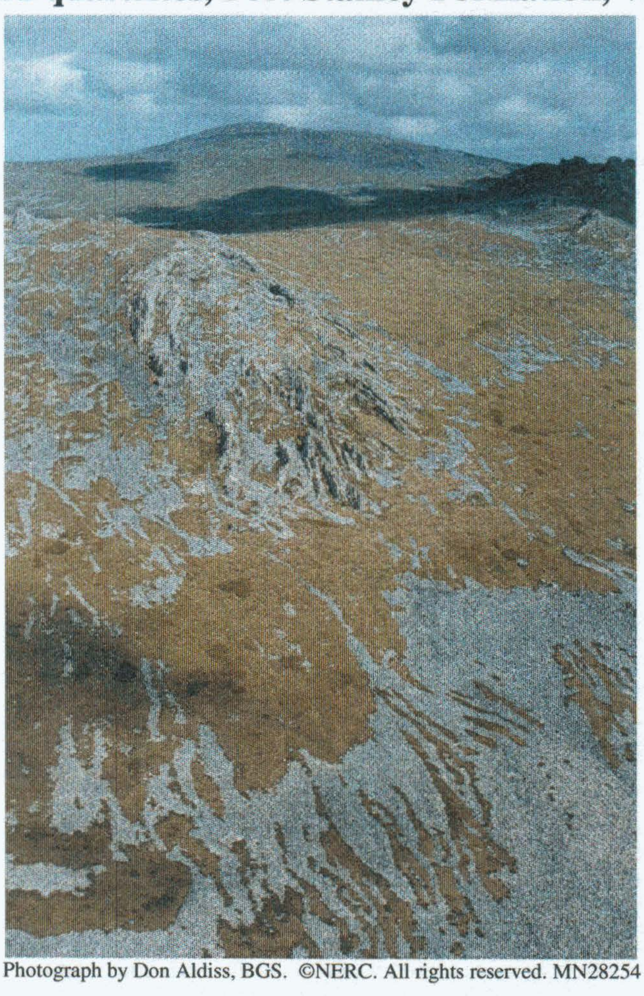
4.2.1.2 South of the D1 deformation front

Although the south-eastern part of Lafonia and adjacent islands lie outside the belt of D1 deformation, it is convenient to comment on their structure here.

Interpretation of remote images shows that Lafonia can be divided into two distinct structural domains by the Goose Green Axis, a NNE-trending lineament aligned with the Goose Green isthmus (Figure 4.1). South of the D1 deformation front and east of the Goose Green Axis, the Permian strata of Lafonia lie in an extremely gentle east-west trending syncline, defined by a zone of apparently horizontal strata either side of Adventure Sound and Low Bay [UC 6 2]. Field observations and photogeological interpretation show that the regional dip in the Driftwood

Plate 26: Folded sandstones, Port Philomel Formation, Chartres

Plate 27: Folded quartzites, Port Stanley Formation, Wickham Heights



Point Peninsula [UC 5 1] is uniform northerly at about 2°, although minor perturbations near fault zones can cause marked variations in local dip directions. To the west, the syncline is truncated near Swan Pond [UC 43 20] by a fault zone aligned with the Goose Green Axis. Western Lafonia is dominated by later deformation.

The general form of the islands in the Sea Lion group suggests the regional dip is to the north-north-west and north, with a series of offsets on northerly-trending faults, each down-thrown to the east. Field observations at the western end of Sea Lion Island indicate a regional dip of 8° to 10° to the north-west. Thus is it quite possible that the beds exposed on Sea Lion Island correlate with strata north of Adventure Sound in the 'transitional zone' between typical Brenton Loch Formation and typical Bay of Harbours Formation (Section 2.4.5). There is no requirement from present evidence for the occurrence of a major structure such as a thrust between the Sea Lion Islands and East Falkland.

4.2.1.3 Structure of Beauchêne Island

In most of Beauchêne Island [UB 3 5] the bedding is subhorizontal, although with a local northerly dip of up to 6° in the central part. Joints are generally orientated WNW-ESE or WSW-ENE, and are vertical. Joint spacing varies between about 10 centimetres and 10 metres. Locally small fibrous quartz veins lie parallel to either joint set. Some joints show a very slight dextral displacement. Two opposed normal faults, striking 330° and 310°, form a small graben in the south of the Island.

Satellite gravity data (Sandwell and Smith, 1997) shows that Beauchêne Island coincides with a linear positive gravity anomaly, suggesting that the island lies in an uplifted basement block elongated ENE-WSW. No evidence for thrust faulting was observed on Beauchêne, and there is no suggestion that this block has been influenced by northerly-verging thrust faults which occur in a narrow zone associated with the Falkland Thrust, fifty kilometres to the south in the South Falkland Basin (Richards et al., 1996b) (Figure 6.1). There is also no evidence that this block is a displaced portion of the Gondwanide fold belt.

4.2.1.4 Central zone of D1 deformation

The central zone of D1 deformation, lying between the Goose Green Axis and the Falkland Sound Fault Zone, forms the hills between Port Sussex [VC 6 7] and Foul Bay [UD 6 0]. The structural trend is predominantly NW-SE, although there are some local deviations from this.

Towards the western end of the Wickham Heights, between Swan Inlet House [UC 88 56] and Black Rock House [UC 81 59], the regional trend of D1 structures changes to a WNW direction as a consequence of later deformation. Seen in detail, it appears that this change takes place incrementally, not gradually. This is clearest in the Port Stanley Formation outcrop where the structural trend within each fault-bounded segment tends to be uniform but slightly different to that of its neighbour. The greatest changes in strike occur at the fault zones marking the Goose Green Axis. The deflection in regional strike is accompanied by a change in the style of D1 deformation, an abrupt northward shift of the D1 deformation front (as indicated by the extent of D1 folding), and a considerable diminution of the intensity of S1 fabrics and in metamorphic grade (Section 4.3). Evidence discussed below shows that tectonic activity occurred on the Goose Green Axis during D1, D3 and D5. Facies variation in the Bay of Harbours Formation suggests that it also influenced Permian sedimentation (Section 2.4.5).

In the area between the Goose Green Axis and the Falkland Sound Fault, D1 structures vary in trend between WNW-ESE and NNW-SSE, forming a series of large-scale kinks between Mount Usborne [UC 74 71] and Wreck Point [UC 54 92]. The multiplicity of folds in the Wickham Heights is replaced by a few large structures: the San Carlos Anticline, the Verde Syncline, the Rookery Sands Thrust and the poorly delineated Coutts Hill Anticline (Figure 4.1). The Verde Syncline and the Rookery Sands Thrust are separated by a markedly linear zone of vertical to overturned strata around Port San Carlos which is effectively the common limb of a south-verging fold pair in which the anticline has been replaced by a thrust. Otherwise, the major folds are open and upright.

Second and lower order folds do occur but they are much fewer than to the east. This simplicity in fold style is clearly seen in the Port Stanley Formation of the Sussex Mountains where no D1 folds occur south of the axis of the San Carlos Anticline, except for the single fold pair forming North West Rincon [UC 50 82]. Further north, minor D1 folds are common in the Fox Bay Formation seen near Queens Brook [UC 85 95] and westwards along the road as far as the New House Fault Zone, but are not seen in the numerous borrow pits and natural exposures in the same formation between the San Carlos River [UC 73 91] and Elephant Beach Pond [UD 77 04], west of the New House Fault Zone.

On remote images there is a very clear contrast in the deformation of the Brenton Loch Formation to the east of the Goose Green Axis, where it is pervasively folded and faulted, and to the north-west, where there is no D1 folding and faulting is less. Sporadic gentle folds, a few metres from peak to trough, are seen within the Goose Green Graben (Figure 4.2, 4.10) at Camilla Creek [UC 649 618]. Strata of the Bay of Harbours Formation near Bodie Creek Bridge [UC 61 53] have been disrupted by several isolated folds and small thrusts. A south-verging asymmetric open anticline occurs some 50 metres south of the bridge. Small scale thrusting and some disharmonic folding have occurred within the anticlinal hinge, whose axis plunges at 20° to N115°, with a steeply north-dipping axial plane. This fold is likely to have formed above a blind reverse fault, but its orientation is consistent with D1 deformation. Although, as expressed by D1 folding, the deformation front is not precisely defined, it appears to have been offset by at least 15 kilometres at the west side of the Goose Green Graben (Figure 4.1).

Partitioning of strain across the Goose Green Graben is also expressed in the S1 fabrics. For example, at Old House Rocks (Ceritos) [UC 700 654], just east of the graben, the tillite and the Black Rock Member carry a pervasive slaty cleavage. A subparallel fracture set occurs in sandstones of the Brenton Loch Formation. Cleavage occurs in places within the northern end of the graben. Immediately west of the graben, near Laguna Babas [UC 69 65], these fabrics are distinctly weaker and more sporadically developed. This is consistent with the generally weak development of S1 fabric in the region west of the Goose Green Graben. No cleavage or subparallel fractures are visible in most exposures around Port Sussex, although a weak cleavage is locally present in the tillite immediately beneath the Port Sussex Formation near Hells Kitchen [UC 62 74] and Port Sussex House [UC 60 75]. Although the Black Rock Formation is folded and very locally tectonically disrupted close to the line of road at the head of Port Sussex, it otherwise displays little evidence of deformation in this area. It does not display the S1 regional cleavage, contrary to the assertion of Curtis and Hyam (1998, p.120). In the coastal sections between Port Sussex and Brenton Loch, the S1 fabric occurs sporadically as a set of narrowly spaced fractures in the fine-grained sandstones of the Brenton Loch Formation. No fabric is observed in the finer-grained lithologies. The local absence of cleavage in this area is attributed to variations in competence of the various rock-types, probably together with local variations in strain.

Deviations in the orientation of S1 in the area between the Goose Green Graben and Port Sussex are attributed to D5 deformation (Section 4.2.5.1). However, it is noteworthy that on the eastern side of the Goose Green Graben near Old House Rocks [UC 701 658], where the Black Rock mudstones change strike by some 50° across a NE-SW fault, the strike of S1 cleavage in the same exposures changes by only 10°. The earliest deformation on the Goose Green Axis thus pre-dated the formation of the cleavage, although probably only dating from an earlier phase of D1.

The difference in fold style and in intensity of S1 fabrics suggests that deformation in the west of East Falkland took place at a higher structural level than to the east, an inference supported by the lower metamorphic grade indicated by clay mineral maturity data (Section 4.3).

The San Carlos Anticline

As shown by Greenway (1972) and described by Curtis and Hyam (1998), the major structure in San Carlos Water is a southerly-verging open anticline. At San Carlos, the main fold axis plunges at about 8° to the north-west (Curtis and Hyam, 1998), but a subsidiary south-verging fold pair extending from Hospital Point [UC 58 91] to Fanning Head [UC 52 97] plunges south-east at up to 30°. The fold axial trace is sinusoidal, probably as a consequence of later deformation. Except at North West Rincon [UC 50 82], parasitic folds are not developed in the Port Stanley Formation forming the limbs of this Anticline. Minor folds observed in the relatively incompetent strata in its core typically have a peak to trough distance of a few decametres and verge to the south-west. The short limbs of the minor fold pairs are commonly replaced by thrusts. Bedding-parallel and discordant thrusts, and reverse faults of south-south-west displacement have been observed at several locations along the coast between Ajax Bay [UC 56 85] and Fern Valley Creek [UC 59 85], together with widespread examples of bedding-parallel slip. These are taken to express disharmonic deformation in the less competent strata in the core of the major fold. In contrast to the eastern part of East Falkland, a weak S1 cleavage occurs only very locally in the least competent strata, in the core of the anticline (Curtis and Hyam, 1998).

At Wreck Point [UC 53 91] the Port Stanley Formation has been overturned, probably due to convergence with the closed anticline exposed on the west point of Fanning Head. Indeed, much of the axial plane of the Fanning Head Anticline appears to have been dislocated: the south-west facing cliffs west of Fanning Head appear massive, bedding having been obscured by small and medium-scale fractures and recrystallisation. A short way to the north of the Fanning Head Anticline [UC 5150 9675] a 40 metre-wide zone of subvertical beds appears to represent the common limb of a south-verging fold pair in which both fold noses have been replaced by faults. It seems possible that as the San Carlos Anticline converges with the Fanning Head Anticline the former could also become dislocated, perhaps passing laterally into a thrust beneath San Carlos Water.

Verde Syncline and Rookery Sands Thrust

The folds observed in the Rodeo Mountains [UC 71 81] very possibly continue north-west into the hills of the Outer Verde [UC 68 82], but not into the broad, shallow-dipping southern limb of the Verde Syncline. Thus they appear to terminate against a set of NNE-SSW faults parallel to the Goose Green Graben.

The straight northern limb of the Verde Syncline is formed by steeply dipping to overturned strata encompassing a major part of the local stratigraphic sequence. Settlement Rocks [UC 6239 9360] and other exposures of quartzite in the ridge north of Port San Carlos are overturned by as much as 50°, as shown by sparse way-up evidence from cross-bedding. A remarkable fabric formed by *Skolithos* lying at 50° to bedding occurs in plane-bedded sandstones of the Port Stephens Formation at Settlement Rocks, although dying out over only 30 metres to the north. The regularity of this fabric and its occurrence within overturned strata suggests that it is a result of ductile deformation during folding, presumably accompanied by bedding plane slip. The sense of rotation of the *Skolithos* tubes is consistent with orthogonal compression within a major fold, that is, the fabric formed before the strata were overturned. There is also later kink-folding which deflects some of the *Skolithos* (Section 4.3).

The ductile style of the deformation at Settlement Rocks contrasts with the faulting and brittle fracture observed only a few kilometres away at Curlew Creek [UC 66 88], but which also seem to reflect deformation within the common limb of this fold. The most spectacular examples of brittle D1 deformation occur in the craggy pinnacles of The Picos [UC 62 97], which mark the highly fractured to brecciated axial zone of an anticline. Similar pinnacles occur at Fanning Head, on the trace of the Fanning Head Anticline, and the same style of brittle fracture and recrystallisation occurs in the hills east of Smylies Creek [UC 65 97].

The Verde Syncline is exposed on the coast about one kilometre south-west of Rookery Sands [5275 9930] marked by an abrupt change in the orientation of the beds. No subsidiary folding was seen near the hinge. To the south, some three kilometres of coastline expose mainly uniformly gently dipping strata, with only a few minor folds and small high-angle faults seen in the cliffs. A gully meeting the coast at [5210 9895] exposes a small thrust, dipping at about 35° to the north, in which beds in the hanging-wall have been dragged into a south-west verging anticline. An antithetic north-easterly verging fold pair occurs some 300 metres to the north-east. This has a gently south-westerly dipping axial plane, partly replaced by a small south-west dipping thrust.

To the north of the Verde Syncline, as far as Rookery Sands [UC 53 99], the beds are uniformly steeply dipping and overturned by up to 64° in places. To the north of Rookery Sands this relatively simple structure is replaced by a zone of complexity interpreted as a rotated thrust duplex. An apparently undeformed thrust plane (dipping NNE at 18°) which crops out in the cliff behind Pebble Beach [UD 53 00] is taken to be the major plane of movement. Strata exposed on the north side of Pebble Beach are subvertical and intensely deformed. Some tight fold closures are visible but there is much dislocation subparallel to bedding planes as well. Some second-phase recumbent folds are apparent, perhaps due to local gravitational collapse. These tight folds die out northwards over one to two kilometres, becoming progressively less disharmonic and more open, to gentle. At least two more thrusts are exposed in the cliffs, one of which replaces a syncline in a fold pair. Some 700 metres north-north-east of Pebble Beach [5380 0130], the cliffs expose a south-west verging fold pair, together with a pop-up fold with a gently south-westerly dipping axial plane about 50 metres to the north. Geometrically similar structures occur elsewhere in the section.

The rather angular fold style, with minor pop-up folds and disharmonic accommodation folds, and the brittle style of deformation, show that in this area D1 deformation was occurring at a relatively shallow level. This is consistent with the low grade of metamorphism indicated by clay mineral maturity data (Section 4.3).

The eastwards (inland) continuation of the Rookery Sands Thrust is obscured by lack of exposure but it seems most likely that it is offset to the south (together with the trace of the Verde Syncline), passing along the limit of the inverted strata north of Findlay Rocks [UC 59 94] and Settlement Rocks [UC 62 93] (Figure 4.1). However, there is no evidence that it extends beyond the eastern end of the Cerro Montevideo ridge [UC 70 92].

Coutts Hill Anticline

In the Smylie Rocks [UD 65 00] and Coutts Hill [UC 68 99] area the Port Stephens Formation is very poorly exposed and except in the north photogeological interpretation yields little structural information. The sparse indications of dip direction, together with the outcrop pattern of the base of the Fox Bay Formation to the north and east, show that it is disposed in a broad easterly plunging anticline. The Fish Creek Member probably continues around the edge of this structure to underlie the low ground near Moss Side [UC 67 95] (as shown on Section H-J in the margin of the East Sheet of the geological map) but it could not be delineated consistently and so is not shown on the face of the map. A zone of D5 high-angle faults separates the Coutts Hill Anticline from the overthrust strata north of the Rookery Sands Thrust (cf. Figures 4.1 and 4.10). It is likely that faulting in the Coutts Hill area is more complex than shown on the map, and it is not possible to plot the inferred anticlinal trace with any confidence.

4.2.1.5 Western zone of D1 deformation

The western zone of D1 deformation, lying between the Falkland Sound Fault Zone and the South Jason Line, forms most of the northern half of West Falkland and the adjoining islands to the north. The structural trend is east-west in the eastern part of that zone, swinging to WNW-ESE in the west.

The style of D1 deformation in West Falkland differs from that seen in eastern East Falkland but is similar to that west of the Goose Green Axis. A south-vergent pair of upright folds, the Roy Cove Anticline and the King George Bay Syncline, can be traced from the centre of the Hornby Mountains almost to the west coast. The chain of islands within King George Bay forms part of the common limb of the fold pair. In the Roy Cove area [TC 6 8] the fold axes multiply and are offset *en échelon*. The only planar fabrics are weak and found very locally in the axial zones of these folds. In spite of this difference in fold style, the east-west to WNW-ESE orientation of the folds, their lateral persistence and their alignment with D1 folding in eastern East Falkland suggest that they also represent D1, and so that they formed at about the same time. Variations in trend and plunge are attributed to D2 folding and probable D3 faulting.

The major folds are gentle to open with a peak-to-trough distance of some four to five kilometres in the east, decreasing to one or two kilometres near Roy Cove. Small and medium-scale folds are confined to the outcrops of the Fox Bay Formation and Port Philomel Formation, and are not seen in the more competent strata of the Port Stephens Formation. Luxton (1994) reports that north-easterly-vergent folds in the Chartres area trend about N150°, locally with weak axial planar cleavage. Although at Teal River the major anticline plunges east, minor fold plunges up to about 20° to both north-west and south-east have been noted in the Chartres area. Variation in plunge is attributed to D2 folding.

Localised zones of folded strata extend as far as the southern shore of Christmas Harbour, for example south-west of Chartres [TC 872 655], and along the coast westwards to near Gun Hill Shanty [TC 820 684]. Folding at Town Point [TC 76 72] is visible on aerial photographs. Foreshore and cliff exposures show open to tight fold pairs, generally in the range two metres to 15 metres from peak to trough, although decimetre-scale folding occurs in the mudstones. The folds are rather angular, commonly with broken or faulted fold axes, sheared common limbs, and no cleavage, demonstrating brittle deformation. The position of the faulted trace of the King George Bay Syncline can be inferred from local changes in the vergence of mesoscale D1 folding across faults exposed on this southern shore [TC 8633 6670]. Cliffs on the east side of the inlet opposite Chartres [TC 872 655] expose eight north-verging, gently plunging fold pairs, most of which have been disrupted by shearing.

Open south-verging D1 fold pairs (for example, about five metres peak to trough, with an easterly axial plunge of 28°) occur on the northern shore of Port North [TC635 924], in association with small-scale thrusts (typically dipping 23° to the NNE). The thrusts are cross-cut by northerly-trending high-angle faults. No cleavage occurs near Port North (Marshall, 1994b).

Most of the inlier of the Port Philomel Formation north of Mt Robinson [TC 9 7] appears to coincide with the axial zone of a large, very gentle anticline of a south-vergent fold pair (Figure 4.1). The corresponding syncline lies just south of Mount Robinson. This pair could be an *en échelon* continuation of the folds near Roy Cove. However, although the Mount Robinson Anticline approximately aligns with the Many Branch Harbour Anticline (Figure 4.2), there is no evidence for folding in the intervening ground, even in the relatively incompetent strata of the Fox Bay Formation (Section 4.2.2.1).

Deformation in the northernmost part of West Falkland and in the adjacent islands (except Pebble Island) is also thought to be part of D1. It is argued in Sections 4.2.4 and 4.2.5 that the high-angle faulting seen at the east end of Port Purvis (which during D5 juxtaposed sandstones low down in the Port Stephens Formation with the Fitzroy Tillite Formation) was preceded by large-scale thrusting during D4. A very similar structural relationship is found at the eastern end of Byron Sound and it seems that the high angle faulting in this area was likewise preceded by a structural inversion due to thrusting, but in this case, during D1. Some low angle faults lying parallel to strike possibly occur within the Port Stephens Formation outcrop between East Lagoon Hill [TC 92 95] and View Hill [TC 97 95]. In the absence of such thrusting, a vertical throw in excess of two kilometres must have occurred on both the east-west and the NW-SE lines of faulting near Sound Bridge [TC 94 92]. The truncated thrust inferred to occur beneath the Port Stephens Formation outcrop north-east of Sound Bridge is here named the Sound Bridge Thrust. Although the Sound Bridge Thrust is nowhere exposed, its lateral continuation can be deduced from other structures.

To the west, the apparent throw on the north-westerly trending faults at Sound Bridge decreases markedly until on Saunders Island between Brett Harbour and Burnt Harbour the displacement has occurred entirely within the Fox Bay Formation. The trace of the inferred Sound Bridge Thrust is therefore thought to strike west through the West Lagoons area, most probably on the north side of the fault-bounded blocks of inverted Devonian strata [TC 85 97], and to continue north of Skip Rock [TC 82 96] into Byron Sound. This implies that the chain of islands including Carcass Island and The Twins [TD 4 1] also lies above the postulated thrust. The localised belt of relatively intense folding exposed within this chain and in the south-western part of Saunders Island can be viewed as a continuation of the folding north of the inferred trace of the thrust (in the hanging wall) at West Lagoons [TC 86 97].

The continuation of the Sound Bridge Thrust west of The Twins is obscured by deformation on the South Jason Line. If this is indeed a D2 structure (Section 4.2.2), then the Sound Bridge Thrust is part of D1. While this is the simplest interpretation and is consistent with evidence discussed elsewhere in this section, the possibility remains that the South Jason Line formed later than D2, removing one constraint on the age of the Sound Bridge Thrust.

To the east of Byron Sound, the complex east-west fault zone thought to have displaced the Sound Bridge Thrust at Sound Bridge continues east into River Harbour [UC 1 9]. Elements of this high-angle fault zone apparently pass south of River Island [UD 1 0] and between Shag Rocks [UD 22 00] and Purvis Rincon, although there is no evidence for a significant structural inversion north of Purvis Rincon. In this interpretation the trace of the Sound Bridge Thrust is overridden to the east by the D4 Pebble Island Thrust (Section 4.2.4.1).

While there is no evidence (other than an approximate alignment) that the Sound Bridge Thrust extended east of Falkland Sound, or that the Rookery Sands Thrust extended west of it, the correlation of these two structures seems plausible. This suggests that the vertical to overturned strata around Port San Carlos (the common limb of an overthrust fold pair – Section 4.2.1.4) are structurally equivalent to (A) the south-eastwards dipping strata of Purvis Rincon, (B) the narrow fault-bounded blocks of Port Stanley Formation strata between River Harbour and Sound Bridge (Section 4.2.5), and (C) the overturned Devonian strata at West Lagoons. The correlation of the inferred Sound Bridge Thrust with the Rookery Sands Thrust requires an apparent dextral offset in the thrust trace of at least ten kilometres along the Falkland Sound Fault Zone, similar to that inferred for D3 faulting in Falkland Sound from other evidence (Section 4.2.3.2), and sufficient to remove the Rookery Sands Thrust south of the D4 and D5 faulting which has obscured the Sound Bridge Thrust (Figure 4.1, 4.10).

The orientation of the broad south-verging east-west trending fold pair between River Harbour and Keppel Sound, within the hanging wall of the Sound Bridge Thrust (Figure 4.1), suggests that this Thrust was emplaced towards the south. This fold pair (which dies out near Shallow Bay, close to the inferred trace of the thrust) is similar in style, amplitude and orientation to the D1 fold pair in the Hornby Mountains. Likewise, the narrow fold belt on the northern side of Byron Sound lies parallel to D1 folds around Roy Cove. These correlations strongly suggest that the Sound Bridge Thrust together with the folding to the north (as far as the Pebble Island Thrust; Section 4.2.4) are part of D1. This is consistent with the presence of east-west trending, north-dipping D1 reverse faults in the north of the Hornby Mountains (Curtis and Hyam, 1998). However, the style of deformation associated with the Sound Bridge Thrust and the Rookery Sands Thrust differs from that found adjacent to the Pebble Island Thrust and its likely correlative, the Sand Grass Thrust (Section 4.2.4).

Richards et al. (1996b) note that the Palaeozoic platform immediately east of the Malvinas Basin (which lies offshore to the west of the Islands) is transected by WNW-ESE-oriented, northwards dipping thrust sheets. These mark the southern margin of a broad negative gravity anomaly to the west and north-west of the Islands, which is taken to be a thick sedimentary sequence, probably a tectonic complex of Permo-Triassic age (Richards et al., 1996b). Its position suggests that thrusts in such a complex are more likely to be continuations of the D1 Sound Bridge Thrust than the D4 Pebble Island Thrust. However, as observed in Section 4.2.4, there is insufficient constraint on the age of these offshore thrusts to preclude assignment to D4.

4.2.1.6 Jason Islands

The Jason Islands lie within the most westerly zone of D1 deformation, west of the South Jason Line (Figure 4.1). The structural trend is NW-SE.

Although the age of deformation in the Jason Islands is uncertain, the style and orientation of folds and fabric are both compatible with an origin during D1. Steeple Jason is part of the south-west limb of a large, upright, south-facing open anticline, with the greater part of Grand Jason forming the north-east limb (Figure 4.1). Stereographic projection analysis of bedding suggests an interlimb angle of about 95°, a steep north-easterly dipping axial plane, and a fold axis plunging at about 5° towards N120°. Estimates of thickness show that approximately the same interval of 1000 metres of strata is present in both limbs of the fold. However, the axial plane of this large anticline appears to have been replaced by a NW-SE trending fault which crosses the south-west extremity of Grand Jason, lying approximately parallel to strike in the western limb of the fold. Outcrop scale folds occur in several parts of Grand Jason. Their orientation is consistent with being parasitic to the main fold, and in some cases the axial plane has also been cut out by faulting.

The sandstones on Grand Jason and Steeple Jason are unusual in the Falklands in displaying a pervasive penetrative planar fabric, which is axial planar to the large anticline passing between the two islands (Plate 28). This fabric is uniform at outcrop scale and appears across the full strike width of both islands. It appears as a slaty cleavage within mudclasts in the sandstones. It does not resemble fabrics of the kinds associated with shear zones or faults. The existence of a regional penetrative fabric on the Jason Islands is consistent with the relatively high grade of metamorphism indicated by clay mineralogy (Section 4.3). Conversely, faulting on the Jasons occurred by brittle fracture, accompanied by formation of cataclasite. No evidence for ductile shearing is seen.

4.2.1.7 Deformation front in West Falkland

The southern extent of D1 deformation observed in West Falkland coincides with the major King George Bay Syncline and its continuation in the southern part of the Hornby Mountains. If extended eastwards, the syncline would meet the inferred line of the Falkland Sound Fault some three to four kilometres north-west of North Swan Island [UC 2 6]. Conversely, if a nominal deformation front lying close to the southern side of Choiseul Sound is extended westwards, ignoring the effects of D2 and D3 deformation in western Lafonia, it would meet the Falkland Sound Fault at about the same place, or within 10 kilometres to the south (Figure 4.1) (compare Curtis and Hyam, 1998, p. 126).

As in southern Lafonia, to the south of the King George Bay Syncline there is a gentle northerly regional dip (generally not exceeding 5°) although this is deflected at intervals by D2 structures. Some steeper dips also occur locally next to faults. For example, sandstones dipping at up to 10° form a natural arch and small peninsula to the south-west of a small west-north-west trending fault in Empire Beach, south of Stephens Peak [TC 3753 1540]. Localised deformation associated with dykes is noted in Section 3.1.1.3.

In the region of gently-dipping resistant strata of the Port Stephens Formation in the south of West Falkland, erosion on joints, dykes (Section 3.1 and 4.4) and minor faults strongly influences landforms, especially on coasts facing away from the direction of dip. Many of the extensive subvertical cliffs (themselves controlled by joints) are penetrated by steep-sided gullies, and stacks and blowholes are fairly common. Locally, intersecting joint sets give rise to notable landforms, such as on the west-facing slopes south of North Bluff (New Island) [TC 055 655], where east-west striking joint sets dipping between 55° to the north and 55° to the south have intersected with each other, and with north-north-westerly joints parallel to the coast, to form a distinctive series of triangular-shaped ridges and gullies (Chater, 1993, p.15).

4.2.2 Second phase of deformation (D2)

The effects of D2 deformation are exemplified by the Hornby Mountains Anticline and the Coast Ridge of eastern West Falkland (Figure 4.2). D2 was also responsible for several other large folds found at intervals throughout the Islands, except in the eastern part of East Falkland. These sporadically developed folds do not conform to a single

Plate 28: Cleaved sandstones, Port Stephens Formation, Grand Jason

Photograph by Don Aldiss, BGS. MN28255

Plate 29: Overturned Stanley Quartzite, the Narrows, Coast Ridge

Photograph by Carol Aldiss.

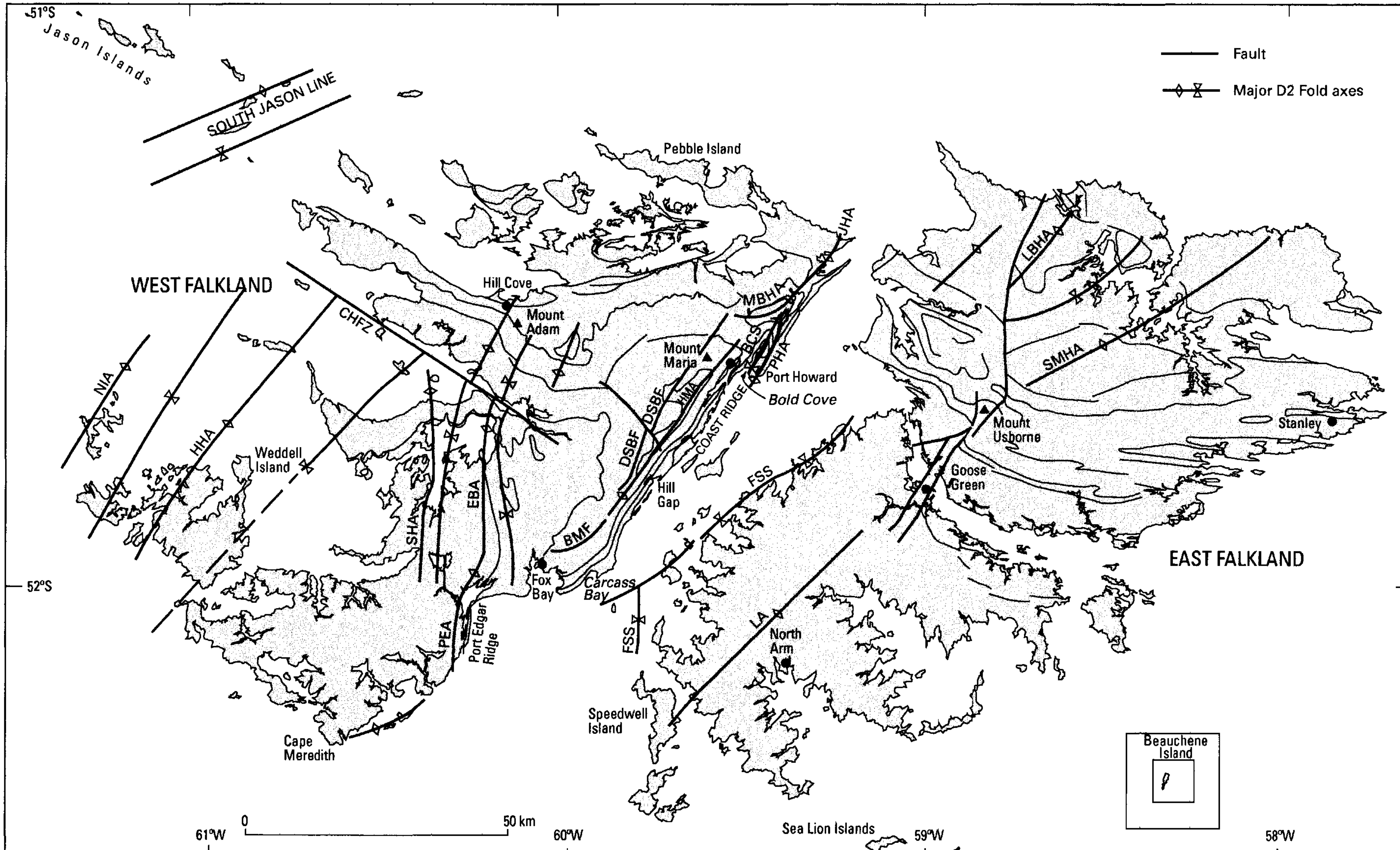


Figure 4.2 D2 structural elements

BCS	Bold Cove Syncline
BMF	Blue Mountain Fault (D5)
CHFZ	Christmas Harbour Fault Zone (D3)
DSBF	Double Stream Basin Fault
EBA	East Bay Anticline
FSS	Falkland Sound Syncline
HHA	Hotham Heights Anticline
HMA	Hornby Mountains Anticline
JHA	Jersey Harbour Anticline
LA	Lafonia Anticline
LBHA	Letterbox Hill Anticline
MBHA	Many Branch Harbour Anticline
NIA	New Island Anticline
PEA	Port Edgar Anticline
PHA	Poke Point Hill Anticline
SHA	Symonds Harbour Anticline
SMHA	Standing Man Hill Anticline

regional orientation, trending between north-south and ENE-WSW. In several cases, the common limbs of the fold pairs are markedly linear and changes in the configuration of individual structures occur rather abruptly. These features are characteristic of drape folds, formed by passive deformation of a sedimentary cover sequence overlying blind faults (e.g. Friedman et al., 1976). Drape folds are individually linear but do not necessarily follow the same trend as their neighbours, reflecting the diverse orientations of basement fractures. The D2 folds are drape folds representing the localised passive response by the Palaeozoic sedimentary cover to east or ESE-directed, up-to-the-west compressional movements between structural blocks in the underlying Proterozoic crystalline basement. Mostly the associated faulting is blind, but it does reach the surface in some places. D2 folds deflect D1 structures in several places.

Curtis and Hyam (1998) deduced that the D2 structures along the east coast of West Falkland arose through a combination of orthogonal shortening across and dextral shear along a NE-SW axis, and inferred a transpressional deformation regime under east-west compressive stress. While this concept explains many aspects of D2 deformation, some D2 structures are themselves deformed by dextral rotation or strike-slip faulting. These movements could be treated as a continuation of the lateral component of D2 deformation, or 'late D2', but it is here considered more convenient to regard them as a D3 deformation. It is assumed that D2 and D3 movements together reflect local accommodation to the stresses generated by large-scale rotation of the Falklands microplate (Section 6). Steeply dipping strata in D2 structures have been intruded by dolerite dykes near East Head (Fox Bay) and south of Port Edgar. Dykes occurring in the area between Port Philomel and Port Edgar also appear to post-date D2 structures.

This section first describes D2 structures in the Coast Ridge area of eastern West Falkland, with particular emphasis on areas of complication around Bold Cove and Carcass Bay. It then deals in turn with D2 structures in the rest of West Falkland, in the Jason Islands, in Falkland Sound and in East Falkland.

4.2.2.1 Coast Ridge area

The east coast of West Falkland between East Head [TC 93 34] and Jersey Point [UC 47 99] is remarkable for an almost continuous ridge of fairly uniform height and width lying parallel to the coast. This, the Coast Ridge, is formed by the Port Stanley Formation (Plate 29). In most of the ridge the strata are near vertical (≥75°) or steeply overturned, but in the south (from midway between Hill Gap and Carcass Bay) and to the north of Bold Cove the dip lessens to between about 40° and 60°. As so defined the central section of the Coast Ridge is part of a 'steep zone' which not only encompasses the younger formations exposed on the eastern side of the ridge but also the older strata to the west, extending down sequence to the Albemarle Member. This steep zone is part of the common limb of a south-eastwards verging D2 monoclinal fold pair, comprising the Hornby Mountains Anticline (HMA) and the largely unexposed Falkland Sound Syncline (FSS) (Section 4.2.2.4).

The HMA intersects the major east-west D1 fold pair in the Hornby Mountains, forming an elongated fold interference structure, but there is no discernible trace of the gentle D1 folds to the east of the HMA axis, where they have presumably been flattened by deformation within the D2 steep zone. To the south of the intersection with the D1 anticline there is a weak culmination in the HMA coincident with a line of D3 faulting north-east of Mount Moody. From there the HMA plunges gently to the south-west. Near the southern end of the steep zone, the HMA passes south-west into a normal fault, the Blue Mountain Fault [UC 03 49]. The sense of offset of the base of the Fish Creek Member indicates that the Blue Mountain Fault is downthrown to the west. This is contrary to the inferred up-to-the-west sense of D2 deformation (discussed further below) but as this fault and others nearby cross-cut and displace dolerite dykes, the westwards downthrow could represent D5 movement (Section 4.2.5.2). To the north of the D1 anticline, the HMA plunges north-east at about 25°, exposing progressively younger strata until it reaches the north end of the steep zone within the outcrop of the Fox Bay Formation, where it is lost in a zone of faulting. No continuation of the HMA into the Many Branch Harbour Anticline can be demonstrated (Figure 4.3).

Curtis and Hyam (1998) record mesoscale dextral faults of both NNE-SSW and NNW-SSE orientations around Port Howard, the former causing bedding parallel slip within the steep zone. It is likely that the whole Coast Ridge has undergone some tectonic thinning. They also note a set of NE-SW low angle thrusts with a north-westerly transport direction concentrated within the Coast Ridge steep zone near Port Howard. These thrusts typically cross-cut anticlinal hinges and are associated with a weak, localised cleavage. They are probably 'out-of-syncline' thrusts formed as the D2 fold tightened.

Between Gorge Stream [UC 11 63] and Double Stream [UC 17 71] the position of the Hornby Mountains Anticline is unclear, although the position of the western edge of the 'steep zone' can be traced on aerial photographs with reasonable confidence. On the south side of Double Stream, between Deep Valley and Stewart's Rock, the change from gentle and moderate westerly dips west of the steep zone to overturned strata within it is seen on the ground to occur within an unexposed area a few hundred metres wide. No gently east-dipping beds (as would be expected close to the fold hinge) are seen there, but several exposures in the adjacent part of the steep zone are intensely fractured, most typically on near-vertical planes striking north to north-east. It seems that the hinge of the Hornby Mountains Anticline has here been replaced by a fault. This is likely to be part of the fault system over which the D2 drape fold formed.

Similarly, the Many Branch Harbour Anticline and the Jersey Harbour Anticline are connected by a reverse fault, south-east of Mount Rosalie [UC 39 94]. The transitions between anticline and fault are both in low, poorly featured ground smothered by superficial deposits.

Photogeological interpretation of the Albemarle Member inlier in the Hornby Mountains shows that a reverse fault, the Double Stream Basin Fault (DSBF), lies parallel to the Hornby Mountains Anticline, a few kilometres to the north-west. Having been offset by the NW-SE D3 fault in The Gorge [UC 11 63], the DSBF repeats the sequence overlying the Albemarle Member about 1.5 kilometres west of Mt Moody. Indeed, a narrow outcrop of the Albemarle Member could occur on the west side of the deep valley north of Mount Moody [UC 08 63]. On the north side of the Hornby Mountains, the Double Stream Basin Fault undergoes another dextral offset and apparently continues to just north of Macdonalds Rocks [UC 23 86]. The dolerite dyke which crops out in Macdonalds Rocks is coincident with this fault, or lies closely parallel to it (Section 3.1).

The formation of the Hornby Mountains Anticline is thought to be a consequence of uplift of West Falkland relative to East Falkland (Greenway, 1972). Macdonald et al. (1996) and Curtis and Hyam (1998) independently proposed that the relative uplift of West Falkland occurred by displacement on a north-westerly dipping blind reverse fault beneath the Hornby Mountains Anticline, not on a fault system in Falkland Sound. Hyam (1997) discusses evidence from seismic surveys and from anomalies in the regional gravity field which indicates that basement faults aligned with the Hornby Mountains Anticline continue offshore to the south-west of the Islands for at least 60 kilometres.

The presence of reverse faults and of overturned strata indicates that some degree of lateral movement of West Falkland towards East Falkland accompanied uplift, at least in the sector corresponding to the steep zone. Curtis and Hyam (1998) calculated that in a section between Shag Cove Mountain [UC 15 68] and Egg Harbour [UC 35 55] 3.9 kilometres or more of orthogonal shortening occurred, with at least 3.3 kilometres of dextral displacement parallel to the HMA. However, in the absence of any information regarding the amount and direction of strike-parallel faulting in the steep limb of the Hornby Mountains fold pair and in Falkland Sound such calculations should be regarded with circumspection (Section 4.2.3).

In marked contrast to the relatively simple structure in most of the Coast Ridge, areas of structural complexity occur at each end of the Hornby Mountains steep zone, around Bold Cove and around Carcass Bay (Figure 4.2).

Bold Cove

On the geological map, one's attention is drawn to the Bold Cove area by the elongate, rather eye-like outcrop of the Port Stephens Formation west of Many Branch Harbour (Figure 4.3). This clearly marks a culmination on the trace of an open, upright anticline, the Many Branch Harbour Anticline (MBHA), and the hook-shape at its eastern end suggests that it is some kind of fold interference structure. However, close inspection reveals that this structure can be interpreted in more than one way. Unfortunately, in West Falkland tectonic fabrics formed during D1 or D2 are extremely localised and there are few minor folds. This presents a serious obstacle to determining which interpretation of the folds in the Bold Cove area is the most plausible.

Greenway's (1972) map shows the MBHA together with the Jersey Harbour Anticline as an *en échelon* continuation of the HMA, together with two large closed folds: the Bold Cove Syncline and the Poke Point Hill Anticline (Figure 4.2, 4.3). The latter both plunge moderately to the south-south-west and are weakly asymmetric, with axial planes dipping steeply westwards. Their similar style suggests they are a west-verging fold pair.

Curtis and Hyam (1998) present a more sophisticated interpretation of the Bold Cove area which takes account of their field studies. They interpret the MBHA as a D1 fold which has been refolded about a vertical D2 fold axis, believing that the Poke Point Hill Anticline is a portion of the MBHA which has been rotated on that fold axis, tightened and extended by the development of a coplanar D2 anticline during D2 dextral transpression (Figure 4.4A).

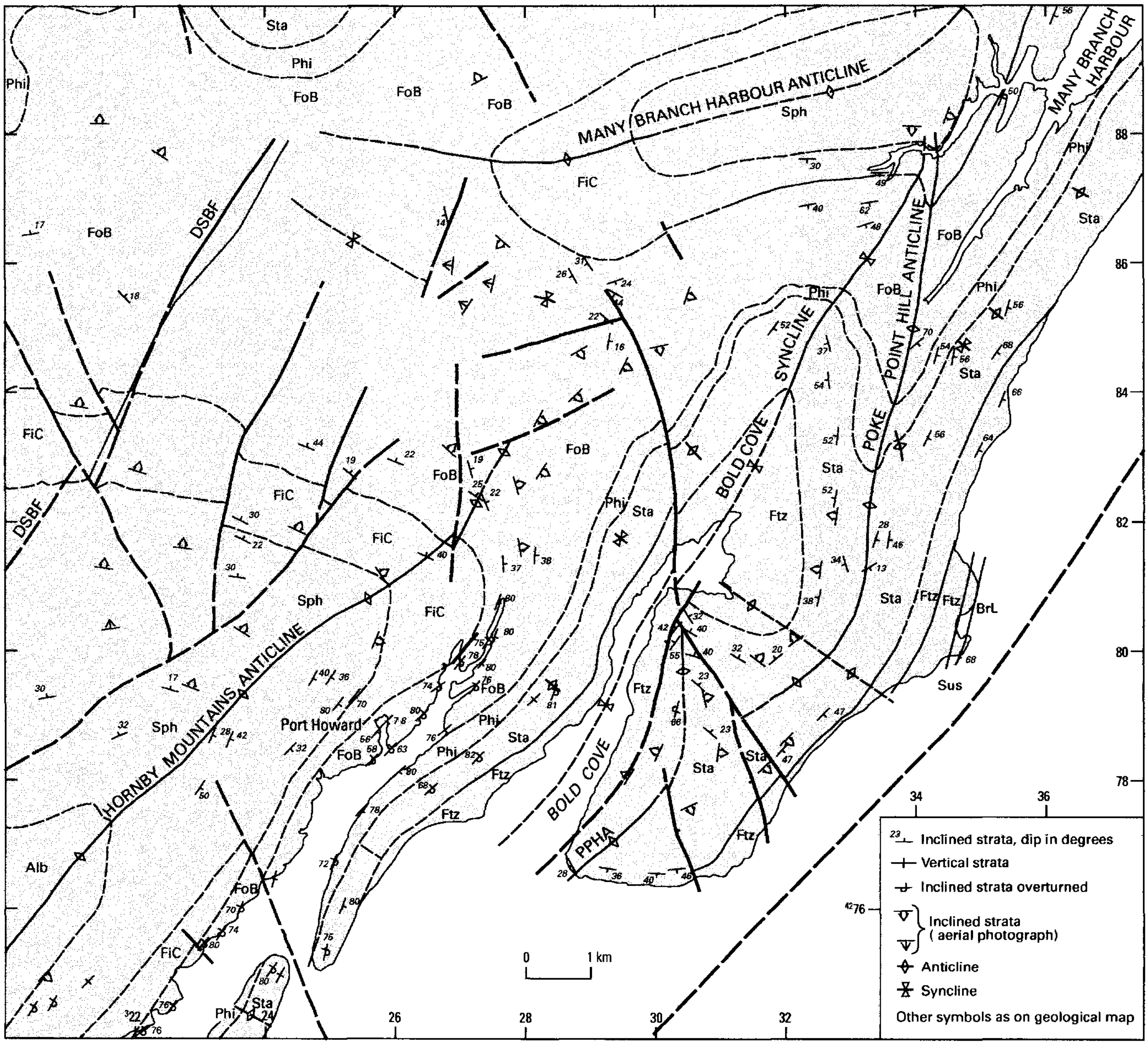


Figure 4.3 Major structures of the Bold Cove area

DSBF	Double Stream Basin Fault
PPHA	Poke Point Hill Anticline

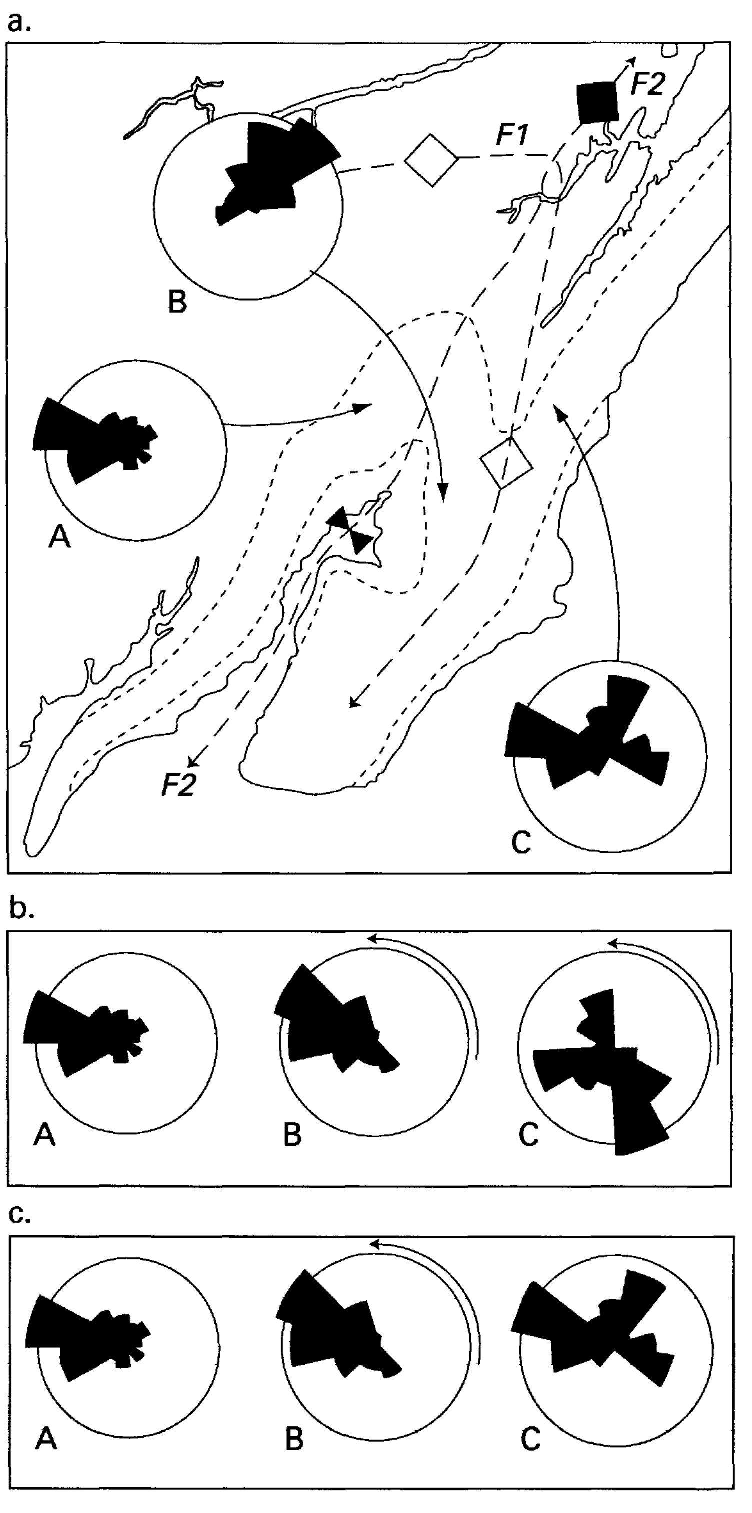


Figure 4.4 Palaeocurrent data for the Port Stanley Formation of the Bold Cove area

After Curtis and Hyam (1998, fig. 11). A: Palaeocurrent data for three major fold limbs without correction for D2 folding.
B: The same data after correction in the Curtis and Hyam model.
C: The same data after correction in the model proposed here.

The Bold Cove Syncline is interpreted by Curtis and Hyam (1998) as a D2 structure which changes into an anticline as it crosses the axis of the MBHA, continuing north-east into Jersey Harbour.

The model presented by Curtis and Hyam (1998) seems not to explain all the structural features of the Bold Cove area, and it is suggested here that the MBHA is actually a D2 structure and that Greenway's (1972) interpretation was essentially correct.

The Port Stephens Formation inlier at Many Branch Harbour is very poorly exposed and reveals very little structure. Thus there is very little direct evidence for the inter-correlation of the major fold axial traces that intersect this inlier (Figures 4.3 and 4.4). For example, although the two NNE-SSW folds at Bold Cove can be traced across the outcrop of the Fish Creek Member [UC 34 87], there is no evidence that they continue more than 250 metres across the western arm of Many Branch Harbour into older strata. The inferred extrapolation of the Bold Cove Syncline into the anticlinal trace to the north-east (Curtis and Hyam, 1998) cannot be observed directly, and would require a double inflection in the fold trace to overcome the slight misalignment.

The outcrop pattern of the Albemarle Member inlier east of the Hornby Mountains clearly shows an interference pattern where the east-west D1 anticlinal axis intersects the D2 HMA, but as noted above no sign of the gentle D1 fold can be recognised in the D2 steep zone. The east-west section of the MBHA anticline has a similar fold style to the D1 anticline in the Hornby Mountains, so if it were also a D1 fold, one would likewise not necessarily expect to see the remnant of the MBHA in the D2 steep zone. Indeed, Curtis and Hyam (1998) find that the poorly defined syncline which seemingly forms a pair with the western end of the MBHA cannot be traced across the northern end of the HMA (Figure 4.3). On the other hand, Curtis and Hyam (1998, p. 126) accept that a D2 anticline exists east of the D2 Bold Cove Syncline. This is here considered a sufficient explanation of the NNE-SSW trending folds at Bold Cove and would explain why the Poke Point Hill Anticline and the Bold Cove Syncline are of very similar style: there is no need to postulate a rotated eastern segment of the MBHA to explain the structure in this area.

In the absence of tectonic fabrics and small-scale tectonic structures Curtis and Hyam (1998) found support for their interpretation only in the local variation of palaeocurrent direction (as deduced from planar cross-bed foresets) across the Bold Cove area. They found that after restoring the measured data to palaeohorizontal, without any correction for their inferred D2 refolding, there was a discrepancy between the data for the three fold limbs (Figure 4.4A). They suggest that if the foreset data for Limbs B and C are then subjected to an anticlockwise rotation about a vertical axis (to mimic the realignment of the Poke Point Hill Anticline with the Many Branch Harbour Anticline in their model), then a consistent palaeocurrent distribution is noted within all three fold limbs (Figure 4.4B). However, while there is clear difference between the unrotated data for Limbs A and B, the unrotated data from the eastern limb (C) falls in three peaks (Figure 4.4A), one which corresponds to the data peak for Limb A, one which follows the regional northerly trend of most palaeocurrent indications in the Falklands Devonian (Section 2.3.4), and a small third peak towards the east-south-east. Thus one can argue that Limbs A and C have in fact undergone the same sequence of deformation (i.e. that conferred by folding on a gently dipping D2 fold axis). It is here considered that a better agreement between the three groups of data is obtained by subjecting only that for Limb B to a further rotation (Figure 4.4C), as would be expected in the common limb of a closed D2 fold pair forming in the dextral transpressive regime proposed by Curtis and Hyam (1998).

An explanation is required for the northwards termination of the HMA, especially as its trace is co-linear with that of the Jersey Harbour Anticline, which can therefore be expected to be controlled by the same underlying basement fracture zone. The abrupt change in plunge and termination of the HMA, which to the south dominates the eastern part of West Falkland, presumably reflects the configuration of some oblique basement feature which interrupts the main NE-SW fracture zone.

Furthermore, although the western portion of the MBHA lies approximately east-west, subparallel to D1 folds to the north-west and south-west, it terminates (together with the corresponding syncline), close to the northern end of the reverse DSBF (Figure 4.3). Although the structure is unclear in the low ground just to the west (south of Mt D'Arcy [UC 21 91] and Green Hill [UC 19 88]), no folding is observed in the well-featured terrain west of Green Hill (Section 4.2.1.5).

The westwards termination of the MBHA at a nearby D2 reverse fault and its association with the abrupt northern termination of the HMA suggests that the MBHA is itself actually a D2 structure, and is most probably a continuation of the D2 anticline in Jersey Harbour. In support, it is noted that the MBHA and the HMA have a similar relationship to NW-SE D3 faults (Section 4.2.3.1), implying some common underlying cause. It is here suggested that the Many Branch Harbour area is underlain by an ENE-WSW basement fracture zone, intersecting the NE-SW zone postulated to underlie the HMA and the Jersey Harbour Anticline, and linking it to the DSBF. Up-to-the-south D2 movement on this easterly-trending fracture zone, synchronous with dextral slip on the NE-SW zone beneath the Hornby Mountains during east-west compression, would be expected to generate a symmetrical anticline in the overlying strata. This is comparable with the formation of the 'isolated anticlines' in Lafonia (Section 4.2.3). The relative incompetence of the Fox Bay Formation around Many Branch Harbour might have assisted the process. This model explains the westwards termination of the MBHA, and the northwards termination of the HMA. The Bold Cove fold pair formed in response to the same eastwards-directed compression as a second-order fold pair on the steep eastern limb of the HMA.

Carcass Bay

The geological map of Greenway (1972) indicates that a short section of the southern end of the Coast Ridge is repeated by a NNE-SSW fault passing into Carcass Bay, and also cross-cut by faults on WNW-ESE and north-south trends. Curtis and Hyam (1998) interpret the Carcass Bay area as a kilometre-scale second order asymmetric fold pair on the eastern limb of the HMA, which has been dissected by a complex series of faults (Figure 4.5A). They believe that the eastern limb of the anticline displays numerous tight to open, overturned south-easterly facing folds with wavelengths of 125 metres to 200 metres. Lower order folds are also present in places, but folding dies out towards the east coast, in the vertical to overturned eastern limb of the anticline. No cleavage is seen in the Carcass Bay area.

Curtis and Hyam (1998) deduced that the syncline in this second order fold pair has been cut out by the Main Carcass Bay Fault, which they interpret as a thrust dipping south-east at about 30° (Figure 4.5b). They also describe two other south-easterly-dipping thrust faults, which they concluded were related to the local folding, although post-dating it. One (their Promontory Thrust) occurs in the small peninsula east of Carcass Bay itself. They also found that there is a set of NE-SW extensional faults, which they thought probably post-date the thrusts. In addition they describe conjugate sets of strike-slip faults lying a few kilometres north-west of Carcass Bay, in which north-south faults apparently display dextral offsets and ESE-WNW faults show sinistral offsets. They observe that this configuration would be consistent with a NE-SW shortening direction.

The present survey found different interpretations of many of these structural elements, and these are now discussed in turn.

The area of folding shown by Curtis and Hyam (1998, fig. 12) (Figure 4.5A) between the two ridges underlain by Stanley Quartzite north-east of Carcass Bay (close to their line of section) is very poorly exposed, with the sloping ground between the ridge crests and the valley-bottom being smothered in solifluction deposits and peat. However, the large scale topographic features on the north-west side of the outer ridge are very similar to those found on the north-west side of the inner ridge. In both places, these features are consistent with those found to develop over an *unfolded* sequence of the lower part of the Port Stanley Formation, the Port Philomel Formation and the topmost part of the Fox Bay Formation (Figure 4.5C).

However, the configuration of the folds in this poorly exposed ground is critical to the interpretation of the MCBF. If the folds (especially the syncline inferred to lie beneath the thrust) do not exist as shown by Curtis and Hyam (1998) (Figure 4.5A, B), it is difficult to see how an east-dipping thrust has juxtaposed the Fox Bay Formation with the Fitzroy Tillite Formation in the Carcass Bay valley. Moreover, very little (if any) displacement occurs at the base of the Fitzroy Tillite Formation on the east side of the Coast Ridge on the line of the MCBF inferred by Curtis and Hyam (1998) (compare Figure 4.5A,C). Instead, the present survey found evidence that the trace of the MCBF crosses the inner ridge [UC 0330 4190] some 4.3 kilometres north-east of Carcass Bay (Figure 4.5C). This configuration suggests that the MCBF dips to the west rather than to the east. The northern end of the MCBF as here interpreted lies parallel with strike and so its extent in that direction is obscured. Towards its southern end, the MCBF divides in two with the eastern splay entering Carcass Bay a few hundred metres to the east. No faults cross-cut the MCBF, or its eastern splay.

The MCBF locally separates the topmost part of the Fox Bay Formation from the Fitzroy Tillite Formation, which indicates that the west-dipping MCBF is a normal fault. The upthrown footfall (to the east of the fault) appears to have brought a steeply dipping to overturned portion of the Coast Ridge D2 fold limb up to the level of a more gently dipping portion. This up-to-the-east movement is contrary to that expected from the mode of formation inferred for the HMA, but it is consistent with the sense of movement on the subparallel Blue Mountain Fault at the south end of

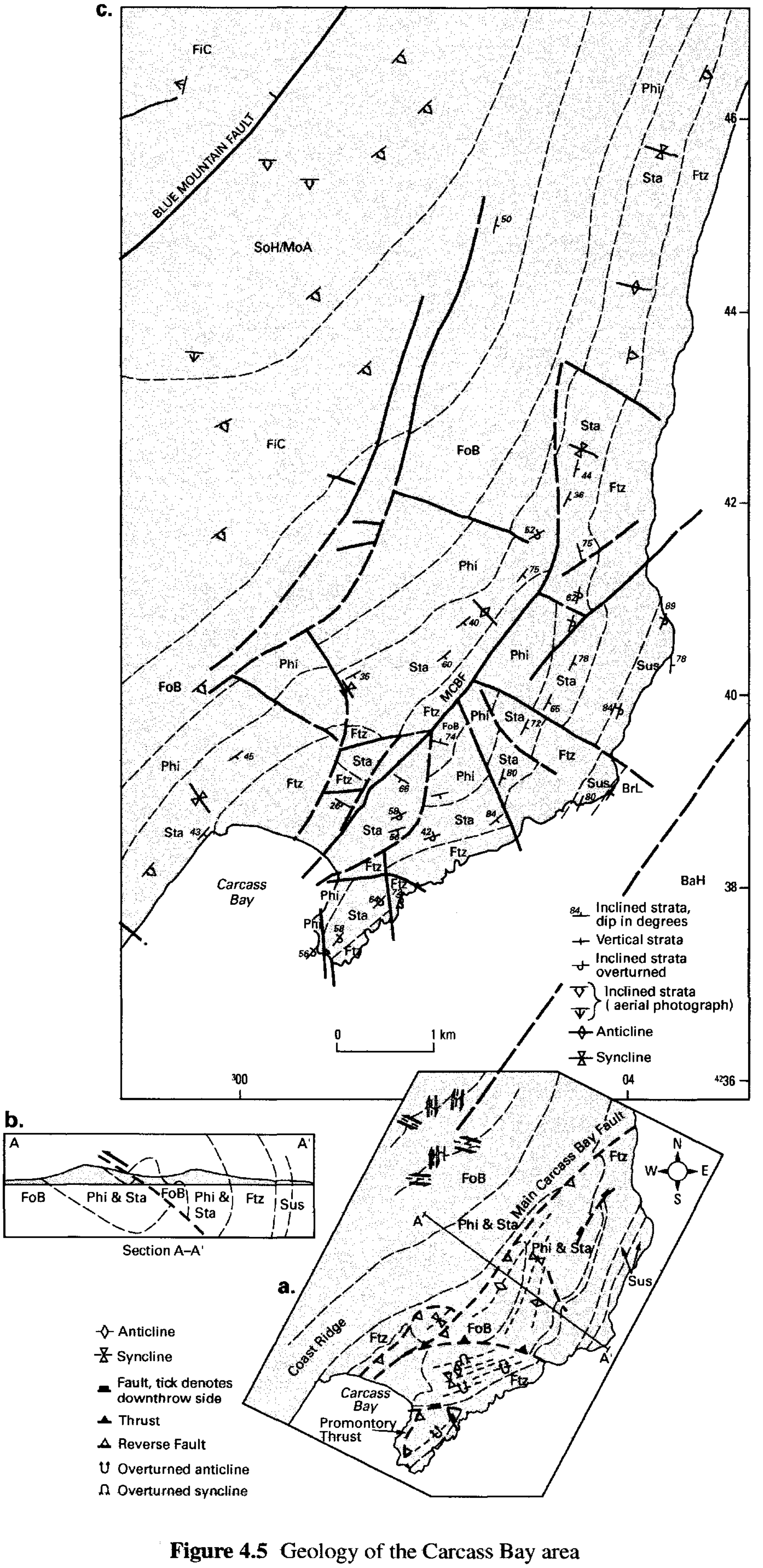


Figure 4.5 Geology of the Carcass Bay area

A: After Curtis and Hyam (1998), fig. 12. B: Cross-section after Curtis and Hyam (1998), fig. 12. C: After Aldiss and Edwards (1998).

the HMA, and with the uplift of eastern East Falkland noted in Sections 4.2.5 and 4.3, implying the MCBF is a D5 structure (albeit perhaps reactivating a D2 or a D3 fault). Indeed, the Blue Mountain Fault cross-cuts a dolerite dyke and so underwent displacement during D5.

The present survey found no evidence for the existence of the 'Promontory Thrust' of Curtis and Hyam (1998). A normal stratigraphic sequence from the Port Philomel Formation to the Fitzroy Tillite Formation is exposed from west to east across the promontory (Figure 4.5C), albeit overturned to the east at about 60° and very possibly shortened by bedding-parallel faults. If the 'Promontory Thrust' exists, it must dip steeply to meet the shoreline west of the (overturned) base of the Fitzroy Tillite Formation at the south-eastern end of the promontory [UC 0093 3728]. In contrast, the field relations observed there during the survey suggest that the western extremity of the promontory has been offset on a north-south high-angle fault zone.

The northern end of the MCBF causes a small dextral offset of the strata, similar to that on the two northerly faults which lie one or two kilometres to the west. These offsets, together with the sinusoidal deflection in the inner ridge between the western end of Carcass Bay and the trace of the MCBF where it crosses the inner ridge, are consistent with the dextral rotation associated with D3 deformation in other parts of the Coast Ridge (Section 4.2.3.1).

Curtis and Hyam (1998) conclude that the complex relationships of the various fault sets around Carcass Bay cannot be explained by a single tectonic stress pattern, and that they probably record a complex evolving stress system post-dating the formation of the HMA. As in the Bold Cove area, this leaves aspects of the local and regional structure unexplained: in particular, their interpretations imply different geological histories for the two areas, although they lie only some 60 kilometres apart along strike on the same structure. The present interpretation finds similarities with structures in the surrounding area, and seems to be compatible with the regional structural development as inferred from evidence elsewhere in the Islands.

4.2.2.2 D2 to the west of the Coast Ridge

The Port Edgar Ridge, which forms the seaward side of Port Edgar, has similarities to the Coast Ridge (Figure 4.2). It is straight, is formed of Port Stanley Formation dipping east at 50° to 60°, and is part of a steep zone (bounded to the west by the Port Edgar Anticline) which extends down-sequence into the Port Stephens Formation. It differs from the Coast Ridge in its north-south orientation. Small-scale folding occurred in the Port Edgar Ridge, and minor structures and fabrics suggest that bedding-parallel faulting has also occurred there. Indeed, the anomalously narrow Fox Bay Formation outcrop is likely to reflect bedding-parallel attenuation of the sequence.

The southwards extent of the Port Edgar Anticline is obscured by faulting but its presence offshore is indicated by westerly dipping strata in the ground around Chaffers Gullet. (This is comparable with the change in strike seen south and east of Port Richards on the west side of the Symonds Harbour Anticline). The topography of the Arch Islands suggests that a segment of the Port Edgar Anticline (here lying parallel to the HMA) also passes through that island group, although the structure is not clear on aerial photographs. In support it should be noted that the distance between Albemarle Rock and the Arch Islands is very similar to the distance between the Edgar Ridge and a parallel inland ridge underlain by the South Harbour Member, and that the structural correlation between the Edgar Ridge and Albemarle Rock is supported by local bathymetry (British Admiralty Chart 2513, 1989). Most strata in the Port Stephens to Port Albemarle region dip towards the north: a shift to easterly dips in the area to the north and north-east of Cape Meredith is presumed also to be an expression of the Port Edgar Anticline or of a related structure.

At the northern end of Port Edgar, the trace of the Port Edgar Anticline is offset eastwards, passing into the gently asymmetric East Bay Anticline (EBA). The dip in the eastern limb of the anticline decreases within only a few kilometres, from about 50° in the Port Edgar steep zone to about 20° near Malo Creek. This inflection in the Port Edgar Anticline coincides with a WNW-ESE structural lineament, which approximately coincides with the abrupt southwards termination of the Symonds Harbour Anticline (SHA, Figure 4.2). The eastern limb of both the EBA and the SHA is steeper than the western one, and in the case of the East Bay Anticline is markedly linear, although both of the corresponding synclines are rather weakly defined. No minor folds and no cleavage are seen in association with these folds. As with the HMA and the PEA, these characteristics suggest that these two east-facing fold pairs are drape folds, formed above blind faults with a considerable easterly downthrow.

Photogeological interpretation shows a zone of relatively steep south-easterly dips in the Hotham Heights area of western Weddell Island. This represents part of the common limb of a north-north-easterly trending fold pair similar to those seen at Symonds Harbour and East Bay. The Hotham Heights Anticline is partly disrupted by subparallel faulting, and it seems that the corresponding syncline has been largely replaced by a fault zone (Figure 4.2). It was suggested in Section 2.3.1.4 that the high ground of Weddell Island represents a continuation of the Mount Alice escarpment. If so, there must be an abrupt discontinuity between the South Harbour Member at Cape Orford and what are probably beds of the Albemarle Member in the south-east of Weddell Island. Together with the considerable change of regional strike in the east of the island compared with South Harbour Rincon and the areas to the east, this suggests that there must be faulting of some kind in Smylie Channel, probably with large uplift to the west. This is supported by relatively steep easterly dips in Stop Island [TC 28 33], but there is insufficient information to suggest a likely configuration of that faulting. Nevertheless, it seems that Weddell Island has been uplifted relative to West Falkland in an analogous manner (albeit to a lesser extent) to the uplift of West Falkland relative to East Falkland.

Reversals of regional dip indicate that Beaver Island is crossed by a syncline and New Island by an anticline. By analogy with large north-trending folds nearby, these are also taken to comprise an east-facing fold pair, although the steepest dips (up to 16° west) occur west of the New Island Anticline.

Photogeological interpretation shows that the north coasts of the Dunnose Head peninsula and of the Passage Islands represent a strike section within the Port Stanley Formation. The changes in strike direction in the Passage Islands describe a very gentle, north-eastwards plunging fold, which is arguably a continuation of the Hotham Heights Anticline (Figure 4.2). By analogy the inflections in the Dunnose Head Peninsula and the Hummock Island chain in King George Bay are also taken to indicate D2 folding. These northerly trending D2 folds evidently post-date the D1 folding, so the trace of the King George Harbour Syncline is inferred to be gently deflected around the north-easterly D2 fold traces crossing the Dunnose Head peninsula. Changes in the strike direction of the D1 fold axes north and north-west of Chartres can also be attributed to D2 folding but the correlation across the Christmas Harbour Fault Zone is not clear (Figure 4.2). The change in regional strike in the Port Stanley Formation at Hill Cove from east-west to WNW-ESE is presumably also a consequence of the D2 folding around King George Bay.

4.2.2.3 The South Jason Line

Although a north-westerly structural 'grain' runs through most parts of northern West Falkland and the adjacent islands, South Jason and the southern part of Elephant Jason are aligned in an east-north-easterly direction, highly oblique to the regional trend (Greenway, 1972). British Admiralty Chart 2514 (1992) shows that South Jason is part of a submarine ridge which continues north-east into Hope Reef: this is also very clear when seen from the air. Chart 2514 also shows the Linblad Reef, with water depths of less than five metres, lying parallel to South Jason and about halfway between South Jason and South Fur Island. This reef continues subparallel to Hope Reef at least as far as Napier Rock, which lies due north of the gap between Carcass Island and the Twins. The chart also indicates the presence of a third submarine ridge, subparallel to the Linblad Reef and about halfway between it and South Fur Island.

The extent of the Linblad Reef and the Hope Reef show that the underlying geological structures cut across those seen elsewhere in the Jasons and which are attributed to D1. It is likely that a subparallel anticlinal trace crosses the southern end of Elephant Jason with the trace of the corresponding syncline lying south-east of the Linblad Reef (Figure 4.2), but otherwise the structures controlling this tectonic axis, the South Jason Line, are unknown. The presence of a D1 cleavage and epizonal metamorphism (Section 4.3) in quartzites of the Jason Islands suggests that they lie on a structural block which has been uplifted relative to West Falkland. This is evidence that the South Jason Line is controlled by a north-west dipping reverse fault, as inferred for the Coast Ridge. Its formation is assumed to be contemporary with D2 elsewhere in the Falklands, although this is not necessarily the case.

4.2.2.4 Falkland Sound Syncline

East of the Coast Ridge, south-easterly dipping strata are also found in the Swan Islands and the Tyssen Islands, indicating that the Falkland Sound Syncline, the geometrically necessary counterpart to the Hornby Mountains Anticline, lies near the eastern side of Falkland Sound (Macdonald et al., 1996). The northerly orientation of the islands between Speedwell Island and Great Island is controlled by regional strike, which in Calista Island [UC 0 3] is close to north-south. Easterly dips observed on aerial photographs of Calista and the most north-westerly of the Elephant Cays suggests that a synclinal axis lies between them and the gentle westerly dips observed on aerial photographs of Clump Island and Ruggles Island. This axis presumably passes north into the NE-SW fault crossing Great Island, which separates areas of west-dipping beds in the south-east of that island from east-dipping beds in the north-west.

A south-west plunging syncline occurs on Shag Rookery Point in the north-west of Lafonia. This is offset by faulting within Kelp Harbour but can be traced south-west as far as Egg Harbour. It seems likely that it is transposed into the line of the Falkland Sound Syncline offshore by further faulted offsets. The northerly continuation of this syncline in Grantham Sound is unclear: it may be truncated by faulting such as that seen at the mouth of Brenton Loch, or it may die out in alignment with the north end of the Coast Ridge steep zone, in an analogous manner to the HMA. Any continuation which may once have existed in the northern part of Falkland Sound has been cut out by D3 faulting (Section 4.2.3.2).

4.2.2.5 D2 in East Falkland

The Wreck Point peninsula, west of San Carlos, is disposed in an angular kink fold, with a southern hinge about two kilometres south of Campito [UC 55 81], a straight north-south common limb and a northern hinge about five kilometres north of Campito [UC 54 88]. This fold is reflected in the sinusoidal trace of the San Carlos Anticline within San Carlos water. A west-dipping reverse fault on the west side of the common limb (which offsets a D1 fold pair in North West Rincon) is assumed to be related to this fold. Its orientation suggests that it is a consequence of D2 eastwards directed compression. A NE-SW striking thrust fault with north-westerly transport direction exposed north of San Carlos settlement (Curtis and Hyam, 1998) could also be part of D2.

Lafonia can be divided into two structural domains by the 'Goose Green Axis', an NNE-trending tectonic lineament aligned with the Goose Green isthmus (Sections 4.2.1, 4.2.5). At this axis, and to the west, the Permian strata are transected by a complex series of linear fault zones, the Lafonia Fault System (LFS), each subparallel to the inferred Falkland Sound Fault. No large strike-slip movements on the LFS are apparent but otherwise there is very little known about the sense and scale of movement on these faults. In the north of western Lafonia, strata dip uniformly to the west (except adjacent to the FSS) but to the south, regional dips in the fault-bounded blocks describe a broad anticline trending NNE-SSW (Figure 4.2). Given the pattern of D2 deformation to the west, it seems likely that this Lafonia Anticline formed in the hanging wall of a blind WNW-dipping reverse fault or thrust repeating the easterly-directed compressional movement beneath the Coast Ridge structure of eastern West Falkland. Alternatively, the anticline in western Lafonia could be a roll-over structure above a north-west dipping normal fault. It might then be contemporary with D5 uplift of East Falkland east of the Goose Green Axis (Section 4.2.5.1).

The eastern extent of D2 deformation in southern Lafonia is marked by a complex fault zone exposed at Swan Pond, east of North Arm [UC 44 21]. To the west of Swan Pond Arroyo, regional dip is uniformly to the south-east at about 10°, increasing to about 23° close to a photogeological lineament parallel to the valley. This lineament marks the western bounding fault of a graben about 800 metres wide. Several subsidiary fault zones are exposed within the graben. Bedding in the narrow intervening fault blocks is tilted either to the west or to the east at up to 50°. This gives the impression that the area within the graben has been gently folded parallel to the fault zone, but nearly all the fold hinge zones are either overthrust to the west, or have been replaced by zones of intense brecciation in which no sense of movement is apparent. The hinge has been preserved in only one, gently plunging, anticline. These structures appear to be the consequence of localised NW-SE compression within the graben, presumably during D2 although possibly later. The same fault zone is exposed some 19 kilometres south-west, at the head of The Mullet [UC 29 08] but here the fault blocks have been tilted at up to 15°, with dips of up to 48° occurring only close to fault zones.

To the north the displacement on many elements of the LFS dies out, possibly being transferred onto faults parallel to Brenton Loch. These faults are well exposed in a complex zone of shearing, including belts of fault breccia up to 20 metres wide, on the coast north-east of Hope Place [UC 58 68]. They pre-date the dolerite dyke near Saladero.

In the north of East Falkland, two anticlines and an intervening syncline lie approximately parallel to Falkland Sound (Figure 4.2), apparently forming a large-scale dome-and-basin interference pattern with major D1 folds. The Standing Man Hill Anticline, between Port Louis and Rincon Grande [VC 1 9], arguably continues south-west to cause the change in plunge direction seen in the Malo Hills Anticline, and so forming the dome-shaped outcrop of the Port Stephens Formation. The syncline forming Bold Point at Salvador [UC 9 9] probably continues south-west through Chata Hill [UC 90 92], intersecting the D1 fold axes at almost 90°. It appears that Bombilla Hill [UC 80 90] marks the trace of the same D2 synclinal axis, having been displaced by WNW-ESE trending faults between Bombilla Hill and Chata Hill. The linearity of the north-western margin of the Malo Hills inlier, lying parallel to this D2 syncline, is reminiscent of the style of D2 deformation in West Falkland. The easterly plunge of the D1 Coutts Hill Anticline might also be caused by D2 deformation (Figures 4.1, 4.2, 4.7).

Although the wavelength and orientation of these D2 folds in East Falkland is comparable to those in the west, their style is more symmetrical, with no clear sense of vergence. They are nevertheless considered to have arisen in the same way, with the differences attributed to a greater depth to the controlling faults in the underlying crystalline

4.2.3 Third phase of deformation (D3)

basement.
Large-scale structures attributable to a third phase of deformation mainly occur in a NE-SW zone centred on Falkland Sound, although D3 faulting and small-scale deformation is more widespread (Figure 4.6). Although it is convenient to treat these structures separately from those formed during D2, as discussed in Section 4.2.2, D3 could be regarded as a late stage of D2. However, as long as the close relationship between D2 and D3 is borne in mind, the terminology used to describe them should not be an obstacle.

D3 was dominated by dextral NE-SW faulting, perhaps in part reactivating older structures, although some folding did occur. The most typical D3 fold style is a subtle dextral kinking of the Coast Ridge. D3 folding in East Falkland includes small-scale dextral kinking. Movement also took place on NE-SW trending faults within Falkland Sound. Although most D3 deformation occurred close to Falkland Sound, widespread NW-SE trending faults which post-date D1 or D2 folds are also thought to have formed in this phase (Figure 4.6). The NE-SW and NW-SE faults comprise a conjugate set resulting from east-west compression. The D3 structures onshore appear to be a consequence of dextral rotation of the crustal blocks on which the Islands are founded.

The D3 fault systems lie parallel to the western margin of the Falkland Plateau Basin (to the east of the Islands) which is formed by a complex series of down-to-the-east normal faults (Richards et al., 1996b). These seem to occur within Permo-Triassic to early Jurassic sedimentary sequences, with onlap replacing active faulting at this margin sometime in the ?Middle to Late Jurassic. Onshore, it can be inferred that D3 pre-dated the Early Jurassic dolerite dykes, so may have ceased somewhat earlier than the faulting at the basin margin.

4.2.3.1 D3 deformation in West Falkland

Examination of the West Falkland Coast Ridge on aerial photographs or from low flying aircraft shows that the exposed quartzites of the Port Stanley Formation, which form the crest of the ridge, lie in a series of straight segments up to five kilometres long, each separated by an angular deflection of between 145° and 175° (Plate 29). In most instances, these deflections form very gentle dextral kink folds with a common limb of between 300 metres and 1900 metres in length but in the extreme case at Shag Harbour [UC 18 64], the dextral deflection occurs over about four kilometres. There are a few sinistral fold pairs. In some cases the nose of the angular, approximately vertical fold formed by these deflections is cross-cut by small strike-slip faults. The ‘synclines’ (facing west) are usually slightly more acute than the corresponding ‘anticlines’ (facing east).

In the Coast Ridge between Hill Gap and Shag Harbour, the common limb of the largest of these very gentle D3 folds is cross cut by an oblique-slip reverse fault, dipping east at about 20°, and striking N025° with a net dextral displacement of up to 400 metres. It seems that if movement on this fault had been more extreme the ridge segments would have developed an overlap like that seen at Carcass Bay. Probable D3 structures at Carcass Bay were discussed in Section 4.2.2.1.

The most extreme example of D3 folding in the Coast Ridge occurs east of Bold Cove [UC 32 79], where the Poke Point Hill Anticline is deflected westwards by about 40° in similar style to the D3 folds elsewhere in the Ridge (Figure 4.3). About two kilometres to the south-west [UC 314 790], the anticline is offset dextrally by a NNW-SSE trending fault. Taken together, the fold and fault which displace the Poke Point Hill Anticline mimic the more gentle D3 fold pairs in style and orientation.

By analogy, the NNW-SSE dextral faults which displace the Coast Ridge immediately north of Bold Cove [UC 30 82], in the Narrows of Port Howard [UC 24 75] (Figure 4.3) and, on a smaller scale, at Shag Harbour [UC 18 64] are also D3 structures. None of these faults is seen to displace the axis of the HMA or the MBHA even though the two larger ones are both aligned with large faults to the north of the respective anticlinal axes (Figure 4.3). This emphasises the close relationship between D2 and D3 structures and incidentally also supports the interpretation of the MBHA as a D2 fold (Section 4.2.2). It is not known if the NNW-SSE faults die out in the fold hinge or if they have been obscured by turning to lie parallel with bedding, in a similar manner to that suggested for these structures by Curtis and Hyam (1998), and seen in the Saddle area [UC 00 74].

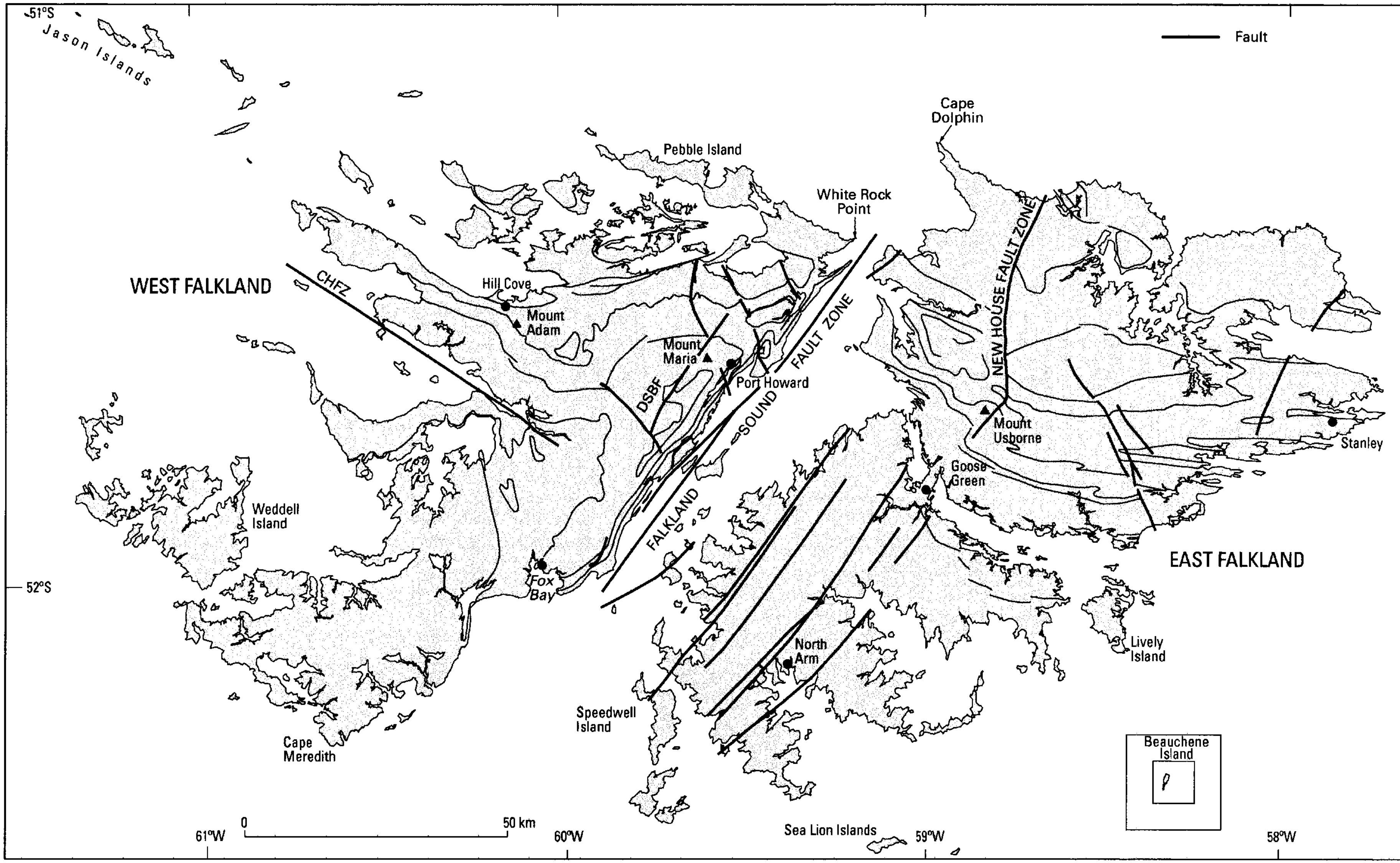


Figure 4.6 D3 structural elements

CHFZ Christmas Harbour Fault Zone DSBF Double Stream Basin Fault (D2)

The presence of a large fault zone within Christmas Harbour (Figure 4.6) can be inferred from the juxtaposition of north-dipping Port Philomel Formation with Fox Bay Formation strata on the north side of Mount Philomel [TC 89 64], from the opposed regional dips of the strata on the northern and southern sides of the harbour near Chartres, and from the presence of subparallel faulting on the southern shore of Christmas Harbour. For example, a subparallel fault plane is exposed on the foreshore opposite Shallop Point [TC 8633 6670] where it replaces a meso-scale synclinal fold axis, possibly the King George Bay Syncline itself.

The Christmas Harbour Fault Zone offsets D2 fold axes (Figure 4.2) and so is unlikely to be older than D3. Although the Christmas Harbour Fault Zone is subparallel to the D5 faults to the north (Figure 4.10), it appears not to displace the dolerite dykes near Chartres. It is therefore unlikely to have seen movement as late as D5. No high angle faults are known to have formed during D4. The Christmas Harbour Fault Zone is thus assigned to D3. It dies out rapidly at its eastern end, which lies within a ENE cluster of dolerite dykes, suggesting a structural relationship with dyke emplacement.

4.2.3.2 Falkland Sound Fault Zone

The presence of a fault zone in Falkland Sound was postulated by Greenway (1972), who suggested it as a normal fault accounting for the relative uplift of West Falkland. Marshall (1994b) regarded it as a dextral strike-slip fault separating zones of differential Gondwanide crustal shortening. Thomas et al. (1997) suggest that as much as 200 to 300 kilometres of dextral displacement occurred along the Falkland Sound Fault Zone during Gondwana break-up in the Early Jurassic, based on a consideration of regional geology of the Falkland Islands and what were once the neighbouring parts of Gondwana (Section 6).

By contrast, Curtis and Hyam (1998) note that if large dextral displacements had occurred on a fault lying close to the east coast of West Falkland (and so bending eastwards around the bulge in the coastline near Carcass Bay) some evidence should be seen onshore in the patterns of faulting at Carcass Bay. No such evidence is seen, and they conclude that significant dextral displacement cannot have taken place along the proposed Falkland Sound Fault. Nevertheless, Hyam (1997; 1998) asserts that dextral movement of up to 60 kilometres has occurred on the Falkland Sound Fault Zone, although he provides no answer to the objections raised by Curtis and Hyam (1998). As described in Section 4.2.1, and as noted by Marshall, (1994b), and by Curtis and Hyam (1998), the style and intensity of D1 changes from East to West Falkland. While this could occur gradually, evidence from the area of the Goose Green Axis, some 30 kilometres to the east (Section 4.2.1) suggests the change is more likely to have occurred by strain partitioning across fault zones (compare Marshall, 1994b).

In the absence of seismic surveys in the Sound there can be no direct evidence for the existence or otherwise of a fault zone. However, results of the Falkland Islands Geological Mapping Project, not available to Curtis and Hyam (1998), suggest that faulting in Falkland Sound is more likely to have occurred than not. Although the British Admiralty Charts do not provide positive evidence for the presence of faults (compare Curtis and Hyam, 1998), they show sufficient detail of the submarine topography to suggest constraints on the position of faults whose existence has been inferred from other information.

The strongest new evidence comes from geological mapping in the north of Falkland Sound. This shows that whereas Jersey Point [UC 47 99] is formed of Port Stanley Formation dipping steeply south-east, Fanning Head [UC 51 96] exposes an easterly plunging fold in the Port Stephens Formation. Some major discontinuity must exist between Fanning Head and Jersey Point. In the absence of any faulting in the Sound, not only would there have to be an abrupt reversal of fold plunge west of Fanning Head, allowing the entire Devonian sequence to be repeated out to a synclinal axis near the west side of the Sound, but the tight D1 anticline in Fanning Head would have to disappear within a few kilometres. Neither circumstance is likely. Indeed, Port Stephens Formation sandstones seem to continue along strike from the Rookery Sands area [UD 53 00] to Tide Rock [UD 50 01]. Conversely, British Admiralty Chart 2558 (1988) shows that the north-east to south-west trending ridge forming Jersey Point continues on the sea floor to Sunk Rock (between Race Point [UD 54 02] and White Rock Point [UD 47 03]) and not to Tide Rock, which appears to stand alone to the east of a relatively deep channel. It is very probable that a major fault lies between Tide Rock and Jersey Point.

The change in style of D1 deformation across the Sound means that it is very difficult to correlate any individual structure in East Falkland with those in the West with any certainty. The most likely direct correlation is between the Rookery Sands Thrust and the inferred Sound Bridge Thrust (Section 4.2.1.5). Even if such a correlation is made, an estimate of the amount of dextral movement parallel to the Sound since D1 would depend on knowing the magnitude of the movement of West Falkland towards East Falkland during formation of the Coast Ridge. That is obscured by an unknown amount of dip-slip faulting in the Sound and in the Coast Ridge steep zone, and by uncertainties over the original thickness of the sedimentary sequences in the deformed areas.

In the simplest case, one can suppose that following the formation of the Coast Ridge during D2, the Port Stanley Formation exposed in Wreck Point at San Carlos [UC 52 92] was aligned with that seen in Jersey Point. The Port Stephens Formation sandstones seen at Fanning Head could likewise be matched with exposures in White Rock Point, although that relationship is thought to have been obscured by D4 and D5 deformation (see below, and Sections 4.2.4 and 4.2.5). This suggests that at least five kilometres of D3 dextral displacement occurred on faults within Falkland Sound. This is comparable with the inferred offset of at least 10 kilometres in the D1 thrust at the north end of the Sound (Section 4.2.1.5) and with a possible offset in the D1 deformation front (Section 4.2.1.7).

Further indications of faulting, albeit more equivocal, is seen by comparing the geology of the Swan Islands with the adjacent part of West Falkland. Whereas both Swan Island [UC 2 5] and West Swan [UC 17 59] appear to be underlain by strata of the Egg Harbour Member (Section 2.4.5.1) the adjacent eastern shore of West Falkland is formed by the lowest part of the Brenton Loch Formation. If the same sequence was once present here as in the Brenton Loch area, some 40 kilometres to the east, then at least 4600 metres of the Brenton Loch Formation and Bay of Harbours Formation has been cut out between West Swan and the mainland. Some of the reduction in thickness in the Coast Ridge steep zone may be due to layer-parallel thinning on numerous relatively minor slip-planes dispersed through the sequence. Conversely, the sequences on West Falkland are generally thinner than on the east. The Fitzroy Tillite Formation is about 30 per cent thinner, and the Port Sussex Formation some 55 per cent thinner in the Coast Ridge than on East Falkland. If there is no faulting between West Swan and West Falkland then the Brenton Loch Formation and the Bay of Harbours Formation would have to be at least 75 per cent thinner, but this is consistent with the sparse clay mineral maturity data (Section 4.3). However, if the formations do become proportionally thinner up-sequence, it is very likely that such rapid thinning would have been controlled by growth faults at the basin margin against West Falkland. Such faults are likely to have been reactivated during D3 or D5 or both. Similar arguments can be applied to the relationship between the Bay of Harbours Formation strata on West Island and the basal Brenton Loch Formation exposed on the coast near Carcass Bay.

Although the presence of dip-slip faults between West Swan and West Falkland cannot be proven at present, circumstantial evidence suggests it should be considered as a strong possibility. The most probable line of faulting is indicated by the deep water between Perks Island [UC 153 592] and an islet west of West Swan [UC 157 587]. To the south-west this aligns with a channel some 35 metres deep between West Falkland and the shoals of Oberon Patch [UC 13 55]. On aerial photographs the rocks in Oberon Patch appear to be well-bedded, forming open folds with north-south axes and a peak to trough distance of about 100 metres. This localised zone of deformation might indicate proximity to a fault zone. No such folding has been noted on West Swan Island. The southwards continuation of this line of faulting, and its relationship with structures near Carcass Bay, is unclear.

The objections proposed by Curtis and Hyam (1998) to large-scale dextral strike-slip faulting near the coast at Carcass Bay demonstrate that if dextral faulting has occurred in that part of Falkland Sound, then it did so somewhat further offshore. The most likely line passes between West Swan and Swan Islands, which have a NNE-SSW channel up to 40 metres deep between them. Curtis and Hyam (1998) found that a set of NE-SW striking, closely spaced joints in the Egg Harbour area of western Lafonia have undergone dextral displacement. This is consistent with dextral movement on the Falkland Sound Fault Zone (Hyam, 1998).

Northern truncation of the Falkland Sound Fault Zone

Three lines of evidence show that the Falkland Sound Fault Zone does not continue as far north as Cape Dolphin (compare Richards et al., 1996, fig. 2; Hyam, 1998). Firstly, field observation shows that the Cape Dolphin peninsula is formed of well-bedded *Skolithos* sandstones of the Port Stephens Formation. The sandstones are gently inclined but are otherwise undeformed: jointing is widely spaced, with no evidence for movement, recrystallisation or veining, even on the most minor scale.

Secondly, British Admiralty Chart 2558 (1988) shows that a submarine ridge extends between the point just south of Cape Dolphin [UD 63 18] and Eddystone Rock [UD 56 27] suggesting that the sandstones forming the ridge have not been cross-cut by significant faulting. This inferred continuation is substantiated in a sketch of Eddystone Rock by Southby-Tailyour (1985, p. 97) which shows bedded sandstones dipping north-east apparently at 20°, approximately consistent with the photogeological interpretation of the Rock and in a similar orientation to sandstones at Cape Dolphin.

Thirdly, there is no evidence in the marine gravity field (SEASAT) for the continuation of the Falkland Sound Fault Zone to the north-east of the northern entrance of Falkland Sound. While Richards et al. (1996b) postulate a link between the FSFZ and north-south faulting in and around the North Falkland Basin, the FSFZ is here thought to have been offset by D4 or D5 faults a little to the north of White Rock Point (Section 4.2.5).

4.2.3.3 Displacement on Goose Green Axis

Different styles of deformation occurred on the Goose Green Axis to the north and to the south of the Wickham Heights during D3.

North of the Wickham Heights an area of subdued topography occurs between the upper reaches of the San Carlos River [UC 77 79] and Elephant Beach Pond [UD 76 06]. Discontinuities in large scale structures across this area suggest that it is underlain by a major fault zone, here named the New House Fault Zone. The lack of exposure in that area and with the generally weak photogeological expression of bedding are consistent with locally intense fracturing. The fault pattern in this zone is very probably far more complex than indicated by the single fault strand shown on the map (Figure 4.7). This New House Fault Zone is taken to be the expression of the Goose Green Axis in northern East Falkland.

There is evidence for at least three kilometres of dextral offset along the New House Fault Zone. This is based on the correlation of three pairs of large structural features: the base of the Fox Bay Formation near Elephant Beach House and its continuation to the south between New House and Cavadi Rocks; the north-south D1 syncline near Elephant Beach Pond and the syncline at Bombilla Hill; and the inferred D1 Coutts Hill Anticline and the Malo Hills Anticline (Figure 4.7). In this interpretation the fault zone would also cross-cut the D2 Letterbox Hill Anticline which might thus continue from the Elephant Beach Pond area towards Coutts Hill, causing the easterly plunge of the D1 Coutts Hill Anticline. The D1 fold axial planes (which are assumed to be steeply dipping) and the base of the Fox Bay Formation (which dips at 60° or less) have all been offset by a similar amount, indicating that fault movement was predominantly strike-slip, rather than dip-slip. The narrow fault-block east of Black Tarn appears to be a graben, consistent with the tensional regime which would be expected at this bend in the Goose Green Axis as a consequence of dextral strike-slip motion. However, the strike-slip displacement dies out north of the base of the Lafonia Group.

The difference in tectonic style means that it is not possible to correlate the Port San Carlos steep zone with any structure to the east of the Goose Green Axis, but if a similar dextral offset has occurred south of the Malo Hills Anticline it seems possible that the Verde Syncline is structurally equivalent to the No Man's Land Syncline (Figure 4.7). The lack of clear structural correlations in this sector and differences in D1 deformation suggest uplift occurred east of the NHFZ, instead of, or in addition to any strike-slip motion.

The zone of steeply dipping and overturned strata at Port San Carlos (Section 4.2.1.4) shows no evidence for large-scale D3 deformation, apparently because dextral displacement was instead taken up on the New House Fault Zone and on NE-SW faults near Race Point [UC 53 98]. Small-scale dextral kink folds lying subparallel to bedding which deform *Skolithos* sandstones at Settlement Rocks [UC 6239 9360] post-date D1.

It seems probable that several phases of fault movement have occurred along the Goose Green Axis. It was suggested in Section 4.2.1 that the change in style of D1 deformation was accommodated by fault movement on the Goose Green Axis, and so these suggested offsets in D1 structures near Mount Usborne could have originated during D1, at least in part. However, the inferred trace of the D4 Sand Grass Thrust shows a net sinistral offset, attributed to D5 uplift east of the Goose Green Axis (Sections 4.2.4.2 and 4.2.5.1). The dextral movement on the New House Fault Zone thus seems to have pre-dated D4, and so is attributed to D3.

The Goose Green Axis is a long-lived tectonic axis akin to the Falkland Sound structure. As such it reinforces the notion that the Falkland Islands is traversed by a set of NNE lineaments founded in the basement. These can be expected to influence offshore structures in a complementary way. Lateral variation in the Bay of Harbours Formation from west to east across Lafonia (Section 2.4.5) suggests the Goose Green Axis also influenced Permian sedimentation.

4.2.3.4 D3 fault movement in Lafonia

Photogeological interpretation shows that faults subparallel to the Goose Green Axis occur as far east as Laguna Isla (mostly throwing down to the west), and to the west as far as Shag Rookery Point (most throwing down to the east, where this can be determined). Most of the faults have very straight outcrop traces, suggesting that they are steeply dipping or vertical at the surface. However, many of the faults in western Lafonia lie subparallel to bedding. This, and the lack of readily identifiable stratigraphic markers, means that it is commonly very difficult to determine the direction of throw, let alone the amount. There is no evidence for large strike-slip movements. These faults possibly mark D2 or D5 movement as well as, or instead of, D3.

Field evidence confirms that there are numerous faults in Lafonia. Many exposed faults are small, with an observed throw of only a few metres, while some have an unknown throw but are marked by fracture zones of up to five metres width. However, only those faults which coincide with a photogeological lineament can be mapped. Conversely, although many lineaments are apparent on remote images of Lafonia, without field evidence only those which mark some contrast in topography are interpreted as possible faults, and only those where bedding traces can be shown to have been displaced are considered definitely to be faults. Unfortunately, where the topography is subdued, bedding traces are very faint and displacement across fractures is very difficult to demonstrate. The consequence is that in Lafonia probably only the largest of the faults are shown on the map, possibly with a bias towards north-easterly trends.

There is some localised evidence for D3 compression within Lafonia. The south-eastern part of Lafonia is notable for the occurrence of swarms of fold structures of a distinctive type here described as 'isolated anticlines'. Where exposed, as on the north shore of North West Arm [UC 3089 1738] and on Horn Hill Rincon [UC 5480 2900], they are seen as abrupt up-domings of otherwise uniformly flat-lying strata. The folds thus formed are some three metres to 15 metres across, with limbs dipping at between 40° and 60°. They are weakly asymmetrical, but in contrast to folding seen in the D1 fold belt there is no corresponding syncline: the anticlines are isolated. (There are no isolated synclines). This style of folding can be attributed to flexure above a blind reverse fault.

In the gently dipping strata of Lafonia the isolated anticlines give rise to a distinctive type of very straight dark-toned photogeological lineament, across which no displacement of bedding is apparent. (A good example occurs just north and north-east of the centre of aerial photograph 28-29, trending NNW.) On the ground these lineaments are found to mark low straight ridges with sparse sandstone float and diddle-dee cover. Several lines of evidence suggest that these lineaments do not mark dolerite dykes: they are shorter than nearly all lineaments known to mark dolerite dykes; none show a central depression, which is usual for weathered dykes; no known dykes in Lafonia form strong lineaments of this type; and no dolerite is exposed on this type of lineament.

Lineaments interpreted as isolated anticlines form several loose swarms which can be several kilometres in extent. South-west of the Bay of Harbours, they are generally aligned NNW-SSE, but the alignment seems to vary slightly from one fault block to the next. However, although most of the isolated anticlines on Horn Hill Rincon and the neighbouring promontory to the south also lie NNW-SSE, a few form a NE-SW conjugate set at an angle of about 60°. The isolated anticlines are thought to reflect localised compression within fault-blocks formed during D3. Their orientation implies maximum compression from WSW to ENE and is consistent with the alignment of D3 dextral displacements.

Small south-dipping thrusts, each with a displacement of a few metres or decametres, are present on the coast between Bull Point and Porpoise Point [UB 43 98] (Geochem Group Ltd, 1996). These thrusts occur in a section which is isolated from the rest of the mainland by intersecting northerly faults passing through Bull Cove [UB 41 98]. Although they are of a different style to the isolated anticlines just described, they are also tentatively attributed to localised deformation during D3.

4.2.3.5 D3 in eastern East Falkland

Large north-easterly and north-westerly faults occurring sporadically east of the Goose Green Axis are tentatively assigned to D3. The most conspicuous of the north-westerly fault zones occurs just east of Mount Pleasant Airport. It displaces bedding traces and D1 fold axes sinistrally by up to 1.9 kilometres, although some dip-slip movement cannot be discounted. The orientation and sense of displacement of this fault zone suggest that it is conjugate to north-north-easterly D3 dextral faults.

Curtis and Hyam (1998) found a set of mesoscale dextral kink bands (up to five centimetres in width) and associated faults post-dating the S1 fabric in eastern East Falkland. These vary in orientation from predominantly north-south to NE-SW. These are plausibly a local expression of D3 deformation. Curtis and Hyam (1998) also observed post-S1 dextral and extensional faults and kink bands striking ENE-WSW in eastern East Falkland. Although their age relative to the north-south to NNE-SSW structures in the same area could not be determined, these structures

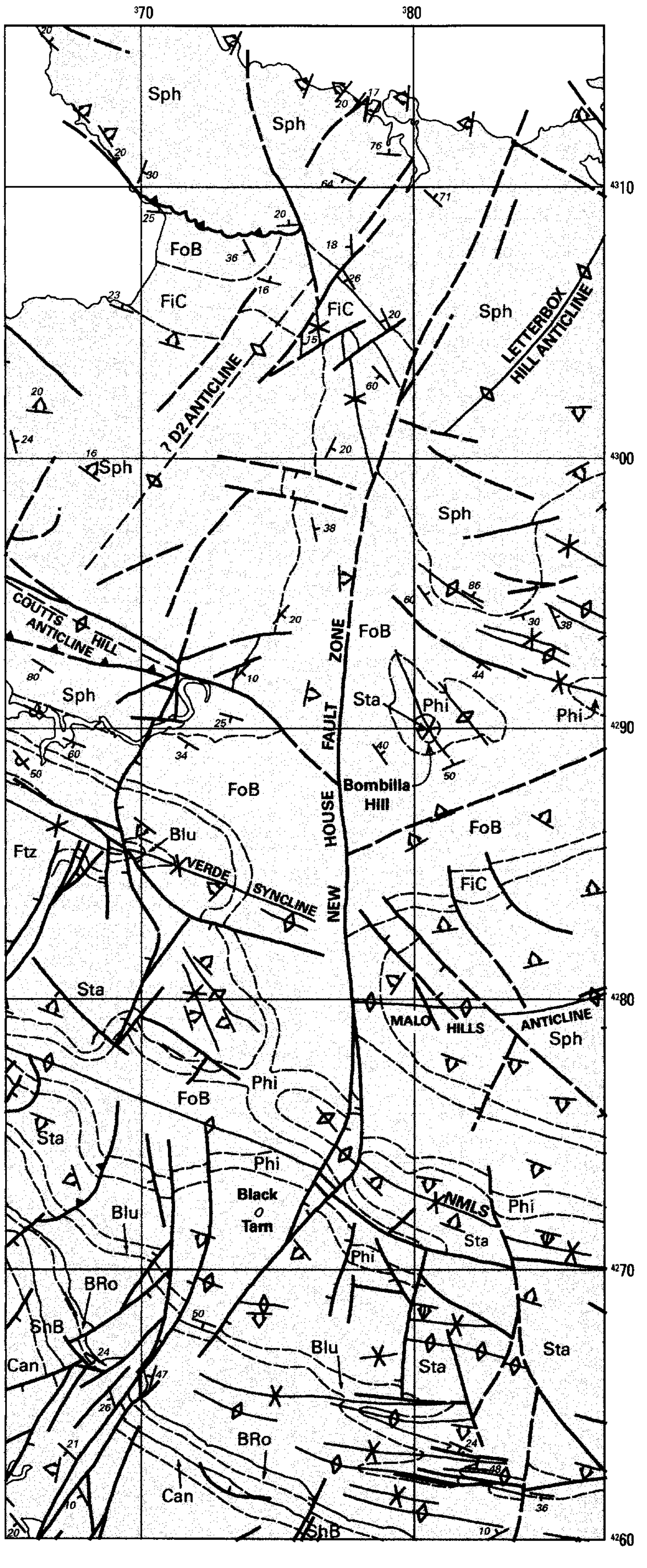


Figure 4.7 Major structures adjacent to the New House Fault Zone

NMLS No Man's Land Syncline Symbols as on geological map

apparently pre-date a late compressional event here assigned to D4 (Section 4.2.4.2), and so are considered to be part of D3.

4.2.3.6 D3 in the Jason Islands

Aerial photographs of Jason East Cay [SD 9 4] reveal a pair of open fractures striking NNW (about N310°), probably only 20 to 30 metres apart, between which the strata have been rotated to a northerly strike, dipping steeply east. The fractures thus appear to form a north-easterly-vergent brittle fold pair or a sinistral kink band. This would be complementary to dextral D3 fault movements seen in other parts of the Islands. Similar structures possibly occur on the Fridays [TD 2 3].

4.2.4 Fourth phase of deformation (D4)

The principal onshore expression of the fourth phase of deformation was the emplacement of the south-west -verging Pebble Island thrust sheet in the north of the Islands (Figure 4.8). It is here argued that components of this thrust sheet, mainly composed of Port Stephens Formation strata, extend from Sedge Island in the west to Cape Carysfort in the east, forming nearly all of the northern coastline of the Islands, and that it was emplaced after D3 deformation.

The D4 thrusting was accompanied by open folding and locally intense brittle deformation, presumably in a high-level compressional regime. Similar structures have been detected offshore to the north and west (Richards et al., 1996), although those to the west of the Islands are more likely to be D1 structures (Section 4.2.1). If D2 and D3 occurred during break-up of the Gondwana supercontinent, D4 is probably a result of compression during this process, and of the rotation of the crustal blocks on which the Falkland Islands lie (Section 6).

4.2.4.1 West Falkland and adjacent islands

The evidence for the existence of a major thrust in the north of the Islands is best demonstrated at the eastern end of Whale Bay (Figures 4.8, 4.9). A prominent negative break of slope in the hillside above that shoreline marks a contact between quartzites of the Port Stephens Formation which form a line of craggy outcrops in the steeper ground above, and strata of the Fox Bay Formation and Port Philomel Formation underlying the lower ground and exposed at the shoreline. The outcrop pattern of the contact indicates that it is approximately planar, dipping gently north or north-east. The quartzite immediately overlying the contact is hard, recrystallised and closely jointed, commonly approaching brecciation. Bedding is unclear. Recrystallisation and the development of joints perpendicular to bedding tend to obscure *Skolithos*, which in some cases is faintly visible only as raised circular structures a few millimetres in diameter on bedding surfaces. Moving away from the contact to the north, fracturing becomes generally less, bedding is clearer (dipping north-easterly at 10° to 20°), and *Skolithos* unobscured.

By contrast, the sandstones exposed at the shoreline around Whale Bay are not recrystallised and are generally soft. The pervasive closely-spaced jointing found in the overlying quartzite is not seen, neither is brittle shearing or brecciation. Bedding dips steadily to the south at 20° or more, although some folding occurs close to the contact. The reversal of the stratigraphic sequence, the metamorphic inversion and the structural discordance together demonstrate that this gently dipping contact is a thrust. The trace of the thrust meets the coast on the north side of Whale Bay [UD 3010 0605]. It reappears on the south coast of Pebble Island, where it is most extensively exposed, with the greater part of the island being underlain by the thrust sheet. It is therefore named the Pebble Island Thrust.

A large north-west plunging syncline within the Pebble Island thrust sheet forms the eastern end of Pebble Island, passing south-east through Ship Harbour [UD 29 09] and continuing to a few kilometres south-east of Goat Hill [UD 39 02]. This, the Pebble Island Syncline, lies oblique to the east-west D1 folds in the Golding Island group. The closely spaced fracturing commonly seen near the base of the thrust sheet typically strikes N124°, subparallel to this fold axis, dipping 70° to the north-east. The orientation of these structures and the regional strike of the thrust plane indicate that the Pebble Island thrust sheet was emplaced towards the south-west.

Localised zones of deformation do occur within the thrust sheet. Brittle deformation in sandstones exposed on the north coast of Pebble Island, west of Green Rincon [UD 093 169] is expressed in kink folding and outcrop-scale thrusting, the latter being cross-cut by a small north-trending extensional fault (D5, Section 4.2.5.2). At Seal Cave [UD 358 077] (north of White Rock House) there is a 200 metre-wide zone of very closely spaced fracturing, striking N261° and dipping 85° to the south-east.

To the west, the Pebble Island Thrust passes offshore just west of Middle Peak [UD 06 15]. Keppel Islet appears to be composed of Port Stephens Formation sandstones dipping to the south-west and so is here thought more likely to be part of the Pebble Island thrust sheet than an along-strike continuation of the north-westerly dipping Fox Bay Formation strata either side of North West Passage. From their along-strike alignment, one would also expect that the Port Stephens Formation sandstones in Port Egmont Cays, the Wreck Islands and Sedge Island likewise form part of this D4 thrust sheet.

There is no evidence for the continuation of the Pebble Island Thrust to the west of Sedge Island. The structural style and orientation of the outer Jason Islands are similar to those of Carcass Island: all are interpreted as part of the D1 fold belt. It has been argued that the deflection of strike near South Jason marks a D2 structure which extends only part-way to Sedge Island (Section 4.2.2.3). If so, this would lie south of the D4 Pebble Island Thrust.

To the south and east of Whale Bay, the lines of D4 thrusting are obscured by D5 high-angle faulting (Section 4.2.5), but can be inferred from the presence of very large apparent throws on high-angle faults (as for the Sound Bridge Thrust, Section 4.2.1.5), especially where this is accompanied by structural discordance and by a contrast in small-scale deformation and degree of recrystallisation. This is seen south-west of Round Hill and at the extreme east end of Port Purvis (Figure 4.9). The shoreline between Ferny Bank and Garden Hill exposes fault-bounded wedges of Fitzroy Tillite Formation and of dolerite [UD 339 000] which lie between recrystallised Port Stephens Formation to the north and Port Stanley Formation to the south.

A similar effect is found on the north side of White Rock Point, where the superb preservation of *Skolithos* in Albemarle Member sandstones exposed just east of Melvern Creek [UD 4570 0305] contrasts markedly with the faint remains of the same trace fossil in the recrystallised sandstones in the lower parts of the D4 thrust sheet nearby to the west. Moreover, the Melvern Creek exposures are structurally continuous with Devonian strata around White Rock Bay and Jersey Harbour, indicating that the trace of the Pebble Island Thrust lies to the north and west of White Rock Point. A zone of structural complexity exposed on the coast north-east of Goat Hill (between Easting 387 and 428) is possibly a consequence of D4 deformation, but its relationship to the D4 thrust plane remains unknown.

The displacement of the Pebble Island Thrust by D5 faults in the area north of White Rock Bay obscures its relationship with the D2 Jersey Harbour Anticline (Section 4.2.2.1), but there is no evidence that the thrust sheet was affected by D2 deformation.

4.2.4.2 East Falkland

The structural relationship described from Whale Bay is repeated on East Falkland in The Sand Grass area, east of Foul Bay (Figure 4.8). Recrystallised and fractured Port Stephens Formation sandstones are exposed in the escarpment to the north of The Sand Grass, and folded but otherwise unaltered Fox Bay Formation strata occur at its foot [UD 69 10]. The outcrop pattern suggests that the contact between is a thrust plane, the Sand Grass Thrust, although some high-angle faults have occurred at or close to the contact where it is exposed on the foreshore north-west of Elephant Beach [UD 691 108]. The similarities with the Pebble Island Thrust suggest that the Sand Grass Thrust is most probably an eastwards continuation of the same structure. The style of deformation seen in the Sand Grass area and to the north is quite different from that seen in the vicinity of the D1 Rookery Sands Thrust, for example at Settlement Rocks and on the coast south of Race Point (Section 4.2.1.4). The differences are compatible with the Sand Grass Thrust having formed later at a higher structural level, resulting in brittle deformation in the thrust sheet.

The Sand Grass Thrust is truncated by NNW-SSE high-angle faulting near Elephant Beach Pond. It is probably offset to the north (Section 4.2.5.1) but the area inland of Smylies Black Point is weakly exposed. Some of the exposures which are present are of very highly fractured quartzite and there is also at least one of a clast-supported breccia. This fracturing might be related to late (D5) high-angle faults, but some could date from D4 deformation. A fault plane dipping at 20° to the east is exposed at the coast near the mouth of Little Creek [UD 782 132]. It is unclear whether this is part of the regional D4 thrusting, or if it is a consequence of local compression between two northerly-trending high angle faults which intersect the coast to either side. Two open fold pairs, each some 25 to 50 metres from peak to trough, are exposed in Chimango Valley [UD 7782 0986]. The folds verge to the south-south-west, plunging westwards at 12°. From the similarity to the folding in the Pebble Island thrust sheet, it seems likely that the folds in Chimango Valley occurred within the Sand Grass thrust sheet, thus constraining the position of the outcrop of the thrust plane.

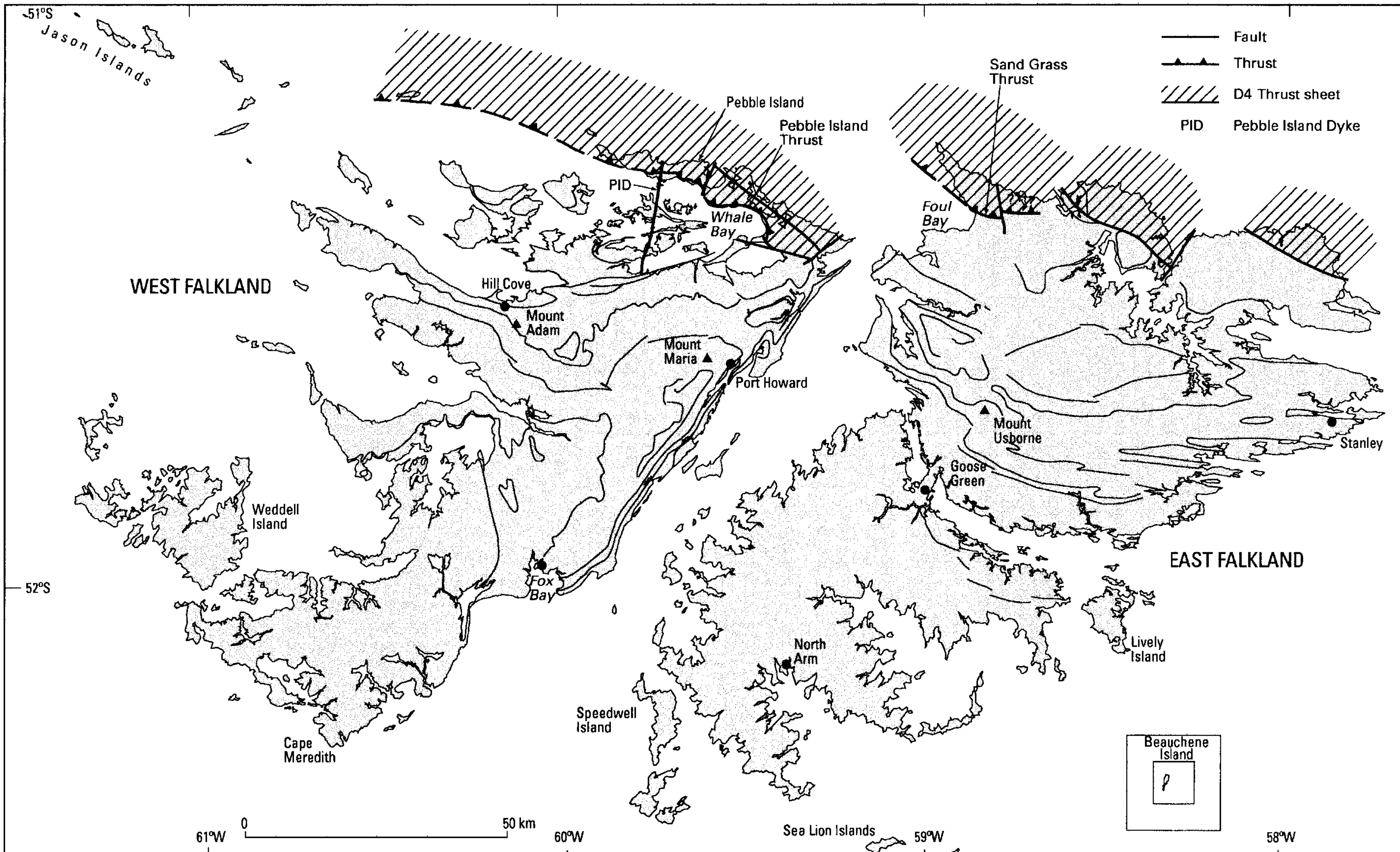


Figure 4.8 D4 structural elements

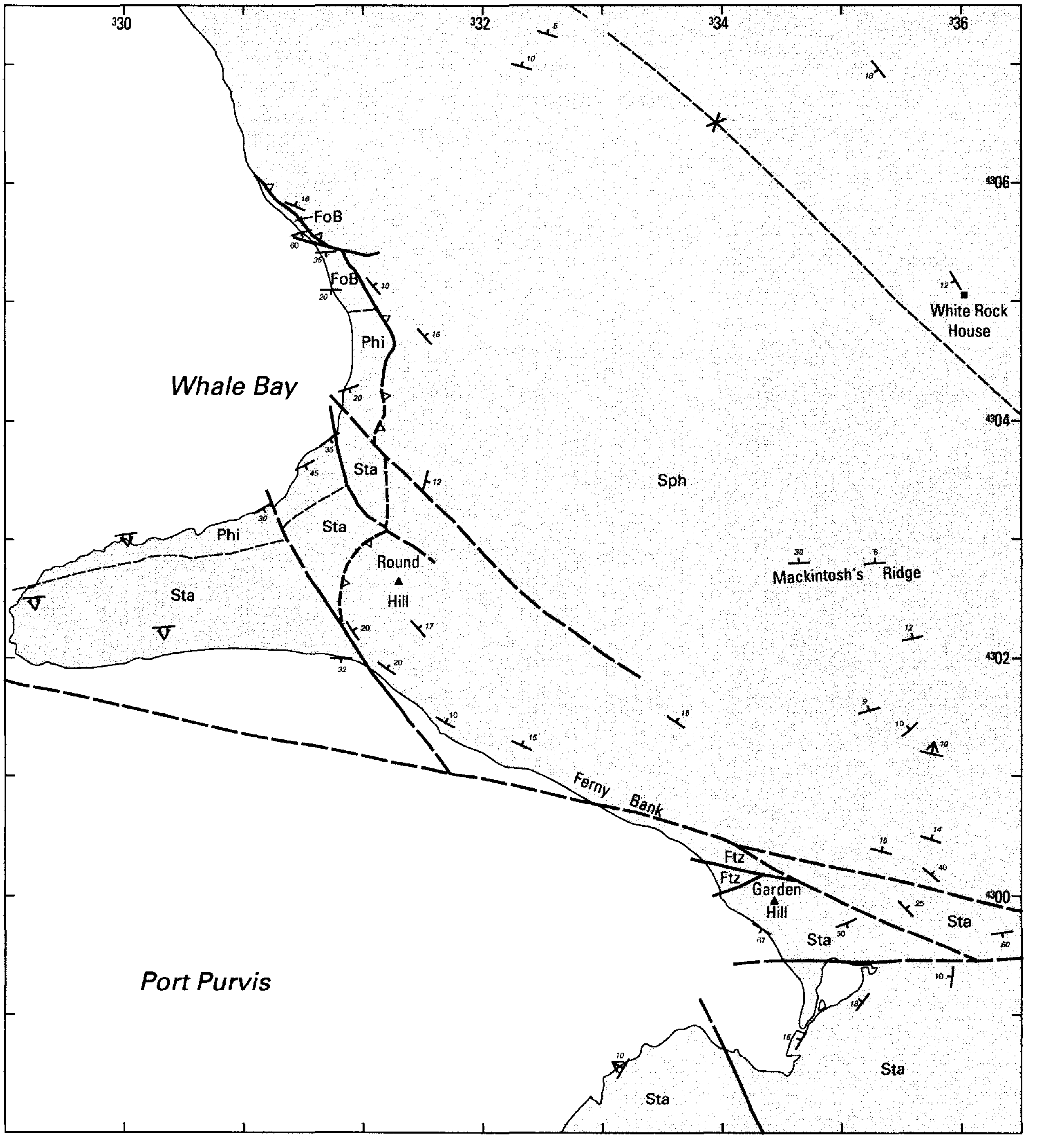


Figure 4.9 Geology around Whale Bay and Port Purvis

Symbols as on geological map

The continuation of the Sand Grass Thrust further to the east is also unclear, but seems to have been entirely hidden by D5 faulting. Its probable position can be traced using the same criteria as in the area between Port Purvis and White Rock Bay, as discussed in more detail in the Section 4.2.5.1. Thus the terranes north of the WNW-ESE D5 fault zones between Limpet Creek and Salvador, and between Swan Pond and Cow Bay (Figure 4.10) are thought to be part of the Sand Grass thrust sheet (Figure 4.8). A WNW-ESE trending D4 anticline can be inferred to occur north of Salvador, associated with a planar fabric in the Port Stephens Formation between Brazo del Mar and Sheilas Creek. As noted in Section 2.3.1.2, the Limpet Creek Member, as mapped near Limpet Creek, is possibly part of the Fox Bay Formation instead. If so, it is probably bounded to the east by a continuation of the Sand Grass Thrust.

Several lines of evidence suggest that the Pebble Island Thrust and the Sand Grass Thrust post-date D2 and D3 deformation, and that they can therefore be attributed to a fourth phase of deformation. As already noted, the syncline in the Pebble Island thrust lies oblique to D1 folds in the Golding Island to Main Point area. This and other evidence for south-west (rather than southerly) vergence shows that the D4 structures post-date D1. D2 and D3 deformation in the Coast Ridge and in Falkland Sound is not seen in the Sand Grass thrust sheet in Cape Dolphin or its undersea continuation to Eddystone Rock (Section 4.2.3). There is no evidence that other portions of the Sand Grass Thrust or of the Pebble Island Thrust have been deformed by D2 folds. The sense of relative offset between the Sand Grass Thrust and the Pebble Island Thrust across Falkland Sound is not compatible with the D3 dextral strike-slip movement inferred from other evidence, although some apparent offset may have arisen by large vertical displacements during D5 faulting.

This conclusion seems to be compatible with the observation by Curtis and Hyam (1998) of a late reverse shear exposed on Whittingtons Rincon, on the south-east side of Dans Shanty Creek. This post-dates an extensional fault which is plausibly part of the D3 deformation. A similar relationship can be inferred near Johnson Harbour [VC 2851 9305] where three intersecting fabrics are exposed in phyllitic siltstone. Bedding is obscured by an ESE-WNW S1 slaty cleavage, dipping south. That is deformed by sparse dextral kink bands about five centimetres wide, trending subparallel to S1 but dipping north. These are attributed to D2. Both S1 and the dextral kink bands are offset by a series of sinistral kink folds, each about one centimetre wide and between five centimetres and 30 centimetres apart, together with subparallel fractures. These strike WSW-ENE, dipping northwards at 20° to 30°, so are consistent with foreland deformation during D4.

There is no strong evidence for the chronological age of the D4 event. According to the geological map (Figure 4.8), the D4 thrust sheet on Pebble Island has been cross-cut by the Pebble Island dyke, assumed to be about 190 Ma old (Section 3.1.1.8). However, the field evidence for this relationship is not conclusive and it is possible that the thrust actually post-dates the dolerite (Section 3.1.1.6). In either case, it seems most likely that D4 thrusting, like D2 and D3 deformation, is related to the break-up of Gondwana during the Early Jurassic.

Richards and Fannin (1997) conclude that NW-SE oriented Palaeozoic thrust faults constrained the development of the southern part of the North Falkland Basin, some 50 to 100 kilometres north of the Islands. Their orientation and position suggest that these could also be D4 structures. Richards et al. (1996b)note that the Palaeozoic platform immediately east of the Malvinas Basin (offshore to the west of the Islands) is transected by WNW-ESE-oriented, northwards-dipping thrust sheets. As argued in Section 4.2.1, these thrusts are more likely to be part of the D1 deformation, but there is insufficient constraint on their age to preclude assignment to D4.

4.2.5 Fifth phase of deformation (D5)

As mentioned in the previous section, the D4 thrusts are in large part displaced and hidden by subsequent high angle faulting. In addition, dolerite dykes thought to have been emplaced in Early Jurassic times are in many places offset by faults. Here all these late-stage high angle faults are attributed to a D5 brittle deformation (Figure 4.10). This D5 faulting follows several different trends and may have occurred in several phases over a period of time. Richards et al. (1996b) and Richards and Fannin (1997) suggest that certain north-south trending faults offshore to the north of the Islands are transfer faults, linking the displacements on NW-SE extensional fault sets. The intersecting fault sets in the north of the Islands, especially in East Falkland, plausibly arose in the same way.

4.2.5.1 East Falkland

A north-easterly trending fault passing into Little Creek [UD 79 10] and a north-north-westerly fault close to Jackass Rincon [UD 73 14] can be inferred to have westerly downthrows from the apparent displacement of the outcrop of the Fish Creek Member around Elephant Beach Pond. The NNW fault strand truncates the Sand Grass Thrust and so presumably displaces its eastwards outcrop to the north, as noted in Section 4.2.4.2. The NNW fault is itself truncated by a set of NE to NNE trending faults including the large New House Fault Zone (Section 4.2.3.3) which is inferred to pass into Laguna Lorenzo. The inferred line of this fault passes close to exposures of highly brecciated quartzite about three kilometres east of Elephant Beach House [UD 809 057], in which north-east striking shear planes are conspicuous.

The New House Fault Zone possibly lies up to one kilometre east of the position shown on the map, indeed it is likely to be a complex of anastomosing fault strands. As this fault zone apparently cross-cuts all other onshore structures it is assumed to displace the Limpet Creek Fault Zone just off the north coast (Figure 4.10). This would be consistent with the position of the deepest water immediately offshore, as shown by the distribution of kelp, and also with a NNE lineament in the regional gravity field. A parallel linear gravity anomaly occurs offshore in alignment with the Salvador Straits Fault Zone (Figure 4.10).

By analogy with the model proposed for D4/D5 faulting in West Falkland (Section 4.2.4.1), the probable trace of the D4 Sand Grass Thrust (as displaced by D5 faults) is indicated by the apparent magnitude of faulting in the NW-SE zone passing from just north of Salvador to Limpet Creek (Figure 4.10). The Limpet Creek Fault Zone reaches Port Salvador on the north side of Sheilas Creek, due east of Salvador settlement [VD 0739 0083 to 0760 0070]. This exposes a 40 metre-wide zone of fault breccia, with highly fractured Port Stephens Formation sandstones (with a WNW-ESE trending cleavage) resting against weakly fractured (and uncleaved) Fox Bay Formation sandstones (part of a NNE-trending D2 syncline). Salvador and Rincon Grande are known for the frequent occurrence of well-developed quartz crystals with pyramidal terminations, consistent with their having formed within cavities in a high-level fault zone.

The lateral displacement of the Sand Grass Thrust to the north-east of Elephant Beach Pond seems to be repeated in the narrows at the entrance to Port Salvador. No D4 structures are seen to cross-cut the D2 Standing Man Hill Anticline east of Horseshoe Bay and Salvador Hill, or in Port Salvador, so it is inferred that the LCFZ near Salvador is offset by a NNE-SSW Salvador Straits Fault Zone (Figure 4.10). Large faults of the same orientation lie onshore in the Rincon Grande area.

Fault zones parallel to the Limpet Creek Fault Zone are believed to occur in the Macbride Head area. This interpretation is based on the presence of weak WNW-ESE photogeological lineaments coincident with indentations in the coastline, changes in the photogeological texture of the cliffs and of the inland topography, and changes in the general orientation of bedding and cleavage. Broad zones of intense brittle fracture and cataclasis are exposed at Swan Pond and at the north end of Cow Bay. These possibly repeat the north-dipping sedimentary sequence in the lower part of the Port Stephens Formation and separate domains of rather different structural styles. To the north of the inferred fault zone in Cow Bay are exposed numerous low angle faults, open folds rather than kinks and much brittle fracturing and shearing. Anomalously steeply dipping strata on the south-west side of faults close to Macbride Head suggest local drag of strata in the vicinity of a high angle fault throwing down to the NE.

Volunteer Lagoon separates areas with opposed dips, but the topography suggests that this reversal occurs across a fault rather than a syncline, an interpretation supported by the recrystallisation seen in sparse outcrops near the western end of the lagoon, and by local disruption of cleavage. This east-west fault seems to displace the north-east to south-west lineament which is taken to mark a D3 fault of relatively small displacement (cf. Figure 4.6). A fault is inferred to lie north of Mt Brisbane from the topography, the repetition of resistant strata along strike to the west and observation of shearing parallel to a valley to the east, between Mt Brisbane and Volunteer Beach [VC 38 95]. The linearity of the north shore of Volunteer Point suggests this could be controlled by faulting offshore, parallel to the faults near Macbride Head.

In this interpretation of northern East Falkland, NNE-SSW D5 high-angle faults are downthrown to the west, progressively displacing the D4 Sand Grass Thrust southwards (Figures 4.8, 4.10). The westwards continuation of the Sand Grass Thrust as the Pebble Island Thrust would be consistent with this pattern. In the absence of evidence for D5 faulting in Falkland Sound or between Cape Dolphin and Eddystone Rock, the inferred D5 displacement between White Rock and Elephant Beach seems most probably to have taken place on the WNW fault zone between Paloma Beach and Moss Side. This lies parallel to the Limpet Creek and Cow Bay Fault Zones and could plausibly pass north of Race Point, truncating the Falkland Sound Fault.

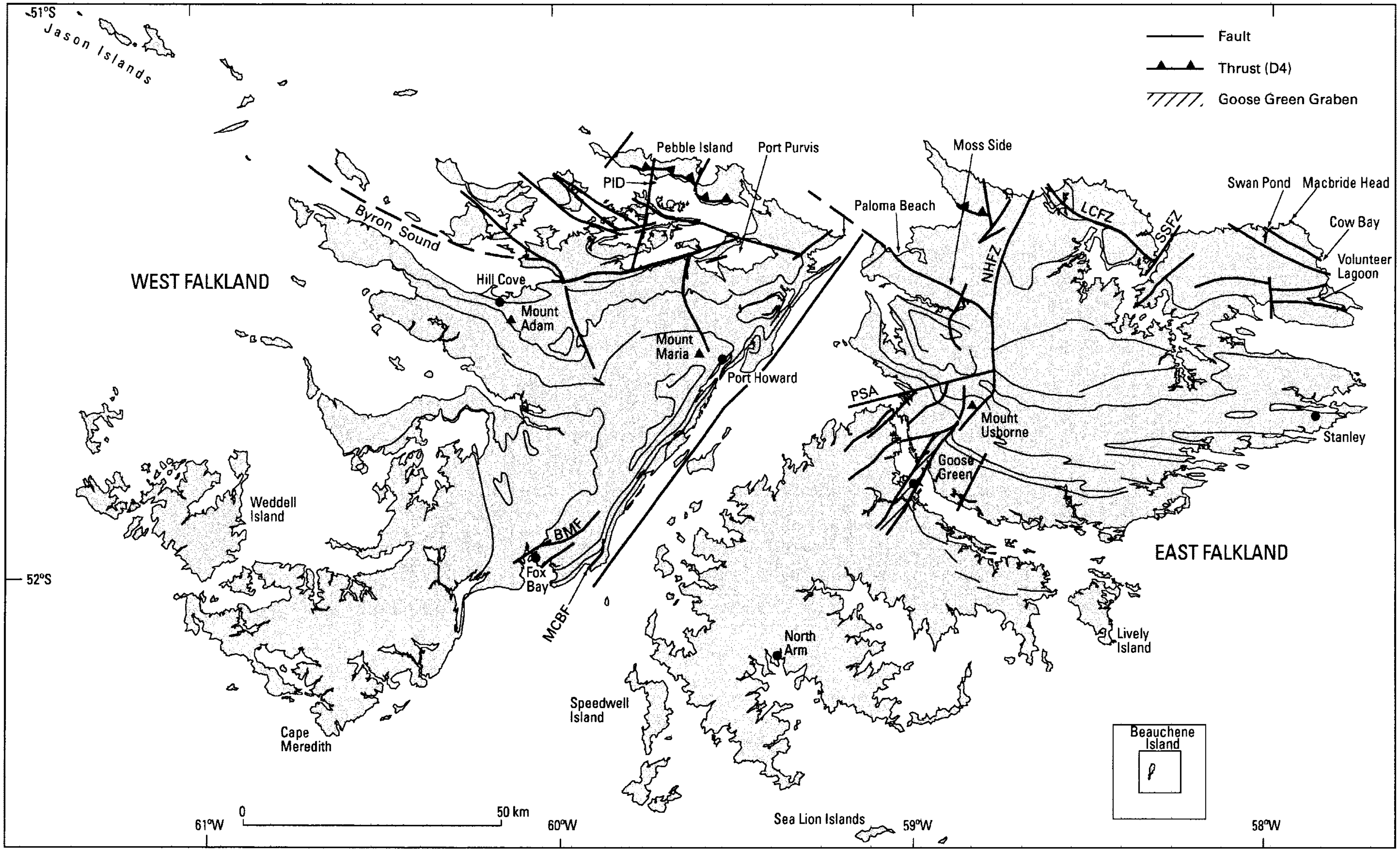


Figure 4.10 D5 structural elements

BMF Blue Mountain Fault MCBF Main Carcass Bay Fault PID Pebble Island Dyke PSA Port Sussex Axis

To the south of the Wickham Heights, the continuity of the Lafonia Group outcrop on either side of the Goose Green Graben shows that it has apparently not been laterally offset on the Goose Green Axis. However, comparison of the linear Black Rock Member outcrops south of Mount Usborne and at Port Sussex shows that although they lie almost parallel at about N115°, they have been mis-aligned by some 5.5 to 6 kilometres, forming a large-scale dextral kink fold pair or drape. This is thought to have occurred during D5, caused by some three kilometres of uplift between Ceritos and Port Sussex (Section 4.3).

The S1 regional cleavage has been displaced by this D5 fold, but with less apparent lateral rotation than the bedding, and not consistently with the lateral offset of the bedding. Thus the offset of the Black Rock Member and of S1 cleavage is attributed to relative uplift east of the Goose Green Graben, although possibly with some lateral (strike-slip) movements on the various faults between Ceritos and Port Sussex.

The eastern axial flexure of this fold pair is formed by the Goose Green Graben. The western flexure, the Port Sussex Axis, strikes at N240° from near the mouth of Brenton Loch [UC 58 70] across the head of Port Sussex to just west of Bodie Peak (Figure 4.10). The two fold axes thus converge and the kink fold dies out to the north-east between the Third Corral Mountains and the Malo Hills range. Localised very tight folding and ENE-WSW trending shears in the Black Rock Member near the head of Port Sussex [UC 6358 7369], and the complex ENE-WSW faulting at the northern tip of Lafonia are attributed to deformation within the western axial zone.

A south-easterly-vergent reverse fault between Bodie Peak and Cantera Mountain suggests compression within the common limb of the D5 kink fold. However, the greater part of the block between the two axial zones (the 'Cantera Block') has been subjected to net extension. This has given rise to a series of NE and ENE normal faults in the Cantera Block, mostly throwing down to the south-east, but the largest displacement occurred on faults aligned with the Goose Green Axis.

The remarkably linear sides of the Goose Green isthmus immediately suggest that this narrow strip of land is bounded by steeply dipping faults. There is abundant confirmatory evidence from the occurrence of fault breccias and fracture zones, subparallel minor faults, offset of bedding and juxtaposition of contrasting rock types. These can be seen in various places on Camilla Creek (especially the west side), shorelines around Darwin Harbour, in Bodie Creek (especially downstream from its junction with Findlay Creek), and at the southern end of Brenton Loch. Exposure inland is poor, but faults can locally be inferred from the offset of topographic features marking bedding. A small quarry for road building materials opened on a hillside about a kilometre south of Salinas Beach [UC6120 5560] exposed part of the fault zone bounding the west side of the isthmus. Here strata dipping east at between 10° and 20° are cross-cut by a 10 metre-wide shear zone. Most of the shear planes strike between N040° and N050° and dip east at 60° to 80°, consistent with normal (extensional) faults downthrowing to the east. However, some more gently dipping shear planes appear to be thrusts subparallel to bedding, which truncate the normal faults, indicating a later stage of compression.

Southerly-dipping geological boundaries exposed between the faults bounding the isthmus are displaced northwards relative to their counterparts to east and west, demonstrating that the isthmus lies within a graben. Vertical movements of 400-500 metres on the west side and of about 1000 metres on the east side of the graben can be inferred from the displacement of the base of the Bay of Harbours Formation and of a horizon of slump folding in the upper part of the Brenton Loch Formation (Section 2.4.4.3).

4.2.5.2 West Falkland

Extensional faults with WNW-ESE orientations occur between the eastern end of Port Purvis and Byron Sound (Figure 4.10). These fault sets are in part bounded by east to ENE extensional faults in River Harbour, Rock Harbour and East Island. They follow the same trend as fault sets in the southern part of the North Falkland Basin, thought by Richards and Fannin (1997) to have formed when late Palaeozoic thrusts were reactivated during the Mesozoic extension, sliding some way back down their soles.

The larger north-trending faults seen on Pebble Island seem to be bounded by these WNW-ESE faults and are oblique to them. They follow similar north and north-north-easterly trends as the large D5 faults in East Falkland, although the fault zone forming the west side of Elephant Bay (and exposed just west of Pebble Island settlement) has a large downthrow to the east. This is demonstrated not only by the apparent dextral sense of displacement of the north-dipping Pebble Island thrust, but also by the juxtaposition of Albemarle Member sandstones west of the fault with South Harbour Member (or Fish Creek Member) sandstones in the north-west plunging syncline east of the fault. Conversely, a NE-SW fault inferred to lie between Pebble Island and Pebble Islet apparently displaces the Pebble Island Thrust down to the west, so that it lies south of Keppel Islet.

It was proposed in Section 4.2.3 that NNW trending faults formed during D3, but in the north of West Falkland faults of this trend cross-cut and displace dolerite dykes in several locations. Furthermore, at Sound Bridge, one of the faults on this trend apparently cross-cuts the ENE D5 faults passing into River Harbour. However, as this fault trend appears to be complementary to D3 structures, it seems more likely that they were reactivated rather than initiated during D5. The ENE D5 faults in River Harbour offset dykes of the Sulivan Swarm (Section 3.1.1.5), including the Pebble Island Dyke (Figure 3.1), by up to five kilometres.

The fault zone between Sound Bridge and River Harbour could be a good deal more complex than shown. A series of fault-bounded blocks of quartzite forming a strip about two kilometres wide lies between the main outcrops of the Port Stephens Formation and of the Port Stanley Formation. Although assigned to the Port Stanley Formation on the map, some doubt remains about the identity of the poorly exposed quartzites in these blocks. Nevertheless, this strip can be regarded as the structural continuation of Purvis Rincon, and with it as the equivalent to the common limb of a south-verging fold pair, analogous to the overturned strata around Port San Carlos, south of the Rookery Sands Thrust (Section 4.2.1.5). In the River Harbour area, however, both the overthrust anticline and the syncline have been obliterated by D5 high-angle faults.

The Main Carcass Bay Fault cross-cuts all other structures in the Carcass Bay area. The Blue Mountain Fault cross-cuts a dolerite dyke. Both are both downthrown to the west (Section 4.2.2.1). This is contrary to the sense of movement expected for faults associated with the D2 deformation in the Coast Ridge, but is consistent with the up-to-the-east D5 movements found in East Falkland.

Richards et al. (1996b) found that the geometry of the North Falkland Basin, especially in the southern portion, is controlled by intersecting north-south and NW-SE trending faults. They postulate that the NW-SE fault trend may have originated by strike-slip and extensional reactivation of subparallel thrusts, such as those identified offshore to the west of the Islands (Section 4.2.1), and that this could have occurred during the Cretaceous or Cenozoic. These major fault trends found offshore conform to the orientation of D5 faults onshore in the north of the Islands, indeed, the two NW-SE faults near Macbride Head are closely aligned with faults found in the offshore areas both east and north of the Islands (Richards et al., 1996b, fig. 2). This is consistent with the D5 faulting onshore having occurred after the emplacement of Early Jurassic dolerite dykes.

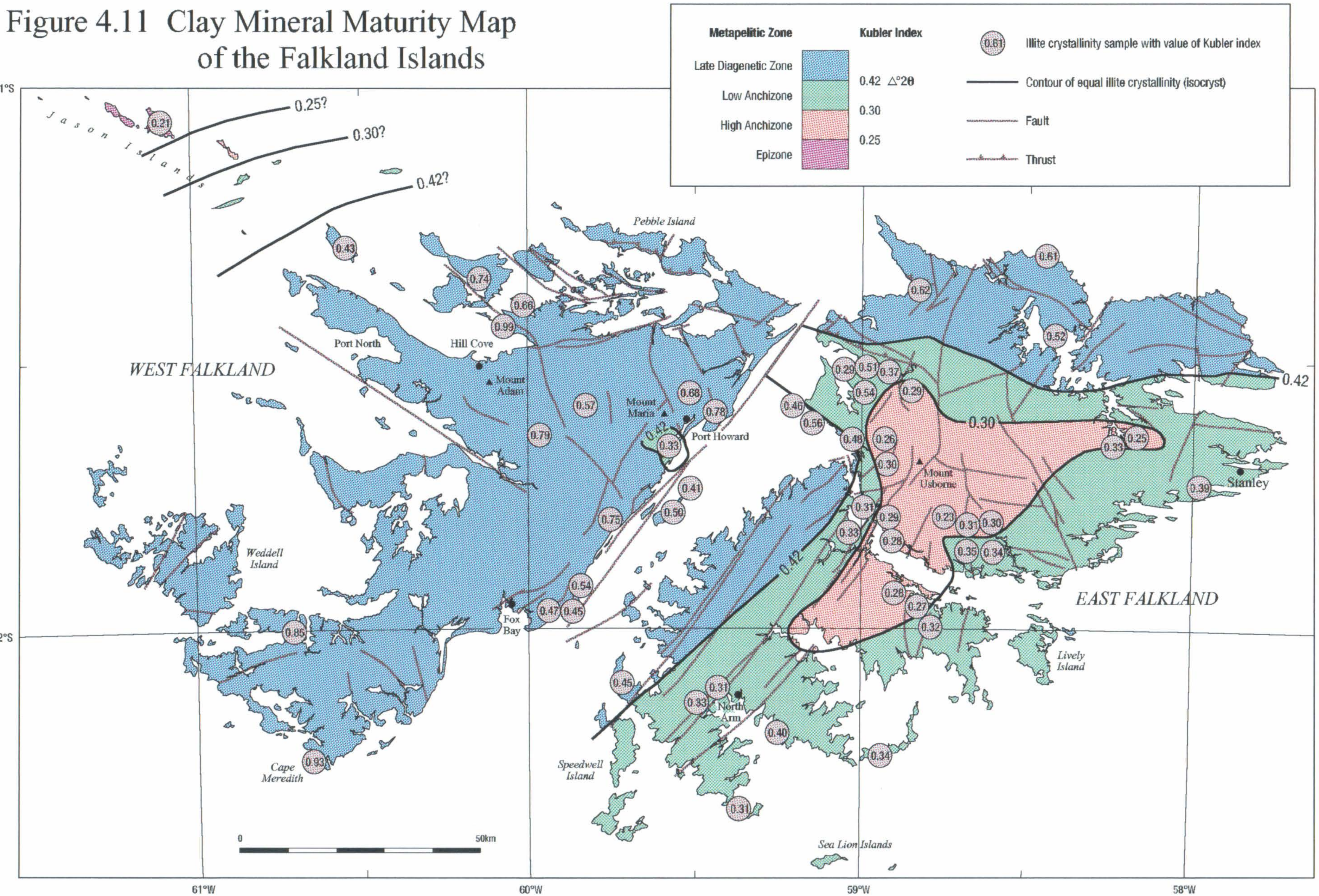
4.3 Metamorphism

The regional deformation described in Sections 4.1 and 4.2 was accompanied by only very low grade metamorphism. While the distribution and intensity of tectonic fabrics gives some indication of how metamorphic grade varies from one part of the Islands to another, determinations of clay mineral maturity and vitrinite reflectance are more informative. These analytical methods show that most of the strata exposed in the Islands remain in the late diagenetic zone (Figure 4.11). The central and southern portions of East Falkland reached the anchizone, but only the centre of that island is in the high anchizone, or locally in the epizone (greenschist facies). The outer Jason Islands show the highest grades of metamorphism in the Phanerozoic sequences, in the epizone. These results show that previous descriptions of the variations in metamorphic grade are too generalised. In particular they indicate that the major variations in metamorphic grade do not follow the trend of the D1 fold belt, as might be expected, but that their distribution is controlled by D4 and D5 structures instead.

High grade metamorphic rocks found in the Cape Meredith Complex are described in Section 2.2. Examples of contact metamorphism and possible hydrothermal metamorphism have been noted in Sections 3.1 and 3.2.

Variations in the style and intensity of the tectonic fabrics associated with D1 and D4 are described in Section 4.2. These provide a broad indication of the variation in metamorphic grade, but the apparent fabric intensity is strongly dependent on local variations in rock type, strain, and even the degree of weathering. It is at best a crude measure. In summary, S1 fabrics are strongest in the outer Jason Islands, and some areas east of the Goose Green Axis in northern East Falkland. Elsewhere, S1 is very weak or absent (Section 4.2.1). A closely spaced fracture set developed close to the D4 thrust planes and in localised zones elsewhere in the thrust sheets (Section 4.2.4). No regional fabrics associated with D2, D3 or D5 have been noted. Previous assessments of regional variation in

Figure 4.11 Clay Mineral Maturity Map of the Falkland Islands



metamorphism concluded that there is an abrupt change across Falkland Sound (Marshall, 1994b), or a gradual decrease from east to west across the Islands (Curtis and Hyam, 1998).

Analysis of clay mineral maturity (illite crystallinity) depends on the observed gradual conversion of smectite to illite with increasing temperature and so with depth of burial (Elsinger and Glasmann, 1993; Merriman and Kemp, 1996; Merriman and Peacor, 1999). In general, this conversion is irreversible so that where basinal sequences have been inverted clay mineral evidence of the maximum burial depth is retained and can be used to estimate the amount of uplift. The technique can be used on most mud-rich rocks, even on mudstone intraclasts in sandstones.

Some 53 new clay mineral maturity determinations are plotted in Figure 4.11 (Kemp et al., 1998). This should be regarded as only a reconnaissance dataset. Detailed studies can require sample densities of one in 10 square kilometres and at this rate more than a thousand determinations would be required to cover the Falklands. Nevertheless, the available clay mineral data mostly conform to a self-consistent pattern, which is also in agreement with vitrinite reflectance data and, in broad terms, with large-scale geological structures. Assuming a 'normal' geothermal gradient of about 25°C/km, these data can be used to infer maximum burial depths across the Islands. This assumption is justified by measurement of the K- white mica *b* lattice dimension in three samples from the D1 fold belt in East Falkland. The values are typical of fold-and-thrust belts found in medium pressure terranes, where a geothermal gradient of 25-35°C/km would be expected (Kemp et al., 1998).

Vitrinite reflectance (VR) depends on the presence of the organic mineral vitrinite and so is more restricted in its application than clay mineral maturity analysis. However, as it employs a completely different analytical technique, it can provide strong corroborative evidence for clay mineral maturity data. Although some VR data for the Falklands has been published, most of the determinations performed to date have been presented in confidential reports. Currently available VR data for the Falklands broadly corresponds to the clay mineral maturity data, according to correlations found in other areas (Kemp et al., 1998; Merriman and Kemp, 1996; Merriman and Peacor, 1999).

Several large-scale features of the clay mineral maturity map (Figure 4.11) seem significant. Note, however, that the isocryst contours on this map are mainly computer-generated and given the sparseness of the data they should be interpreted with reference to the individual data points.

Nearly all of West Falkland is in the late diagenetic zone, but there are some informative local variations. At Port North [TC 6 9] Marshall (1994b; 1994a) found that VR increased from 1% to 1.4% up sequence, attributing this inversion within the late diagenetic zone to increasing proximity to the D1 fold belt. In the present interpretation it is linked with proximity to the trace of the Sound Bridge Thrust and associated folding (Section 4.2.1.5). The clay maturity data suggest that in general the Port Philomel Formation of West Falkland had undergone burial to a depth of about four to six kilometres before being uplifted to present-day outcrop (Kemp et al., 1998). This implies that in addition to the 1.6 kilometres of younger strata presently exposed on West Falkland, 2.4 to 4.4 kilometres have been removed by erosion. These missing strata presumably belonged to the Brenton Loch Formation and Bay of Harbours Formation at least in part, making the Lafonia Group only about 3.5 to 5.5 kilometres over much of West Falkland. This inferred thickness would be broadly consistent with the approximately 55 per cent reduction in thickness seen in the Port Sussex Formation from east to west. However, if the previous overburden to the Port Philomel Formation also included portions of the Sound Bridge or Pebble Island thrust sheets, then the Lafonia Group would be proportionally thinner.

The determinations of very low metamorphic grade for samples from the Port Stephens Formation in the south of West Falkland suggests a burial depth of only four to five kilometres at South Harbour and three to four kilometres at Cape Meredith. If these are representative, and if the Devonian formations were not significantly condensed in the south of West Falkland, they indicate that the Lafonia Group had reduced to only two to three kilometres at South Harbour (a reduction of about 80 per cent compared to East Falkland) and could have been absent at Cape Meredith. Richards et al.(1996b) infer that the Lafonia Group is locally absent offshore to the south-west of West Falkland, where it has been overlapped by younger Mesozoic strata resting on the West Falkland Group.

Hyam et al. (1997) determined VR for organic matter from sedimentary dykes in the South Harbour area (Section 4.4) considering a value of 0.79% to be representative. They inferred that the area had undergone only about two to three kilometres of burial since mid-Carboniferous times. This is consistent with the prior removal of some two kilometres of Devonian strata, and with the reduction of the Lafonia Group inferred from clay mineral maturity data.

Most determinations for the Lafonia Group from West Falkland are from within the Coast Ridge steep zone where increased strain would produce relatively high metamorphic grades (Kemp et al., 1998). Moreover, burial will have been in part due to tectonic thickening. These data are not used to estimate formation thicknesses because of these uncertainties.

In East Falkland, the area of highest metamorphic grade lies within the D1 fold belt, centred on the Wickham Heights. Burial depths in the order of eight to 10 kilometres are indicated. However, metamorphic grade and apparent burial depth decrease northwards, with even though the age of the strata increases in that direction, in general. This metamorphic inversion is thought to be a consequence of D4 and D5 faulting, and possibly also D1 thrusting. These and other findings modify previous suppositions that metamorphic grade decreases from east to west (Marshall, 1994b; Curtis and Hyam, 1998).

The major change from the late diagenetic zone in the west to the anchizone in the east occurs close to the Goose Green Axis, not to the Falkland Sound Fault Zone. Low Kubler index values from the Port Stanley Formation near Curlew Creek are attributed to relatively localised deformation in the Port San Carlos steep zone, consistent with the ductile deformation at Settlement Rocks (Section 4.2.1.4). Near Port Sussex, the grade increases from the diagenetic zone to high anchizone within a few kilometres at most. The clay mineral maturity data thus support the interpretation of the Goose Green Axis as a major tectonic lineament, at least for most of its length. Furthermore, to the east of the Goose Green Axis, metamorphic grade decreases towards the east. Also, as indicated by the intensity of S1 cleavage in the outer Jason Islands, the metamorphic grade there is higher than anywhere else, attaining the epizone (Figure 4.11).

The marked increase in metamorphic grade parallel to regional strike between Carcass Island and Grand Jason confirms that a large differential uplift occurred there following D1. This uplift could have been as much as four kilometres (Kemp et al., 1998). From structural evidence, this is most likely to have occurred on the South Jason Line (Figure 4.1) during D2/D3. Note that the three isocryst contours shown between Grand Jason and Carcass Island on Figure 4.11 were interpolated by computer and that the full extent of epizonal metamorphism in the Jasons is unknown. It seems likely that a much steeper metamorphic gradient occurs in the vicinity of the South Jason Line.

Conversely, the configuration of the metamorphic zones in East Falkland suggests that an area east of the Goose Green Axis has been uplifted more than to the west, as inferred from D5 deformation (Section 4.2.5.1). The D1 strain partitioning postulated to have occurred on the Goose Green Axis (Section 4.2.1.4) might be thought to account for the metamorphic gradient, instead of or as well as uplift, but eastern Lafonia south of the D1 deformation front (Section 4.2.1.2) has also been uplifted relative to the west by at least two kilometres (Kemp et al., 1998). The decrease in grade from high to low anchizone which occurs across regional strike in eastern East Falkland implies that the degree of uplift decreases to the south-east, imparting a regional tilt. Burial depths of more than six kilometres indicated near the top of the Bay of Harbours Formation in the southern part of Lafonia suggest that the Jurassic and younger sequences seen offshore (Richards et al. 1996) once extended that far onshore. The uplift cannot then be due to D3 movement, which pre-dates the Early Jurassic dykes, and so is assigned to D5.

This uplift of eastern East Falkland is contrary to the sense of up-to-the-west regional uplift deduced for the D2 deformation (Section 4.2.2). From structural considerations, West Falkland clearly underwent a considerable uplift during D2 to form the Coast Ridge steep zone. The thickness of exposed strata east of the Hornby Mountains Anticline shows that this uplift must have been at least four kilometres and perhaps as much as seven kilometres, consistent with the four to six kilometres of burial inferred for the Port Philomel Formation on West Falkland. An area of the steep zone near Port Howard is in the low anchizone, as shown by clay maturity data (Figure 4.11) and a VR value of 2.1% (Marshall, 1994b; Marshall, 1994a). The presence of low anchizone metamorphism close to the core of the Hornby Mountains Anticline can be attributed to this uplift.

The proposed reversal in the polarity of regional uplift during D5 is consistent with up-to-the-east movements found on the Blue Mountain Fault and Main Carcass Bay Fault (Section 4.2.2.1), and on northerly trending D5 faults in the north of the Islands (Section 4.2.5). The distribution of amygdaloidal dykes (Figure 3.1), which are confined to the far west of the Islands, also suggests that up-to-the-east uplift took place after Early Jurassic dyke emplacement. In summary, it appears that whereas D2 deformation involved up-to-the-west movement of major structural blocks, D5 saw the reverse.

4.4 Sedimentary dykes

Sedimentary dykes are a minor but widespread and rather unexpected component of Falkland Islands geology. Although none are large enough to show on the geological map, they can yield useful information and may be the only local representation of some periods of the Earth's history. Whereas igneous dykes (Section 3.1) are the result of intrusion of hot molten rock into a vertical fracture, sedimentary dykes formed when a fissure was filled by unconsolidated sediment. Small intraformational sedimentary dykes can form by processes such as de-watering and are found sporadically in parts of the Fox Bay Formation and in the Permian sequence. This section considers only sedimentary dykes which formed after the host formation had become lithified.

The most closely studied examples occur in the South Harbour area. Baker (1924) reported that diamictite (assumed by him to be a tillite – Section 2.4.2) occurs together with minor shale near South Harbour. Hyam et al. (1997) showed that these lithologies occur in a series of subvertical sedimentary dykes and are not part of the Port Stephens Formation sequence. The following account is taken largely from their description, together with some observations made during the present survey.

Eleven sedimentary dykes seen near South Harbour House [TC 43 31] by Hyam et al. (1997, fig. 2a), together with one additional one found during the present survey west of Signboard Hill [TC 3989 3268], are in the range 0.2 metres to 3.4 metres wide, with straight subparallel walls formed of gently dipping sandstones of the South Harbour Member (Section 2.3.1.5). All are subvertical, and most trend roughly north-south, with a few at intermediate orientations. The exposed vertical extent of each dyke ranges from one to ten metres. All are exposed in the foreshore, although one in south-west Anthony Creek can be traced across an inlet and apparently continues through a steep-sided gully some 200 metres to the north-west. All twelve recorded sedimentary dykes occur within only four kilometres of South Harbour House and it is probable that more are present in the surrounding area. Other examples are known on Dyke Island (M J Hole, oral communication, 1998), and at Island Harbour, Weddell Island.

The rock forming the dykes is typically a grey to grey-brown poorly sorted muddy sandstone or diamictite with scattered granules or small pebbles up to five centimetres in diameter. It is broadly similar to massive diamictites in the Fitzroy Tillite Formation. Most pebbles are composed of vein quartz but some are of quartzite, and there are few clasts of igneous or metamorphic rock types. One example of a mudclast was seen. Vertical lamination is common, especially close to the margin of the dykes. In some dykes there are also subvertical partings of mudstone or muddy silt from a few millimetres to five centimetres thick at or close to the margins. Vitrinite reflectance data from these dykes is discussed in Section 4.3.

Hyam et al. (1997) concluded that the sedimentary dykes near South Harbour had been emplaced rapidly by downward injection of fluidised sediment, rather than by the passive infill of open fissures. This is consistent with formation beneath a glacial ice sheet. They found that two of the dykes in this group contained a mid-Carboniferous palynomorph assemblage and concluded that they had been emplaced sometime from the late Viséan to the early Namurian. The dykes thus preserve a deposit from a period which is not represented in the West Falkland Group or the Lafonia Group, and only poorly elsewhere in central Gondwana.

One of the two Carboniferous sedimentary dykes is exposed on the shore of Anthony Creek, about 950 metres ENE of South Harbour House [TC 4422 3195]. It apparently cross-cuts and dextrally displaces a narrow dolerite dyke by some ten metres horizontally, although the dolerite dyke is assumed to be of Early Jurassic age (Section 3.1). It seems likely that the displacement of the dolerite dyke was caused by later slip along the eastern margin of the sedimentary dyke.

Dawson (1967) describes a sedimentary dyke within the Bluff Cove Formation exposed on the foreshore close to Bluff Cove settlement [VC 1945 6613]. According to Hyam et al. (1997) this is subvertical, oriented north-south, and is about 0.8 metres thick. Hyam et al. (1997) also describe a single NE-SW sedimentary dyke which crops out on the south shore of Monty Dean's Creek, near Green Patch [VC 2080 8634], where it intrudes the Fox Bay Formation. It is between 35 and 40 centimetres thick. Two similar dykes, each about five centimetres thick, occur some 25 metres to the east. All are filled with greyish brown sandy diamictite similar to that found in the South Harbour dykes. Hyam et al. (1997) found no palynomorphs in these dykes but concluded from their field relationships that they are likely to have been formed in the Late Carboniferous or Early Permian, during the deposition of the Fitzroy Tillite Formation. The dyke at Monty Dean's Creek is cross-cut by a faint planar fabric parallel to S1 (Section 4.2.1.1), which is consistent with this interpretation.

Other examples have been found at widespread localities. Narrow sedimentary dykes occur in coastal exposures of the Fox Bay Formation in the White Rock Bay area, about 400 metres west of Old House Bay [UC 413 995] and on the south side of Big Rincon [UC 423 977] (M. Hunter, oral communication, 1998). Several narrow sedimentary dykes composed of yellowish brown siltstone and fine-grained sandstone occur in the Port Philomel Formation at the Neck, Saunders Island [TD 7430 1083]. These infill fractures associated with northerly-trending faults, and range from five to 30 centimetres across. Cliff-top exposures on New Island show that the few dolerite dykes which occur there are very deeply weathered, so that it is usual for intact dolerite to be visible only 20 metres or more below the adjacent cliff-top. Where not covered by vegetation, the material overlying the weathered dolerite appears to be poorly consolidated sparsely pebbly clayey sand. This deposit varies from dark grey at depth to brown closer to the cliff-top, indicating that it is of sufficient age to have been exposed by erosion of the cliff-line. It is possibly a submarine deposit dating from a time of higher sea level.

It cannot be assumed that these other sedimentary dykes are either mid-Carboniferous or Permo-Carboniferous in age, or that they all formed in the same manner. Some may be the result of Quaternary weathering and superficial deposition.

4.5 Geophysical surveys

Geophysical information from the Falkland Islands onshore area is very sparse and so far it has made little contribution to the geological interpretation.

Short gravity surveys were carried out by McNaughton (1972) and Martin and Sturgeon (1982), providing data for a total of 38 stations close to sea level around the Islands. The base station at Stanley established by McNaughton (1972) and McGibbon (1988) was relocated to Sapper Hill [VC 38 71] during 1999. The onshore gravity data is too sparse to allow a quantitative interpretation. However, it is notable that the largest Bouguer anomaly value was obtained on Weddell Island, within the area of the Cape Orford Dyke Swarm and close to an inferred volcanic centre (Section 3.1).

The gravity anomaly map shown as an inset on the East Falklands geological map sheet is based primarily on the free-air anomaly values derived from satellite altimetry and released into the public domain by the Scripps Institute of Oceanography as version 7.2 (Sandwell and Smith, 1997). The satellite gravity data are not valid over land and the limited number of available Bouguer gravity anomaly values from stations observed on the Falkland Islands themselves (Martin and Sturgeon, 1982; McGibbon, 1988) were used to provide control here. Some marine gravity data has been collected opportunistically during scientific cruises near the Islands, especially to the east (Barker, in press). Detailed marine gravity surveys have been undertaken as part of hydrocarbon exploration activity in the 1990's but these remain confidential at this time and, because of their restricted extent, they do not significantly enhance the appearance of the gravity anomaly field at the small scale used.

The satellite data as displayed have been filtered by upward continuation to reduce the influence of track-related noise which appears as a 'dimpling' effect over the image as a whole. The continuation distance of two kilometres effectively raises the observation level from sea level and thus attenuates anomalies of higher frequency (or longer wavelength). In practice, anomalies with a wavelength of less than 20-30 kilometres cannot be resolved reliably from the satellite data even though the satellite tracks from which they are derived are typically only five to ten kilometres apart.

Free-air gravity anomaly values are not corrected to allow for variations in the depth of sea-water. As the density contrast between water and rock is relatively large, large effects can be observed on the margins of the continental shelf and over specific bathymetric features (compare gravity and bathymetry insets on the East Falklands map sheet) where the relief changes rapidly. Where the sea-bed relief is more subdued the anomalies are more obviously of geological origin: gravity anomaly lows may be associated with sedimentary basins or granites, whereas anomaly highs suggest elevated basement blocks or the presence of basic igneous intrusions. The effects of topography on gravity measurements taken on land are more of a problem in distorting the geological component as the Bouguer correction is usually applied to take account of the varying thickness of rock and the change in height between the observation point and sea level datum. As the station elevations on the Falkland Islands are small, the corrections here are not critical.

An interpretation of the offshore gravity anomaly data was not within the scope of this project but it is clear that they do contain valuable information relating to the regional tectonic setting (e.g. Barker, in press).

A ground magnetic survey of the Stanley area was carried out by Ashley (1961), also reported by Mansfield (1965). No clear relationship between the magnetic field and known bedrock features can be made. Although there is some deflection of the magnetic field in the vicinity of the fault passing west of Sapper Hill, other disturbances are apparent in areas where no faulting is suspected.

Two brief reconnaissance aeromagnetic surveys have been carried out in the Falkland Islands. Six north-south lines were flown over the northern part of East Falkland in April 1997. The lines show very few high frequency anomalies, and little detail which can be related to known geological features. They are too widely-spaced to allow useful quantitative interpretation. A single extended aeromagnetic traverse across both main islands was flown in May 1997. This dataset remains confidential.

Marine magnetic and gravity data from the east of the Islands has been interpreted by Barker (in press) (Section 3.1.1.7). Some marine magnetic data was acquired during 1997 for hydrocarbons exploration but this remains confidential. Southby-Tailyour (1985) reports the occurrence of local magnetic anomalies in several places close to the Falklands coast line, possibly caused by sunken wrecks.

5. CENOZOIC GEOLOGY

As noted in Section 2.1, a distinction is commonly made between the 'solid geology' of an area, which broadly corresponds to 'bedrock', and the 'superficial geology' (or 'drift' geology), which encompasses material lying between the soil and the bedrock. In the Falklands nearly all the known superficial deposits were formed during the late Quaternary and are less than about 50 000 years old, but one onshore deposit, the West Point Forest Bed, dates from the Neogene. The apparent absence of deposits of any intermediate age, especially from the Pleistocene, remains problematical, but might be a consequence of late Pleistocene uplift amongst other factors (Sections 5.2.5 and 5.2.10).

A map of the superficial deposits was compiled at 1:250 000 scale during the Falkland Islands Mapping Project. It remains unpublished, but can be consulted at the Department of Mineral Resources, Stanley. It is also available in digital form on CD-ROM (Appendix 1).

5.1 West Point Forest Bed

In 1899, the then owner of West Point Island, Mr Arthur E Felton, observed the presence of numerous tree trunks on a sheltered beach close to the settlement. He drew them to the attention of a visiting geologist, Halle, in 1907. Halle (1912) discovered that the tree trunks had been eroded from a bed of organic clay which occurs intermittently along at least 600 metres of coastline and up to 30 metres inland. This underlies a stony sandy clay (diamicton) of similar aspect to the solifluction deposits (Section 5.2.1) which are widespread elsewhere in the Islands, although on West Point Island an admixture of well-rounded and angular boulders suggests that older raised beach deposits have been incorporated with debris derived from weathered bedrock. Parts of the overlying diamicton are organic rich or sulphatic. Pieces of tree trunk more than about two metres in diameter occur, together with twigs of five millimetres or less. Some have been pyritised. Halle (1912) found that the Forest Bed includes remains similar to two gymnosperm species which are endemic to South America: *Austrocedrus chilensis* and *Podocarpus salignus*. At the present day neither species grows south of latitude 45° S in southern Chile (Moore, 1968; Veblen et al., 1995).

Although Baker (1924) argued that the tree trunks from the West Point Forest Bed represent a recent accumulation of driftwood, as is seen in certain south-west facing coves elsewhere in the Islands, its position beneath solifluction deposits clearly shows that this is not the case. Moreover, Halle (1912) noted that the tree trunks within the deposit showed no sign of marine erosion, commonly retaining their bark, and inferred that the Forest Bed is a Quaternary deposit. However, no trees occur in growth position. Birnie and Roberts (1986) confirmed that the wood is preserved in an indigenous terrestrial deposit. They extracted a palynomorph assemblage dominated by podocarp-type pollen, including material comparable to *Dacrydium*, *Dicksonia* and *Nothofagus fusca,* which they inferred to be of local origin. They point out that podocarp-type pollen has not been found in post-glacial deposits in the Falklands (Section 5.2.7). They concluded that the Forest Bed is much older than suspected by Halle (1912), suggesting that it records a Tertiary, possibly early Oligocene, forest environment in the Falkland Islands.

Palynomorphs from the upper part of the Forest Bed were discussed by Macphail and Edwards (in prep) who argue that the deposit is no older than Middle Miocene and no younger than the Neogene. They suggest from negative evidence that the assemblage is consistent with deposition during the Early Pliocene.

Although the Forest Bed may have been moved down-slope by solifluction from its original site, it indicates that some part of West Point Island was emergent during the Pliocene. It is quite likely that Tertiary or earlier Quaternary deposits elsewhere in the Islands have been incorporated within the mantle of solifluction deposits (Section 5.2.1), although remnants could be preserved beneath them in places.

5.2 Quaternary superficial deposits and associated landforms

5.2.1 Solifluction deposits

Solifluction is the slow down-slope flow of soil and superficial debris. It occurs in ground which is thawing after having been frozen. Solifluction typically causes movement of a few centimetres a year and can take place on slopes as gentle as 1° (Washburn, 1979; Ballantyne and Harris, 1994, p. 114). Solifluction does not imply the presence of permafrost, but does require a colder climate than presently experienced in the Falklands. The term 'solifluction' was coined by Andersson (1906) following his research in the Falkland Islands, and on Bear Island (Latitude 74.5° N) where solifluction is currently active.

Solifluction deposits (or periglacial slope deposits) are the product of gradual down-slope mass-movement by solifluction. (In the UK they are usually known as 'head'). They are the most widespread category of superficial deposit in the Falkland Islands. They are almost ubiquitous on steep or high ground underlain by the Port Stanley Formation or the Port Stephens Formation, including the more hilly offshore islands such as New Island and Grand Jason. Solifluction deposits appear not be significant in most low-lying areas, especially those underlain by the Fox Bay Formation, the Fitzroy Tillite Formation and younger bedrock formations.

The deposits are most commonly made up of pale grey unsorted stony sandy silty clays (diamictons) and stony sandy silts. They often contain subangular quartzite boulders similar to those seen in stone runs although without any superficial rusty staining (Section 5.2.2). Clast-supported accumulations of large blocks with a fine-grained matrix may occur in places, but the solifluction deposits generally are thought not to intergrade with the stone runs. Deposits of this kind underlie the soil in most parts of the Stanley area, for example, and are commonly seen beneath the peat in road cuttings and trenches. They represent the collective weathering products of the local bedrock, probably together with elements of any pre-existing superficial deposits, such as raised beach deposits (Section 5.2.10). It has been argued that the solifluction deposits include *remanié* elements of Tertiary deposits, but this is not proven (Clapperton, 1975; Clark, 1976).

In most areas, the solifluction deposits form a debris mantle without any characteristic landforms, which almost completely covers the local bedrock. Topographic features which are evidently related to stratigraphic boundaries or structures in the bedrock are usually visible despite this overburden, implying that the solifluction debris mantle is generally of slight and uniform thickness. In most places it probably does not exceed about three metres in thickness, although direct evidence for this is sparse. In some areas, however, bedrock features are completely obscured. The greatest thickness observed in exposures of the solifluction deposits is about 10 metres (R. Clark, written communication, 1999). Where the solifluction deposits are thicker, surface landforms have developed. The most usual are turf-banked or stone-banked solifluction lobes and terraces (Plate 33), formed by a more mobile part of the deposit over-riding another down-slope (Washburn, 1979; Ballantyne and Harris, 1994). The downhill margin of these features is typically several metres high but can be as much as 10 metres (Clark et al., 1994). Clark (1972) observed that solifluction terraces occur between 60 metres and 450 metres above sea level (ASL) on slopes at angles of between 8° and 12° and of all aspects. The terraces follow the contours or lie slightly oblique to them, and are commonly several hundred metres in length. Many are simple, with straight edges, but some are compound and lobate. Good examples occur on the slope rising west of Colarado Pond [UC 97 68] and on the eastern flanks of Mount Kent [VC 25 74]. Roberts (1984) gives further details of these phenomena.

Clark (1972) also notes that the solifluction terraces on the northern flank of Mount Usborne are interrupted by irregular depressions, some discontinuous, which pass down-slope. They appear to be contemporary with the terraces, and are possibly caused by meltwater from snowfields on the higher ground. They are probably related to a series of narrow elongate ridges, lying oblique to the line of slope, which have developed in the solifluction mantle above the valley which contains the headwaters of Arroyo Malo and Dougherty's Brook. These appear on aerial photographs to be depositional landforms rather than erosional ones.

Although there is no evidence that solifluction is occurring at the present day in the Falklands, minor frost-induced weathering still occurs, especially at high levels, although it has been noted at sites as low as 35 metres ASL (Wilson and Clark, 1991). This is associated with frost-sorting which forms small-scale patterned ground (Clark, 1972; Clapperton and Roberts, 1986; Clark et al., 1994; Wilson and Clark, 1991). Some forms of patterned ground which are not frost-related are described by Wilson (1995).

Plate 33: Solifluction terraces

East flank of Mount Kent. Photograph by Don Aldiss, BGS. MN28263

Plate 34: Stone runs (stone stream type)

South flamk of Mount Usborne. Photograph by Don Aldiss, BGS. MN28260

Most of scene is underlain by Port Stanley Formation. Base of Bluff Cove Formation is marked by negative break of slope below patch of dark-coloured vegetation in centre of picture. Foreground is underlain by Fitzroy Tillite Formation, exposed in bank on left side of the picture (in shadow).

The mantle of solifluction debris is commonly associated with stone runs of all types (Section 5.2.2), with which it is co-eval. Deposits which are the product of rapid mass-movement, including land-slips and debris flows (Section 5.2.3), occur locally in association with solifluction deposits. They are overlain by upland peat deposits (Section 5.2.7). Many valley bottom deposits probably include a component of solifluction debris, underlying or possibly interbedded with alluvium or peat, although some upland streams have dissected solifluction deposits. Some solifluction deposits extend below sea level. Dispersal of the fine-grained matrix by wave action can then leave a bouldery lag deposit which could be mistaken for a stone run (Clark, 1972; Roberts, 1984).

Radiocarbon age determination of peat (Section 5.2.7) indicates that the solifluction deposits mostly formed during the last glaciation, between about 13 600 years and 26 000 years BP (before present). There is some evidence for older solifluction deposits (Birnie and Roberts, 1986; Clark et al., 1998) but it is likely that most superficial deposits on hills were remobilised and reworked by solifluction during the last glaciation.

5.2.2 Stone runs, blockfields and scree

Stone runs are accumulations of boulders with no finer material and virtually no vegetation, which occur on slopes or valley bottoms at all levels. They are relict landforms, the result of mass-movement due to past periglacial processes in frost-susceptible sediments. They everywhere occur in association with solifluction deposits and have flowed down-slope together with them, as part of a near-continuous debris mantle which covers the larger hills in the Falklands. The smallest stone runs are only about a metre across, whereas the largest are up to four kilometres long and several hundred metres wide. They are most common on outcrops of the Port Stanley Formation but also form on the Port Stephens Formation. Some extend down-slope onto adjacent formations (Plate 34). They are most widespread and voluminous on East Falkland, where the folded quartzite beds of the Port Stanley Formation form numerous exposures and a broad outcrop, but also occur on West Falkland and the larger of the other islands including Weddell, Saunders, and Keppel.

The local term 'stone run' has been generally used in scientific literature about the Falklands since Baker's (1924) report. Other names have been used in the past and in other parts of the world. Some stone runs which veneer moderate or steep slopes can be classified as blockslopes, and those in valleys are blockstreams (Washburn, 1979; Ballantyne and Harris, 1994). Some other Falklands stone runs form stone stripes, a variety of sorted patterned ground (Ballantyne and Harris, 1994).

There are some apparently similar types of deposit, some closely related to stone runs, some less so. Blockfields are areas of level or gently sloping ground largely covered by angular or subangular boulders (Washburn, 1979; Ballantyne and Harris, 1994). They differ from stone runs as described here in that they represent detritus of weathered bedrock which has not undergone significant lateral movement. In the Falkland Islands, blockfields occur on the summit of the higher hills underlain by Stanley Quartzite (Clark, 1972) but are not particularly extensive compared with stone runs. Most if not all probably pass into high-level stone runs (see below), from which they are not clearly distinguishable on aerial photographs.

Scree (or *talus)* is clast-supported rock debris which has fallen from steep rocky exposures. Scree thus differs fundamentally from mass-movement deposits formed by solifluction or landslip. Small scree deposits occur locally on the larger steep rocky faces of Stanley Quartzite. They are usually associated with the far more voluminous solifluction deposits and may have become partly incorporated within them, or with stone runs. Scree formation occurred under periglacial conditions and no active screes are observed.

Stone runs should not be confused with debris flows or landslips (Section 5.2.3), which are deposited rapidly; glacial deposits, which are of very limited distribution in the Falklands (Section 5.2.5); or relict rock glaciers (Sissons, 1975; Barsch, 1996), which do not occur in the Falklands.

There is a considerable literature on the Falklands stone runs although many of the published descriptions are rather brief and derivative. The more useful and interesting accounts are given by Moody (1842, reprinted in Falkland Islands Journal, 1969), Darwin (1845), Davison (1889), Andersson (1907), Baker (1924), Joyce (1950), Greenway (1972), Clark (1972), Clapperton (1975), Roberts (1984) and Clark and Wilson (1997). References to other reports are made in these papers. The following account is based both on this literature and on observations made during the present survey (including aerial photograph interpretation).

Many characters are shared by all the stone runs, regardless of size, form or situation. They are almost entirely composed of locally derived quartzite blocks, with exceedingly rare examples of dolerite or diorite (Section 3.1) which are presumably also of local origin. Most blocks seen at the surface are between 30 centimetres and two metres across, and rarely up to five metres long. The range of block size can vary locally, however. For example, two block streams meet low on the southern flank of Mount Usborne (Plate 34). Blocks in the main, western stream are of the usual size range, while those in the smaller eastern stream are mostly between 10 and 50 centimetres across, with some up to 1.5 metres. The size of the smallest blocks in a stone run is probably related to the average block size, here thought to be a consequence of the size of the spaces left between the blocks while the stone run was developing.

Some blocks in stone runs are equant, but most are tabular or elongate, their shape and size reflecting the common distribution of joints and bedding planes in nearby quartzite exposures. The blocks tend to be angular to subangular, with rounded edges and corners but little other sign of abrasion, although some of the smaller blocks are subrounded. Many stone runs display a marked fabric in which the tabular blocks are packed together on edge (Plate 35). In many places, these blocks tend to be oriented parallel to the slope, especially near the margins of the stone run. Elsewhere they are aligned across the slope and then are typically imbricated, dipping up-slope. A few blocks, including some of the very large ones, were not incorporated within the fabric and these lie randomly on top of the stone run. Some blocks tilt when stepped on. Some stone runs have no clear fabric. Excavations in stone runs (seen by the road near Mount Kent, and on the south side of the Sussex Mountains) reveal weak reverse grading to cobble size, but normally no gravel, sand, clay or peat is seen to occur within the stone runs. Bellosi and Jalfin (1984a) carried out morphometric analysis of the stone run south-east of Blue Mountain [UC 820 640]. They found general downstream trends of decreasing diameter and flatness of blocks, and of increasing sphericity and roundness. Other details are given by Roberts (1984).

Although undisturbed stone runs are uniformly pale grey in colour, excavations show that where blocks have been protected from the weather within the stone run they are invariably stained reddish brown, orange or pink by iron oxides. Where stone run blocks have been broken the rusty colouration can be seen to spread through the rock. It is attributed to prolonged chemical weathering in an oxidising environment, with traces of iron presumably derived from the weathering of feldspar and mica within the quartzite. Its absence at the surface of the stone runs is attributed to prolonged leaching by rainwater. The basal part of the stone run can be stained blackish brown by precipitation from running water. Note that similar quartzite blocks enclosed within the solifluction deposits are entirely grey in colour, due to reducing conditions.

On slopes, stone runs generally occur only in lateral contact with solifluction deposits, and not against other types of superficial deposit such as peat, and not in isolation. They are, however, locally overlain by peat (Clark, 1991). Where stone runs occupy valley-floors they are generally contiguous with valley-bottom deposits of other kinds either downstream or upstream. Parts of such stone runs appear to be buried by waterlain deposits. This is seen in the Moody Valley (Rosenbaum, 1996), for example, and in the upper portion of Green Cottage Brook, just north of The Onion mountain. Other examples have been observed by R. Clark (written communication, 1999). One example of a debris flow post-dating a stone run is described in Section 5.2.3. Rosenbaum (1996) suggests that raised marine clays (Section 5.2.10) overlie stone runs near Stanley, but no other evidence for this relationship is known.

Stone runs usually lie at some distance from rock exposures and some occur where no bedrock can be seen. Very few can be traced directly to a potential source exposure although some seem to merge with scree deposits. Some extend upslope to ridge crests and summit areas, where they merge with blockfields. There is no association with open scars at the upslope end of stone runs, from which they might have arisen as a landslip. Some areas of hillside have only a few patchy stone runs surrounded by solifluction deposits, and many stone runs enclose isolated vegetated patches, with all intermediate proportions. The smaller stone runs commonly form regular stripes running down-slope, typically alternating with vegetated stripes of similar width (Plate 36). Stone runs can pinch out or divide down-slope, but commonly amalgamate with their neighbours to form progressively larger streams.

The lateral margins of the stone runs are commonly linear and curve to follow the line of greatest slope. The upper and lower margins can be more irregular, tending to interfinger with adjacent vegetated areas. In some stone runs, including some of the largest, the lower margin tapers down-slope to a point as wide as a single block (Plate 34). Others form a broad horizontal front, especially where there is a break of slope. However, all margins are abrupt, or with a narrow transitional zone in which blocks are visible through the grass or dwarf shrubs on the adjacent ground. If present, this marginal zone is usually less than one metre across but can be as much as ten metres wide at the

Plate 35: Alignment of blocks in stone run

Keppel Island. Photograph by Don Aldiss, BGS. MN28261

Plate 36: Stone runs (stone stripe type)

Near Bluff Cove. Photograph by Don Aldiss, BGS. MN28262 termination of large stone runs. There is usually little if any encroachment by vegetation onto the stone runs, although Rosenbaum (1996, fig. 3) records an exception. Excavations across the margins of stone runs (as near Mount Kent) show a well-demarcated, steeply dipping contact between the blocks forming the stone runs and the clay-rich solifluction deposits (with scattered, matrix-supported blocks) which underlie the adjacent vegetated ground. Thus it seems that in most instances the deposit underlying the vegetated areas does not intergrade with the stone runs. This is contrary to the commonly-expressed opinion (Andersson, 1907; Baker, 1924; Joyce, 1950; Rosenbaum, 1996) that many of the vegetated areas between stone runs are essentially 'over-grown' stone runs. In excavations, the base of the stone runs seems to rest abruptly on solifluction diamicton, although Clark (1972) records a natural exposure in which a stone run lies on grit, sand and silty mud. Dodds (1969) notes rare examples of stone runs directly overlying bedrock quartzite.

Stone runs commonly lie on slopes of up to 10° on valley sides, or as little as 1° in valleys. The surface of the stone runs is nearly always approximately level with the adjacent vegetated ground. There is some suggestion that stone runs near hill tops are more uneven than those lower on the slopes (Dodds, 1969), but there are usually no hummocks, ridges or terraces within individual stone runs, or at their down-slope termini, as seen in solifluction lobes or rock glaciers. Separate stone runs do not pile up against each other to any great extent, although differences of up to a metre can occur in the levels of large confluent stone streams. In common with the solifluction deposits, however, stone runs drape over bedrock features, apparently maintaining a uniform thickness which is insufficient to obscure the rockhead topography. This is well displayed on the southern flank of Mount Usborne [UC 720 675] (Plate 34). Conversely, small stone runs can pass over (or through) solifluction terraces up to five metres high without interruption. This relationship is also seen at the Stiperstones (Shropshire, UK) (Clark, 1994, p. 469). Running water can be heard within many stone runs, commonly emerging into small streams at the down-slope end. Clark (1994) found that in places there are multiple narrow subsurface channels, some of which are braided. This water can rarely be seen but usually sounds as if it is within one to three metres of the surface. Excavations and natural exposures suggest that this is the usual range of thickness of stone runs.

Shallow longitudinal declivities appear within some large stone runs, or at one margin. They coincide with streams within the stone run, as seen low on the southern flank of Mount Usborne, and in Green Cottage Brook in the Malo Hills [UC 881 810]. A stone run 'south of Port Salvador' has a broad central tract of surface boulders rounded by water erosion but now lichen covered. The channel width is more than would be expected for normal run-off (Clark, 1972).

Few stone runs reach sea level. Of those that do, some pass beneath the water with no apparent break. This can be seen on the east side of Bold Cove, where the stone run has been partly reworked by the sea. By contrast, the prominent stone run on the south-west side of Keppel Island stops abruptly about two metres above sea level and about five metres from the high tide mark, forming a low rocky embankment. The adjacent beach is a narrow shingle strip with no large boulders, passing offshore to sand. This abrupt termination is unlike that noted at any other stone run. It is taken to mark a previously higher sea level and is thought to show that active stone runs were disrupted by sea water.

The Falklands stone runs can be classified by their surface form and the geomorphological situation in which they occur. This newly proposed classification is based on a comprehensive study of aerial photographs of the Falklands, including some taken with hand-held cameras, and on field observations in all parts of the Islands.

Stone patches are mutually isolated concentrations of boulders on slopes. They are the most common type. The patches vary from a few metres to more than one hundred metres across, and may be irregular, equant or elongate in overall shape. Elongate stone patches may be termed '*stone streams*' (Plate 34). The larger patches include vegetated areas and they commonly pass laterally into stone stripes. Some stone patches would be classified as blockslopes (Ballantyne and Harris, 1994).

Stone terraces and *stone fans* are aligned with transverse breaks of slope. Terraces are elongate, whereas fans are more equant. The use of these terms is not intended to imply a similar origin to structures such as river terraces and alluvial fans, etc., but to be merely descriptive. Both tend to have linear down-slope margins but can pass into stone patches or stone stripes in any direction (Rosenbaum, 1996, fig. 2). Some terraces are largely made up of stone stripes. Good examples occur on the south flank of Mount Low [VC 42 78], and on the north side of Prince's Street [VC 22 81].

Stone stripes are narrow linear parallel stone runs alternating with stripes of vegetated ground of similar width (Rosenbaum, 1996, figs. 2 & 3) (Plate 36). Some sets of stripes merge laterally into other forms of stone run, up-slope or down-slope, but some are isolated. Some stone stripes form a very regular pattern. Those seen beside the main road near Bluff Cove are between one metre and five metres wide, while Clapperton (1975) notes stone stripes ranging from three metres to 15 metres wide.

Stone rivers are large valley-bottom accumulations of boulders, formed by the down-slope amalgamation of other types. They correspond to blockstreams (Ballantyne and Harris, 1994). It is possible to distinguish between 'ragged' and 'smooth' stone rivers, based on their appearance on aerial photographs. The former are the more usual type and have numerous tributary stripes and patches along one or both sides which confer a 'ragged' appearance. The latter have no such tributaries and their 'smooth' margins run approximately parallel to the valley. This occurs where the stone river has moved beyond any local sources of boulders to feed the tributaries. Although stone rivers may be several kilometres long, most if not all of the blocks have travelled a much shorter distance from the nearby slopes. Stone rivers are more voluminous than stone streams, they have more tributaries and they generally lie on gentler gradients.

The largest stone river and the best known example is Princes Street, between Estancia and Berkeley Sound. It is about four kilometres long and roughly 400 metres wide in places. It is superbly illustrated and mapped by Andersson (1907; photograph reproduced by Joyce, 1950, fig. 1). Other very large examples occur in the Wickham Heights (e.g. Rosenbaum, 1996, fig. 1) and stone rivers three kilometres or more long occur north-east of Mount Adam and north-west of Mount Edgeworth, West Falkland.

High-level stone runs occur above about 300 to 350 metres on some hills, including Smoko Mountain, Mount Simon, Mount Usborne, and Mount Rosalie. As seen on aerial photographs, these 'high-level stone runs', appear to have a different texture to those on the lower slopes, being much finer, more closely spaced, and possibly more diffuse. Observations by Dodds (1969) on Mount Usborne suggest that they are more uneven and 'jumbled' than those at lower levels. It seems probable that they intergrade with completely undifferentiated mobilised hill-top debris. A line can be usually be drawn between the high level stone runs and the more usual types on lower ground nearby, some of which lie on slopes of a similar angle. On the south side of Mount Simon, for example, it appears that the high-level material is over-riding stone runs lower on the slopes. The high-level stone runs either formed later and are less differentiated than the usual stone patches, or they are the product of a different process, perhaps having formed under lower temperatures.

The Falklands stone runs seem to be unique in terms of their variety, size and abundance, but examples are also known from elsewhere in the World. Goudie and Piggott (1981) and Clark (1994) describe stone stripes and boulder nets on the Stiperstones, Shropshire, England. The stone stripes are of similar dimensions to those of the Falklands, formed of quartzite blocks with long axis lengths between 0.2 metres and 2.5 metres. They note that the long axes of the blocks are frequently aligned down-slope and that individual slabs are often arranged on edge rather than lying flat. Large stone stripes in North Wales are described by Ball and Goodier (1968). The pattern formed by each stone stripe and the adjacent vegetated stripe repeats every five to eight metres. The blocks are composed of sandstone and conglomerate and are generally between 0.6 and 1.5 metres across. In both instances, the stripes have been proposed as the result of stone sorting under periglacial conditions (Ballantyne and Harris, 1994, p. 98), although Clark (1994) questions this, proposing that they can be explained by the development of a drainage system on the debris sheets and consequent washing-out of fine-grained debris. The Hickory Run boulder field, Pennsylvania, USA, is irregular in outline, and about 120 metres by 550 metres in extent. The blocks are composed of locally derived quartzite and conglomerate (Smith, 1953). DTA has observed small stone-runs formed of granite blocks in the Cheviot Hills, England. Greenway (1972) refers to other examples.

Numerous theories have been proposed for the origin of the Falklands stone runs but it is now generally accepted that (in common with the solifluction deposits, and with stone stripes and allied phenomena elsewhere) the stone runs formed during the last glaciation as a result of intense frost action alternating with periods of thaw. In understanding their formation, it is important to realise that stone runs moved down-slope as part of a near-continuous solifluction debris mantle. Stone runs did not move by themselves like glaciers or rivers. Their well-developed sorting, internal fabrics, overall form and relationship to the solifluction deposits, together with their local extension onto the Fitzroy Tillite Formation, shows that they are not the result of in situ weathering (Joyce, 1950). However, some questions about their formation remain (Clark and Wilson, 1997).

At least five processes seem to have been involved in the formation of stone runs: weathering, solifluction, frost-heave, frost-sorting and washing. Both the Port Stanley Formation and the Port Stephens Formation include feldspathic sandstones, with some siltstones and mudstones. These rocktypes would be broken down by frost-shattering and chemical weathering to sand, gravel and clay, with blocks of the very hard quartzites. As with the solifluction deposits, there may also have been some pre-existing clay-rich superficial cover which became mixed with this debris (Section 5.2.1).

As this unconsolidated material was subjected to repeated freezing and thawing it would gradually move down-slope by solifluction (Section 5.2.1). At the same time, frost heave would tend to move the quartzite blocks towards the surface of the deposit, and frost-sorting would cause them to be grouped together. (These mechanisms are explained in detail by Ballantyne and Harris, 1994). Some accumulations of blocks may have arisen originally as scree, landslips or debris flows but the mechanism of frost-sorting does not require pre-existing concentrations of blocks. On level ground, frost-sorting can give rise to polygonal patterns, but on slopes these pass into stripes. The pattern width generally increases with clast size and the larger forms imply (but not prove) the presence of permafrost (Ballantyne and Harris, 1994, p. 92). However, Clark and Wilson (1997) question if frost heave and lateral sorting could move the large blocks the distances required to create such large-scale forms.

One of the major problems in understanding the formation of stone runs concerns the nature of the matrix which lay between the blocks while the stone runs were being formed and still actively moving down-slope (cf. Clark, 1994, p. 465; Quinn, 1987; Clark and Wilson, 1997). It is mechanically unfeasible that stone runs could have moved without any matrix to separate the blocks, with the general lack of abrasion suggesting some form of lubrication.

While it is assumed that the blocks originally lay in a clay-rich diamicton similar to the solifluction deposits, no remnants of such a matrix are now recognised within the stone runs. It has been suggested that a clay-rich matrix would have been progressively removed by rain-wash and by streams within the stone run (Clark, 1994). This mechanism is here rejected as the *sole* agent for separating the fine-grained matrix from the blocks. Although Clark (1994, p. 469) states 'the simple patterns of stripes show the limited development of integration of drainage paths expectable from combination of distributed water input and relatively smooth hillsides', this mechanism does not explain why the margins of the stone runs are so abrupt, or why the stone stripes are so regular. It is not clear how water flowing between the blocks could transport and remove gravel-sized debris, or even sand: the blocks are more likely to act as a natural baffle and trap such sediment. Washed-out debris does occur at the terminus of some stone runs (Clark, 1994) but alluvial deposits are not apparent at the snouts of most; indeed, many small stone runs have no efferent stream. Moreover, no water-sorted detritus is seen at the base of the stone runs in excavations. The red colouration of blocks in the stone runs indicates a long period of exposure to air, not burial within a fine-grained matrix.

Furthermore, this mechanism does not explain why the surface of stone runs is generally level, when removal of differing concentrations of the matrix would cause differing amounts of settlement. Conversely, the occasional appearance of gentle longitudinal depressions within the larger stone runs could indicate the limited removal of finer material by flowing water. Clark (1994) records up to three metres of settlement following wash-out. The rare occurrence of rounded boulders on the surface of a stone run (Clarke, 1972) indicates some surface flow. Where the surface channel contains no such rounded material, as on the Mount Usborne stone run, flushing out of fines by water flow might have taken place within or at the base of the stone run.

The occurrence of peat as a matrix during formation of the stone run is also rejected. Relatively small volumes of peat are found in areas where stone runs are most common and there is no obvious mechanism whereby a sufficient volume of peat could be introduced between the blocks. No residue of a previous peat fill is seen in stone runs.

Although washing-out of the fine matrix by running water has played a part in stone run formation, it is here thought that gradual replacement of the clay-rich matrix by ice was a more important process. Rock tends to have a higher thermal conductivity and a lower heat capacity than an adjacent moist fine-grained matrix. As the ground freezes, the freezing front will advance more rapidly through the blocks than the matrix, especially if it is at the ground surface, allowing ice to form around it (Ballantyne and Harris, 1994, p.88). This ice will effectively 'push' the fine-grained matrix, together with the smaller stones, away from the block. As diurnal and seasonal freezing and thawing continues, the build-up of ice would progressively increase. At times of thaw some of the fine-grained matrix would probably be removed by water-flushing (cf. Clark, 1972, p. 42). The smaller rock fragments would fall or be washed down between the larger blocks, accounting for the vertical size gradation. This type of frost-sorting, acting on an unsorted bouldery diamicton, would eventually generate an ice-bound concentration of large blocks corresponding to the stone stripes or smaller stone patches.

It is probable that the clay-rich solifluction deposits moved down-slope at a different rate to the stone runs, depending on the gradient and the distribution of frozen material. At times when the solifluction mantle was unfrozen and the stone run was ice-bound, it seems probable that a discontinuity would form at the ice-front and the material ejected from the stone run would tend to be carried away down-slope. This would account for the abrupt marginal contacts with unsorted diamicton. Moreover, an ice-bound stone run would be in hydrostatic equilibrium with the enclosing solifluction mantle. Any tendency for lobes to develop within the stone run would be countered by lateral expansion, accounting for the uniform height of the stone runs and the adjacent ground. Lateral expansion and the faster down-hill movement of the solifluction deposits would both tend to align the blocks in the stone run parallel to the margin and to progressively remove clay-rich material from between adjacent stone patches. Continued down-slope movement of the ice-bound stone runs within the solifluction blanket would allow them to merge and diverge, ultimately amalgamating in stone rivers on the valley floor. The lack of terraces and transverse ridges in the stone runs and in the adjacent portions of the solifluction deposits indicates that they moved faster as they descended lower on the slopes. This is presumably a consequence of more water being present in the deposits, having drained from above, perhaps with more frequent cycles of freezing and thawing at lower altitudes.

The different forms of stone run and the passage from one form to another probably represent variations in the proportion of quartzite blocks in the original weathered debris, and in the absolute and relative rates of movement down-slope by solifluction, and of movement laterally by frost sorting. Other variables such as the angle of slope and the thickness of the debris mantle might also be significant. The formation of some stone terraces might be related to spring-lines. The terraces are often found at breaks of slope. These commonly mark mudstones in the bedrock, such as that marking the base of the Port Stanley Formation (e.g. on the north side of Princes Street) (Section 2.3.4). At times, springs at these mudstones could have contributed either to the ice forming between the blocks in the stone run, or to water flowing out between the blocks and so flushing away the fines, so that a concentration of blocks would form preferentially along the break of slope.

The close spatial and genetic relationship between the solifluction deposits and the stone runs shows that they are co-eval. The stone runs are therefore considered to be between about 26 000 years and 13 600 years old (Section 5.2.1). The volume and extent of the stone runs, and the probability that the upland areas were subjected to periglacial conditions during several periods of the Quaternary, suggest that they could have started forming during earlier cold periods. The degree of differentiation involved in the formation of stone runs suggests that some positive feedback mechanism is involved. Thus, when periglacial conditions resumed after an interglacial period, the processes previously involved in stone run formation would tend to preserve or enhance existing stone runs. Once established, stone runs are likely to persist. However, there is no direct evidence that stone runs formed progressively during several glacial periods of the Quaternary.

5.2.3 Landslips

Evidence for landslips has been noted in only a few places. Other examples may have been obscured by later solifluction or by marine erosion. There are historical records of peat slips.

Several shallow translational landslips have occurred on the southern shore of Estancia Creek [VC 17 77]. One, at least, has an arcuate back-scar and its sole is below sea level at the coastline. The slipped material is a diamicton including both rock fragments and peat clasts (R. Clark, written communication, 1998). These landslips were not observed on aerial photographs. A large landslip occurred during the austral winter of 1994 in the valley intersecting Long Mountain, north-west of Fox Bay. Similar examples could be widespread.

Large landslips and associated debris flows occur on the southern flank of the steep-sided valley on the north side of Mustard Mountain (about 12 kilometres north-west of MPA) [UC 910 674], originating on what is probably a fault scarp crossing the head of the valley. They were observed only on aerial photographs. The landslip scars form a series of arcuate steep-walled hollows, altogether about 1400 metres long. Part of the slipped material is composed of very coarse rock debris, possibly including some blocks 10 metres or more across. The debris flow deposits form a strip only about 100 metres wide immediately below the scars and for some 1200 metres downstream, but on reaching a less steep part of the valley their width increases up to 320 metres wide in a fan-shape over a distance of 200 metres, thereafter diminishing over another 800 metres to a snout less than 100 metres wide.

Photogeological evidence shows that these debris flows truncated a series of stone-covered solifluction terraces and adjacent valley-side stone runs, clearly post-dating them. On the other hand, the debris fan has been strongly dissected by the valley bottom stream, suggesting that it formed at a time when the ground was still poorly vegetated. The landslips and debris flows therefore probably formed about the end of the last glacial period, roughly 14 000 years ago.

Peat slips can be regarded as a particular kind of landslip. Peat deposits on elevated ground, including ridge crests, are fairly common in parts of the Falklands. Water is prevented from draining freely by the underlying clay-rich solifluction deposits, and when water-logged, the thicker peat deposits can loose internal cohesion, resulting in a bog-burst (Clark, 1991). At some springs, a layer of peat and vegetation covers a pool of water, forming a 'water blister'. Although these features are perennial, some eventually collapse, forming a landslip (e.g. on the east side of Long Mountain, north-west of Fox Bay. Peat slips seem to present a more serious potential hazard in the Falklands than landslips in other materials. A peat flow occurred in lowland peat in Ball Mountain Camp [UC 9 8], c. 1980 (R. Pitaluga, written communication to R. Clark, 1992). Examples of peat slips are described by Wilson et al. (1993). Weller (1975) attributes the formation of some ponds to slippage in peat.

Two large peat slips are known to have occurred in Stanley. The first occurred on 30 November 1878, close to the position of what is now Philomel Street. It is fortunate that there were no fatalies (Bailey, 1879). The second, on 2 June 1886, was some 180 metres to the west of the first, close to the site of St Mary's Cathedral. It caused two deaths (Strange, 1983). Both slips were presumably the consequence of water having been allowed to accumulate within peat diggings at the top of the slope on which the town stands, leading to the saturation of the peat banks and their eventual collapse. Subsequent peat digging has reduced these particular banks to mere remnants. However, where there is excavation within or close to peat banks, the potential consequences of impeded drainage and saturation of the base of the peat should borne in mind.

5.2.4 Ramparted hollows (pingo scars)

Closed ramparted hollows occur on fairly low ground between Swan Inlet and Brenton Loch, between about 30 and 60 metres ASL, with some possible examples in the Doyle Valley area. They appear to have formed under periglacial conditions and are probably relict pingos. Pingos are considered to be diagnostic of the existence of permafrost.

The best examples of closed hollows occur low on north-facing slopes in the ridge-and-valley terrain typical of the Brenton Loch Formation (Section 2.4.4). They range from about 20 metres to 120 metres in diameter and are mostly enclosed by a low rampart up to 20 metres wide. The rampart is generally up to about two metres higher than the surrounding ground, although disappearing on the up-slope side of the hollow. The ramparts slope at up to 20° on the inside and about 10° on the outside. The hollows occur singly or in short chains at about the same height along the slope. In some cases the ramparts touch and merge, in others they do not. The interior of the hollow can be a blind irregular depression as much as one or two metres below the level of the ground outside the rampart, or it can be flat at a similar level to the surrounding ground. A few hollows hold small pools. None have afferent or efferent channels, although water could drain into the hollows from the adjacent hillside. The ramparts appear to be constructional landforms, not erosional ones.

Weathered bedrock probably lies close to the surface on slopes adjacent to the ramparted hollows. It appears that the ramparts are mostly formed of unconsolidated thin stony sandy clay similar to the local regolith, but in at least one locality the rampart includes well-sorted sand. The flat-bottomed hollows within the ramparts are infilled by thin peat and organic clay (up to 0.7 metres) underlain by smooth blue-grey clay (in one case more than two metres thick), interpreted as lacustrine deposits. The other hollows appear to have been free-draining. Some thin peat occurs in the adjacent valley bottoms.

A line of six hollows occurs beside the track to Cantera House, about one kilometre west of Camilla Creek [UC 6500 6335]. A line of larger and deeper hollows occurs south of the Darwin Road [UC 7073 5940], continuing along this valley for at least 800 metres to the south-east. A single isolated hollow occurs a few kilometres west of Swan Inlet

[UC 8992 5192]. The closed hollows are too small to be identified on aerial photographs with confidence unless they have first been located on the ground, and so could be more numerous than so far recognised.

The form of these ramparted hollows suggests that they originated as ice-cored mounds, with the ramparts representing material that slipped from the middle to the edge of the mound. They are of comparable size to pingo scars. Pingo scars can occur low on valley sides, and on sloping ground the rampart is typically absent in the upslope direction (Flemal, 1976). Pingos form where water progressively seeps into a frozen zone close to the ground surface, forming a lens of ice. This happens only under permafrost conditions, and it seems that springs would have occurred at the sites of the ramparted hollows described here only when the ground under the adjacent valleys was frozen. Clark (1972) suggested that some lakes in the Doyle River area originated as pingos or other thermokarst depressions, but no soil fabrics diagnostic of permafrost (such as involutions) have yet been reported from the Falklands.

Another form of ice-cored mound, the mineral palsa, is broadly similar to the pingo but forms only in fine-grained materials (Ballantyne and Harris, 1994). Mineral palsas are generally smaller than pingos but lie within the same size range (Flemal, 1976). Some of the problems of distinguishing relict pingos from relict mineral palsas are discussed by Gurney (1995). The position of the ramparted hollows in the Falklands on valley sides, and apparently not in a particularly fine-grained substrate, suggests that they are relict pingos, not palsas.

5.2.5 Glacial deposits

Glacial deposits and glacially eroded hollows (corries) were recorded by Clapperton (1971) and described further by Clark (1972) and Roberts (1984). They occur only on hills rising above 500 metres, on Mt Usborne, in the Hornby Mountains, near Mount Maria, and in the Mount Adam range. They are the product of small glaciers no more than 2.7 kilometres long (Clapperton and Roberts, 1986).

The corries are characterised by a semi-circular outline, a vertical back wall from 15 to about 200 metres high, and a low hummocky embankment of moraine at their mouth. Many contain small lakes (tarns), such as Black Tarn on Mount Usborne (McAdam and Roberts, 1981). There were at least two periods of corrie development (Clapperton, 1971; Roberts, 1984). The older corries are partly covered by solifluction deposits in addition to moraine. The younger landforms were little affected by periglacial activity.

Glacial till (composed of boulders and stones in a compacted sandy pebbly matrix) forms a deposit up to 25 metres thick in Double Stream Basin, south-west of Port Howard, extending down to 165 metres above sea level (Clapperton, 1971; Clapperton and Sugden, 1976). Roberts (1984) describes till in four other valleys, all in West Falkland. Roberts (1984) suggested that the valley tills were deposited during an older glaciation prior to about 26 000 years BP (before present), and that the younger corries were formed during the last Wisconsin glaciation by smaller glaciers confined within the corrie threshold, prior to 9 300 years BP. Subsequent work on the age of basal peat layers further constrains the age of the last glaciation to before 13 600 years BP (Clark and Wilson, 1992) (Section 5.2.7).

No evidence for more widespread glaciation on the Islands has been found. It seems unlikely that any glacial deposits or landforms that might have existed on lower ground have been completely covered or obliterated and so it is thought that there were none. The lack of more extensive glacier development can be attributed to a lack of snowfall in the lee of Patagonia. Also, it has been suggested that tectonic uplift had raised the upland areas to the threshold of glaciation only by the onset of the Wisconsin glaciation (Roberts, 1984; Clapperton and Roberts, 1986) (Section 5.2.10).

In addition to the landforms formed by glaciers, Roberts (1984) identified 56 nivation features: hollows, crests and corries formed by mass wasting at the margins of snow-patches. Most are in the same general area as the glacial landforms, with just one on Mount Philomel. These also appear to have formed during the Wisconsin glaciation (Clapperton and Roberts, 1986).

Clapperton (1971) and Roberts (1984) noted that most of the nivoglacial landforms are on north-east or south-east facing slopes. This was attributed to snow being swept off the south and west-facing slopes by the prevailing winds and accumulating in the lee of the hills. It contrasts with the more usual situation, where the orientation of corries is controlled by aspect relative to solar radiation.

5.2.6 Aeolian deposits

Wind-blown deposits are mainly situated down-wind of lakes or of west to south-west facing sand beaches. They are thus most extensive on West Falkland, particularly in a belt encompassing Lake Sulivan, the Doyle Valley and the Blue Mountain area. The aeolian deposits are mainly composed of fine to medium-grained sand, but sections through blown sand deposits commonly also reveal numerous thin interbeds or laminae of wind-transported peat particles. Examples may be seen in cuttings along the tracks between Stanley Airport and Cape Pembroke, and in other thick aeolian sand deposits.

The thickest aeolian deposits include relict sand dunes, most of which are vegetated and partly degraded. Extensive dunes occur behind many of the large west-facing sand beaches, for example Elephant Beach (Pebble Island), Elephant Beach (Foul Bay), Yorke Bay and Paloma Beach. Relict dunes up to 20 metres high occur in Centre Camp, downwind of Lake Sulivan. In the 'lakeland' terrain around Mount Sulivan and extending east to the Blue Mountain area, many of the lakes have aeolian deposits on their eastern margin. Indeed, many of the lakes may owe their origin to the creation of deflation hollows by wind erosion (Wilson, 1994). In this area with many closely-spaced lakes and drainage channels, a complex interleaving of aeolian, lacustrine and alluvial deposits can be expected (e.g. Clark et al., 1994, fig. 1).

In addition to hummocky sand deposits, there are also thin sand sheets which drape large areas without establishing a distinct landform. These occur as 'sand blows' from west-facing beaches and from some of the larger lakes. In an extreme case, the prevailing wind has carried sand eastwards from the Blue Mountain area across the Coast Ridge. Indeed, some of the sand around Blue Mountain may itself have been blown from East Bay, through the Doyle Valley.

There are some isolated small sand deposits not visible on aerial photographs (e.g. beside the track north-east of North Arm [UC 393 240]) which could represent relict dunes which have migrated far from their source.

Aeolian sand overlying and interbedded with in situ peat occurs near Lake Sulivan House and around Blue Mountain, amongst other places. So far it has not proved possible to correlate the alternating periods of stabilising peat growth and of blowing sand either between sites, or with regional climatic variation (Wilson, 1994; Wilson, 1998). The age of interbedded peats indicates that some sand accumulated in the early Holocene or late Pleistocene, but that most was deposited in the mid to late Holocene (Wilson, 1994).

Ventifacts (clasts which have been abraded and polished by wind-blown particles) of quartzite, mudstone, dolerite and diamictite occur at the base of the peat in a variety of situations in West Falkland (Clark and Wilson, 1992). Polished surfaces of Stanley Quartzite occur locally on some rock outcrops in the Cape Pembroke peninsula. Clark and Wilson (1992) concluded that most ventifacts are associated with a major period of wind activity following a episode of severely cold climate but before the restoration of vegetation cover and peat growth, that is prior to about 13 000 years ago.

Many aeolian sand deposits are indicated on topographic maps of the Islands by a fine stipple ornament designated as 'sand or clay'. Note that in some places aeolian deposits are more extensive than thus indicated, but also that some areas shown by this ornament are actually eroded ground without any aeolian deposits. These eroded areas are vegetation-free interfluve areas, not subject to tidal inundation, where exposed bedrock, regolith or some superficial deposit other than blown sand is subject to active erosion. Most are in coastal areas which were apparently once covered in tussac grass, but some occur inland where other vegetation has been destroyed. The eroded areas commonly include patches of wind-blown sand. Soil erosion is discussed in detail by Wilson et al. (1993).

Some fine-grained wind-blown debris is carried to the Islands from distant sources. The most dramatic component of this kind of far-travelled deposit is volcanic ash. In August 1991, fine-grained tephra from an eruption on Mount Hudson (in the Chilean Andes at latitude 46° S, some 1000 kilometres distant, Naranjo et al., 1993) was carried to the Islands by north-westerly air flows. The ash completely covered the ground in places and created considerable difficulties for some time afterwards. Preliminary searches for traces of volcanic ash in Falklands peat deposits for the purpose of tephrochronological studies have shown that similar ash-falls have occurred on numerous occasions in the past 10 000 years or so (Clark and Wilson, 1997, P. Wilson, written communication, 1998; Holmes et al., 1999), and they presumably also occurred prior to that. A volcanic ash layer up to 15 centimetres thick found in Tierra del Fuego has been traced to an eruption of Mount Hudson (900 kilometres distant) some 6 800 years ago (Stern, 1991).

5.2.7 Peat

Peat accumulations occur from below high tide level to more than 600 metres above sea level. Three broad categories of post-glacial peat deposit are distinguished here: upland peat, lowland peat and tussac peat. More detailed descriptions are given by Wilson et al. (1993) and Clark and Wilson (1997). There are a very few known examples of pre-glacial Quaternary organic deposits. Wind-blown peat debris is noted in Section 5.2.6.

'Upland peat' usually overlies the resistant lithologies of the Port Stephens Formation and Port Stanley Formation. It generally forms upstanding blankets and peat hags on interfluves, including some on quite narrow ridges. The eroded margins of these deposits can stand more than two metres high. The hags are commonly associated with small shallow ponds and areas of bare rock and regolith (Plate 39). Upland peat overlies solifluction deposits, and in some places also raised beach deposits (Plate 37). It is particularly widespread in northern East Falkland. Further details are given by Clark (1991) and Clark et al. (1994).

Peat blankets which lie close to the present-day drainage channels are here categorised as 'lowland peat'. They can occur on bedrock of any type. The most widespread occurrences are relatively small, narrow or thin deposits in the floor of river valleys where they commonly overlie alluvium and, perhaps, solifluction deposits. The most extensive deposits occur where drainage channels are widest. This can be in the headwater areas which have experienced little dissection, but can also occur near the coast if valleys are broad. Lowland peat also occurs in dried-out lake beds of all sizes. It locally occurs at sea-level, for example east of Hookers Point, near Stanley, and near Black Point, Port Louis.

Tussac grass (*Poa flabellata*) is generally underlain by peat which can exceed 10 metres in thickness. 'Tussac peat' tends to contain hard lignitic material, although this readily crumbles to pieces the size of coarse sand or granules. This was noted on Beauchêne Island by Lewis Smith and Clymo (1984), who observed that it was apparently anomalous for lignite to form in peat deposits which are so young and thin. However, observations during the present survey indicate that this is a usual characteristic of tussac peat. Tussac peat occurs only in coastal locations. It is very susceptible to erosion so tends to be preserved only where there is at least a partial cover of living vegetation.

Bog wood is absent from the peat deposits, indicating that prior to settlement no trees have grown in the Falklands in the post-glacial period (cf Birnie and Roberts, 1986). However, a notable concentration of sub-fossil bird bones occurs within the peat on West Point Island, some nine feet [2.7 metres] below the original surface. The bones are distributed in a layer at least 30 centimetres thick and are partly enclosed in a clayey matrix. The bones are disarticulated but otherwise undamaged. A diverse assemblage is represented, including penguin, caracara, and finch, but no evidence has yet been found for species not currently present in the Islands. This deposit has been suggested as the site of a pond (Hattersley-Smith and Hamilton, 1950; Adams and Woods, 1997, R W Woods, written communication, 1998), although the evidence for this seems largely circumstantial.

Field relations show that most peat is younger than the solifluction deposits and so post-dates the climatic amelioration at the end of the Wisconsin glaciation. Although some of the earlier radiocarbon age determinations from the Falklands suggested that post-glacial peat growth started about 10 000 years ago (Barrow, 1978; Roberts, 1984), Clark and Wilson (1992) report ages of basal peat layers of 13 610 ± 45 and 11 080 ± 45 years BP (before present) from near Lake Sulivan, and (quoting J R Pilcher, personal communication) of 13 077 ± 69 years BP from Eliza Cove, Stanley. Lewis Smith and Prince (1985) obtained a radiocarbon age of 11 600 years BP from the basal layers of tussac peat on Beauchêne Island.

Organic clay from about 10 centimetres below the floor of Black Tarn gave a radiocarbon age of 6460 ± 110 years BP (McAdam and Roberts, 1981). Peat layers interbedded with a sequence of aeolian sands up to ten metres in thickness near Blue Mountain gave radiocarbon ages between 8590 ± 40 and 815 ± 45 years BP (Wilson, 1994). Peats in similar sequences up to three metres in thickness near Sulivan House gave radiocarbon ages ranging from

Plate 37: Raised beach deposits and peat, Grand Jason

 MN28266

Note small rounded boulders in background in addition to pebbles and gravel in foreground

Plate 38: 'Lakeland terrain': Lake Sulivan

 MN28264

7650 ± 45 to 600 ± 45 years BP (Wilson, 1998). Bird bones from peat on West Point Island have yielded a radiocarbon age of 2 200 ± 100 years (Weller, 1975).

There are only two known examples of thin pockets and layers of peat which pre-date the solifluction deposits. The first to be documented is exposed in a low sea-cliff at Kelly's Garden, San Carlos [UC 596 834]. Roberts (1984) records a three centimetre thick layer of black amorphous peat, interpreted as a relict soil, overlain by 1.5 metres of poorly sorted silty sand, regarded as a solifluction deposit. That is in turn overlain by some 30 centimetres of fibrous peat beneath the present-day soil. The relict soil is underlain by some 40 centimetres of unconsolidated pebbly silty sand, also of probable solifluction origin, which in turn overlies weathered bedrock sandstone. The thin peat layer yielded a radiocarbon age of 26 060 ± 400 years BP (DD Harkness, personal communication, 1981 in Roberts, 1984). The pollen assemblage within the peat is dominated by grasses (*Graminae*), and is broadly similar to that of post-glacial peats (Birnie and Roberts, 1986).

Two organic-rich silt layers are exposed in low cliffs on the east side of Plaza Creek [UD 965 057]. They occur within a sequence of sands and silty clays interpreted as valley-side periglacial deposits. Their radiocarbon ages fall in the range 35 970 ± 280 to 28 150 ± 350 years BP, in the late Pleistocene. As with the Kelly's Garden deposit, the contained pollen assemblage is dominated by grasses. Sedge pollen also forms a significant contribution and there is a variety of other pollen in very low abundance, including some tree pollen (Clark et al., 1994; Clark et al., 1998).

5.2.8 Alluvium and river terrace deposits

At times in the past, erosion of river valleys in the Falklands has taken place when the sea level has been much lower than at present (Section 5.2.10). The various bays and inlets in the convoluted coastlines of the Islands are river valleys drowned by rising sea level. As sea level rose, the rivers had to readjust to a higher base level and consequently are now sluggish and underfit (Plate 38), although many are also slightly incised, presumably due to late rejuvenation. Although many valleys are quite narrow, especially in their lower reaches, they have flat bottoms across which the streams meander. Unconsolidated deposits underlying these flat valley bottoms are thought to be mainly alluvium (river deposits) with peat. In the upper part of the rivers, the valleys tend to broaden out into enclosed flat-bottomed basins with very peaty floors.

River bank exposures of gravels and sands can be seen in places, but the alluvium probably includes silts and muds in addition. It is usually closely associated with lowland peat, which commonly forms a surface blanket on valley bottoms. Indeed, in the smaller drainage channels, it is usual for the stream to be confined within peat. The channel sides are then near-vertical as a consequence of the upwards increase of the peat deposit.

In places, the alluvium might also be interbedded with or overlie mass-movement deposits, but these are most likely to have formed during periods of cold climate, when erosion of the river valleys would have been more active. Given that the river valleys have been flooded by rising sea level (Section 5.1.10), alluvial deposits can be expected to pass offshore, and to be overlain by or interfinger with estuarine, lagoonal or beach deposits. Some tidal inlets have gravel bottoms (and so support the passage of vehicles), which are presumably of alluvial origin.

River terraces do occur, for example on the Arroyo Pedro (Greenway, 1972, fig. 2), but are not very extensive. This can be attributed to two main factors. Firstly, when sea level was higher than at present, the Islands would have been much smaller and so the rivers would have been correspondingly shorter. Secondly, as sea level fell, for example during the Wisconsin glaciation, river terraces which then formed would be prone to erosion as the river valleys adjusted to the lower base level, and to reworking by periglacial mass-movement. It is possible, however, that some river terrace deposits are preserved beneath the mantle of solifluction deposits.

A small area of semi-consolidated superficial conglomerate occurs in the bottom of the valley at the eastern end of Limpet Creek [UD 9082 0878], overlying bedrock on a subhorizontal contact about four metres above the high tide mark. This deposit is one to two metres thick, and comprises open-textured, clast-supported breccio-conglomerate with very little matrix, passing up into sandy conglomerate. The constituent pebbles and cobbles are angular to subrounded, and from gravel-sized up to about 25 centimetres across. They are all composed of quartzite similar to the local bedrock. This conglomerate appears to be a river terrace deposit.

An isolated outcrop of weakly bedded breccio-conglomerate is exposed in the cliff at the head of Port Sussex [UC 6365 7357]. This is similar to the occurrence at Limpet Creek, but consists largely of subrounded fragments of tillite and subangular clasts of mudstone, both exposed nearby. Rare cobbles of quartzite are well-rounded and could have been derived from the tillite. There is one thin layer of laminated mudstone. Bedding in this exposure dips north-east at 22°, away from the adjacent shoreline, due either to local tectonic displacement, or to subsidence following erosion.

5.2.9 Lacustrine deposits

Small lakes and ponds are a familiar feature of the Falklands landscape. They occur in a range of substrates, including peat, sand, solifluction debris and bedrock, and in a variety of situations ranging from coastal to hilltop. While some of those in low-lying areas underlain by alluvial, aeolian or lacustrine deposits may be in hydraulic continuity with the local groundwater, many Falklands ponds are 'perched' above an impermeable layer of weathered bedrock or solifluction debris. Indeed, many ponds stand at an appreciably different level to their immediate neighbours (perhaps only a few metres distant) and so must be isolated within a substrate of very low permeability (Plate 39).

The formation and biodiversity of ponds in the Falkland Islands is discussed by Clark (1972) and Weller (1975). Many of the flooded depressions in either bedrock or superficial deposits probably formed as deflation hollows during dryer periods (Clark et al., 1994; Wilson, 1994) (Plate 38). Others might reflect marine or fluvial erosion at times of higher sea level. A few occupy glacially-eroded hollows (Section 5.2.5). Some, especially the many near-circular ponds in the 'lakeland' terrain of West Falkland, might have originated through the formation and melting of ground ice. Although the lowlands were not glaciated (Section 5.2.5), there is some evidence for the past occurrence of permafrost, in the form of pingo scars (Section 5.2.4) (Clark, 1972, p. 46). Many smaller ponds seem to have originated by cracking or slippage in peat. Others occupy portions of alluvial plains, perhaps in depressions formed by differential compaction of the alluvium. Small ox-bow lakes are common near meandering streams. Coastal lagoons formed behind barrier beaches are discussed in Section 5.2.10. In many cases, the ponds could have been initiated as quite small shallow depressions. Once established, through whatever cause, they will tend to have increased in size by wave action and (during dry periods) by further deflation. Unless constrained by a resistant substrate, many ponds will thereby develop and maintain a circular outline. The elongate outline commonly seen (Plate 38) is generally due to the ponds having been confined between bedrock ridges.

Some lakes eventually become infilled by sediment, including peat, while others have diminished through drainage capture. Dried-out lake beds are a common feature of the 'ridge-and-valley' terrain between Grantham Sound and Choiseul Sound but occur widely elsewhere. The Purvis Pond Airstrip at Port Howard stands on an old lake bed.

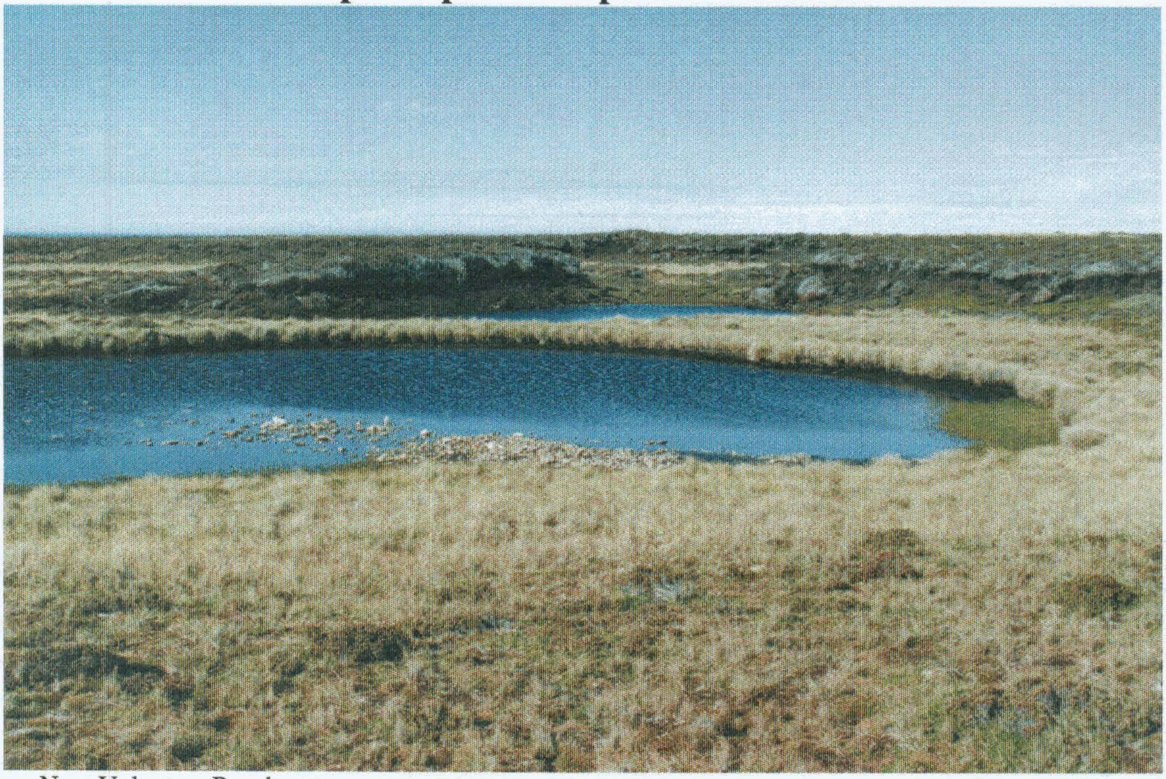
Lacustrine deposits include peat, presumably with clays, silts and sands, including some wind-blown material. Their thickness is unknown, but might be sufficient to present obstacles to construction (and construction traffic). McAdam and Roberts (1981) describe some 30 centimetres of clays (some organic) and silts from the bottom of Black Tarn.

5.2.10 Coastal deposits and marine erosion levels

Coastal deposits related to the present sea level are widely distributed but mostly of limited extent. They are chiefly beach deposits, with some estuarine and lagoonal deposits. These are often closely associated with wind-blown sand and with alluvium and peat. Evidence for past marine erosion levels is more extensive, both above and below the present sea level, although raised beach deposits are relatively rarely seen.

Most coastal deposits appear to be composed of sand, gravel or cobbles, but some silts and muds can be expected, especially in the lagoonal deposits. Some beach sands include a proportion of calcareous debris, chiefly fragments of mollusc and brachiopod shells, with some foraminifera. Relatively extensive shell limestone deposits near Shell Point, Fitzroy [VC 15 60] are described by Adie (1953) and also noted by Clapperton and Roberts (1986) who ascribe them to their 6-8 metre raised beach level. A few modern beaches are composed mainly of gravel and sand-sized debris of the calcareous bryozoan *Lithothamion* (Section 7.7.1).

Plate 39: Upland peat with ponds at different levels



Near Volunteer Beach. Photograph by Don Aldiss, BGS. MN28265

Plate 40: Barrier beach, coastal lagoon and lakes

Bertha's Beach. Photograph by Don Aldiss, BGS. MN28256

Barrier beaches are one of the most characteristic landforms on the Falklands coastline (Plate 40). These form through wave action, particularly at river mouths. Most rivers are too sluggish to redistribute the material thrown up by the sea, so bars are more common than spits. Brazo la Mar, south of Walker Creek, provides a rare example where the stream flow is a stronger force than the local wave action. The aerial photographs show that a spit extends from the south-west bank across the estuary mouth, but does not block the river to form a complete bar and lagoon. The spit shows successive growth positions, indicating that this situation has prevailed over a long period of time.

Where bars have formed, they typically create a freshwater pond at the mouth of the river which has been blocked. These ponds are generally a metre or more higher than mean sea level, although some may be inundated by the highest tides. Some of these barred ponds have surface outflows, and some not, although this may depend on the time of year and local rainfall. While the surface deposits in many beach bars are well-sorted, wave-washed gravels and clast-supported cobble-conglomerates, the main part of the deposit must include an impermeable matrix.

In many places, barrier beaches have also formed between one island and another, creating a tombolo. A superb example more than 500 metres long occurs at the north-east end of Barren Island. The peninsula of Cape Pembroke is joined to the rest of East Falkland by a raised tombolo, on the landward side of Surf Bay. East Cove, south of Mount Pleasant Airport, is closed to the east by a complex tombolo (joining two islands to East Falkland in series) which is the southern extension of Bertha's Beach. Some of the many excellent examples on Lafonia and the surrounding islands are double tombolos. For example, the Twin Ponds on the west side of Speedwell Island are bounded by two tombolos, each linking a rocky islet to the main part of the island. Flores Harbour Island [UC 23 10] actually consists of two rocky islets joined by a tombolo at either end of the central shallow pond.

Although the tidal range is not large, the power of tidal streams is locally very great. Although some inlets have narrow entrances, tidal flow is apparently sufficient to prevent a significant accumulation of material by wave action. The balance between wave action and tidal currents can explain landforms at Volunteer Lagoon. The north-east arm of the lagoon is barred by the tombolo forming Volunteer Beach. The eastern arm is partly blocked by a spit on the northern shore, Lagoon Bar. Aerial photographs show that for several kilometres west of this spit the lagoon is very shallow with a sandy bottom, but that the central and western part of the lagoon is deeper. It seems that a build-up of sediment through wave action has almost blocked the eastern mouth of the lagoon, but that tidal currents sweeping over the barrier prevent it from being built above sea level.

At Elephant Beach (Foul Bay) the balance seems to have tilted towards wave action. Here a beach some 3.5 kilometres long is backed by a complex of superficial deposits extending about five kilometres east to Elephant Beach Pond, and so mostly infilling a large tidal inlet. Remarkably, significant sedimentation has occurred in this area within the last 150 years. It is reported by those who know the area that the first house on Elephant Beach Pond was brought to the site by boat from the sea, during the 19th century. This would no longer be possible, even though the flood of the very highest tides still reaches Elephant Beach Pond. The presence of remnants of a small jetty near the head of Smylie's Creek, now barely navigable even at high water, suggests fairly rapid sedimentation in this estuary also.

As first noted by Andersson (1907), it is clear that the deeply indented coastline of the Falklands represents a drowned topography of river valleys incised when sea level was relatively much lower than it is now. Low sea level tends to occur during glaciations, and rates of erosion may then be increased due to more intense frost action and less extensive vegetation cover. Marine erosion levels which may exist below present sea level are difficult to investigate, for obvious reasons. Roberts (1984) notes that incised channels in the sea bed, which apparently mark the position of rivers at times of lower sea level, can be traced on bathymetric maps to a depth of about 50 metres. More locally, it is reported that divers have found that the cliffs on the west side of New Island are undercut below sea level. This undercut, apparently formed by wave action at a time of lower sea level, is associated with accumulations of boulders, presumed to be beach deposits (I J Strange, oral communication, 1997).

Evidence for raised beach deposits and marine erosion levels is widespread. It was first noted by Andersson (1907), although most of his examples are questionable. In particular, rounded cobbles of pink quartzite found at about 60 metres above sea level near Cape Meredith have been eroded from basal conglomerates in the Port Stephens Formation (Section 2.3.1.3) and do not mark a raised beach, contrary to the interpretations of Andersson (1907). Roberts (1984) and Clapperton and Roberts (1986) reviewed the topic in detail. They describe evidence for raised beaches at five different levels. They found deposits at two to four metres ASL (above sea level) and six to eight metres ASL which are younger than about 4 000 years BP. Others at one to four metres ASL, 20 to 25 metres ASL and 60+ metres ASL all predate the solifluction deposits and are older than about 26 000 years BP.

However, much caution is needed in the interpretation of raised marine erosion levels and associated beach deposits. While there are extensive marine erosion levels, many of the terrace-like features seen in the Islands reflect structures in the bedrock (Adie, 1953) or in periglacial deposits (Sections 5.2.1 and 5.2.2). Recognition of raised marine deposits can be problematical. Spreads of pebbles and small cobbles are fairly common near the coast. While some of this material probably does represent raised beach deposits, some has been carried onshore by seabirds or seals (Section 3.2), or possibly was thrown up during storms. Where raised beach deposits do exist, they may have been redistributed down-slope by solifluction or land-slip. This is very likely in the case of the older, higher levels.

Nevertheless, evidence for raised beach deposits at similar levels to those described by Clapperton and Roberts (1986) was found in many places during the present survey, for example at about three metres ASL at Shallow Bluff, Shallow Harbour, beneath up to two metres of peat near the southern end of Port Edgar, between five and six metres ASL on Grand Jason and about 10 metres ASL on Carcass Island.

There is also some evidence for raised beach deposits at other levels. Rosenbaum (1984; 1985) notes that several metres of green-grey marine clay mantle the Stanley Quartzite in the Cape Pembroke peninsula, where it rises to 35 metres ASL, and elsewhere in the Stanley area. This may be of similar age to massive slightly sandy clayey silts and clayey very fine sands exposed in two roadside gullies on the north side of the MPA road at Bluff Cove [VC 2002 6768]. It underlies about two metres of stony solifluction deposits on a narrow terrace at around 35 to 40 metres ASL. This clay is grey, weathering to yellowish brown. A smooth blue-grey clay, exposed in a trial pit, occurs beneath the surface peat deposits close to Cape House, near Cape Dolphin at about 45 metres ASL.

Small pockets up to one metre thick of loose clast-supported conglomerate overlie the Stanley Quartzite in Ponys Pass Quarry [VC 340 697]. Very well-rounded cobbles and boulders of quartzite in a gravely sandy clay matrix underlie a thin solifluction diamicton and peaty soil. Some of the boulders have small crescentic fractures indicating a period of high-energy abrasion, as on a beach. Parts of the irregular underlying quartzite surface have been smoothed by subaqueous abrasion. The base of this deposit in the south-east corner of the quarry lies at 41.8 ± 2 metres ASL (R. Challoner, oral communication, 1997).

Evidence for raised marine erosion levels is found in large-scale landforms in the north of the Islands. Andersson (1907) noted what appears to be a horizontal terrace crossing the southern flank of Mount Low, north of Stanley. This is part of a widespread sloping erosion surface between about 80 metres and 30 metres ASL which is found in many parts of northern East Falkland. Isolated hills which stand higher than this surface tend to be bounded by a distinct negative break of slope which is unrelated to bedrock structure. This surface is extensively dissected by river valleys and it is largely covered by solifluction deposits and peat. The surface is interpreted as a marine erosion level, or series of such levels, pre-dating the last glaciation. Few marine deposits have yet been found on this surface, although they could be widespread beneath the upland peat and solifluction deposits. The uniform height of many ridges and summits throughout the Falklands suggests that remnants of still higher erosion levels could be present (Clark, 1976, p.172).

Similar sloping geomorphological surfaces are clearly seen on the larger islands to the north of West Falkland. Pebble Island and the larger of the Jason Islands each have upstanding relatively steep-sided rocky hills separated by a persistent subhorizontal negative break of slope from areas of gently undulating or sloping low-lying ground. Rocky outcrops above this break of slope tend to be rougher and more craggy than those below it, reflecting differences in erosional history. As in the north of East Falkland, the upstanding hills are unrelated to bedrock structure, and so the surface enveloping the lower ground is interpreted as a marine erosion level.

On the larger Jason Islands the break of slope at the foot of the upstanding hills is between about 40 and 45 m ASL. The marine erosion level slopes down to 10 or 12 metres ASL near the coast. Any intermediate breaks of slope have been obscured by solifluction deposits and peat. (The smaller islands in the Jason Group all lie below about 15 metres ASL). The main marine erosion level is at a similar height range on Carcass Island, but another erosional terrace at 70 metres is seen on the south side of Jason Hill, at the south end of the island. A strong break of slope also appears on Pebble Island. This is morphologically very similar to that seen on the Jasons, but lies at about 55 metres ASL, 10 to 15 metres higher. Discontinuous sections of erosional terrace also occur at levels up to 110 metres ASL on Pebble Island, but none can be traced very far. A negative feature at 24 metres ASL west of Marble Mountain possibly correlates with the 10 to 12 metre level on the Jasons. To the east of Pebble Island, the ground between the Tamar Pass, Foresight Hill and Goat Hill appears to be an erosion surface. Negative topographic features occur at 110 metres on the north side of Goat Hill, between 85 and 95 metres on the north side of Foresight Hill and from 55 metres to 70 metres between Mare Rock and the north side of Goat Hill. Some elements of these features may be structurally controlled. Remnants of a raised beach at 30 to 40 metres ASL can be seen on New Island, and at 20 metres and 40 metres on Beauchêne Island. Raised sea stacks occur at the southern end of Beauchêne Island.

Some traces of possible marine erosion levels can also be seen on Saunders Island but their extent is masked by bedrock features and by superficial deposits. These factors may explain why marine erosion levels are not clearly seen in other parts of the islands. Nevertheless, there is considerable scope for a detailed survey and analysis of raised beach deposits and erosion levels throughout the Islands, although it would be limited by the relatively coarse resolution of the available topographic maps.

Clapperton and Roberts (1986) consider that the various raised beaches occur at approximately the same level in different areas, and that there is no evidence for differential uplift across the Islands. However, this seems inherently unlikely, given that a very small regional tilt would be required to create a significant difference in elevation across the width of the Islands. This is supported by the apparent difference of 10 to 15 metres in the height of the principal erosion surface on Grand Jason and on Pebble Island, for example. As there is evidence for several successive sea levels, and as the evidence for any one of them is at best fragmentary, correlating an individual level across the Islands is fraught with difficulty.

Clapperton and Roberts (1986) suggest that the two highest levels of raised beach deposit which they identified formed during interglacial periods. However, they note that as no large glaciers formed on the Islands during the Quaternary, the subsequent uplift must be largely attributed to regional tectonic uplift rather than glacio-isostatic recovery. The cause of the uplift remains unknown but is possibly related to compression on the northern margin of the Scotia Plate (Figure 6.1). Formation of a high level marine erosion level during the Sangamonan interglacial (Oxygen Isotope Stage 5, immediately prior to the Wisconsin glaciation), followed by regional uplift prior to about 35 000 years BP, could help explain the lack of Pleistocene deposits in the Falkland Islands.

6. REGIONAL SETTING OF THE FALKLAND ISLANDS

This section considers the regional setting of the Falkland Islands, both in the present and when they were part of the Gondwana supercontinent in the Early Jurassic.

The Islands lie on the southern end of the South American tectonic plate, separated from the main part of the plate by the Falklands Fracture Zone and from the much smaller Scotia Plate by the Falkland Thrust (Figure 6.1) (Richards et al., 1996a). North of the Falklands Fracture Zone, the eastern edge of the South American continent lies at about the same longitude as East Falkland, as expressed by the bathymetric contours (Figure 1.1). South of the Falklands Fracture Zone, the Falklands Plateau and the Maurice Ewing Bank are formed by a complex of continental crustal blocks, together with some oceanic crust (Figure 6.1) (Lorenzo and Mutter, 1988; Barker, in press). The Falkland Islands lie on a part of the Falklands Plateau known as the Falklands Microplate (Mitchell et al., 1986; Richards et al., 1996a). The Scotia Plate and most adjacent parts of the Antarctic Plate are formed by oceanic crust (Cunningham et al., 1998).

The geology of the Falklands offshore area has been described by Platt and Phillip (1995), Richards et al. (1996b) and Richards and Fannin (1997). As shown on the regional map inset on the East Sheet of the geological map (Aldiss and Edwards, 1998), the Islands are surrounded by an interconnected series of major sedimentary basins. These are: to the north, the North Falkland Basin; to the west, the Malvinas Basin; to the south, the South Falkland Basin; and to the east, the Falkland Plateau Basin. These basins appear to have formed initially as extensional rifts associated with the break-up of Gondwana. They are infilled by sequences of Jurassic to early Cenozoic sedimentary rocks, underlain by Middle to Late Jurassic volcanic rocks in some areas. These basinal sequences overlie the offshore continuations of the Lafonia Group and the West Falkland Group.

Onshore structures indicate that the Falklands Microplate is itself made up several crustal blocks, which have at times moved several kilometres laterally or vertically relative to each other (Section 4). Nevertheless, the consistent orientation of the D1 structures shows there has been little differential rotation between these component blocks. Apart from the reverse fault movements during the formation of the D2 drape folds (Section 4.2.2), the only evidence for convergence between the blocks within the microplate is found in the north of the Islands, in the zone of D4 thrusting (Section 4.2.4). To the south-east of the Islands the Falklands Microplate is bounded against the Falkland Plateau Basin by a rifted volcanic margin (Richards et al., 1996a; Richards et al., 1996b; Barker, in press). Gravity data suggest that the inferred western margin of the microplate lies close to the eastern side of the Malvinas Basin. It is assumed to be faulted, although interpretation of geophysical information has failed to locate a major fault zone in that area (Richards et al., 1996b). The original nature of the southern margin is obscured by movements along the northern margin of the Scotia Plate (Richards et al., 1996a) but presumably was a rifted or transcurrent margin. The northern margin of the microplate is assumed to coincide with the Falklands Fracture Zone (Richards et al., 1996a) although the D4 thrusting suggests the possibility of differential block movement in the area north of the Islands. Integration of structural geology interpretations of the onshore and offshore areas should help elucidate these matters.

Although the Falklands Fracture Zone is not now seismically active (Figure 6.1), Rapela and Pankhurst (1992) suggested that together with the Gastre Fault System of South America it was once part of a major transcurrent/transform fault zone. In a reconstruction of Gondwana in the Early Jurassic (Figure 6.2), this fault zone separated the Falklands Microplate from the African continent, and the Southern Patagonia block from the rest of South America (Lawver and Scotese, 1987; Storey et al., 1992).

Adie (1952a) first realised that in a reconstructed Gondwana the Falkland Islands most probably lay to the south-east of South Africa (Figure 6.3). He based this remarkable inference (which pre-dated the formulation of the theory of plate tectonics) on the striking similarity of the geological formations in the Falklands to their equivalents in South Africa. The Cape Meredith Complex is very similar to the crystalline basement in the Natal Metamorphic Complex (Section 2.2), the formations of the West Falkland Group all have correlatives in the Cape Supergroup (Section 2.3), and the formations of the Lafonia Group can all be matched within the Dwyka Group, the Ecca Group and the lower part of the Beaufort Group in the Karoo Basin (Section 2.4). The Falklands dolerites have similar compositions to Karoo dykes in South Africa (Section 3.1). Adie's (1952a) reconstruction requires the Falklands to be inverted relative to their present position, aligning the D1 fold belt of the Falkland Islands (Section 4.2.1) with the Cape Fold Belt of South Africa, and the Lafonia Group with the Karoo Basin (Figure 6.3). The reconstruction shown in Figure

6.2 also aligns the other dispersed segments of the Permo-Triassic Gondwanide fold belt, found in the Pensacola Mountains and the Ellsworth-Whitmore Mountains of Antarctica (Curtis and Storey, 1996) and the Sierra Australes and adjacent areas of Argentina (Rossello et al., 1998).

Adie's model is now supported by considerable additional work including palaeomagnetic analysis (Mitchell et al., 1986; Taylor and Shaw, 1989), geochronology of the dolerite dykes (Mussett and Taylor, 1994), stratigraphic studies in the West Falkland Group (Marshall, 1994b), and the Permian (Macdonald et al., 1996), Permian trace fossil assemblages (N H Trewin, written communication, 1997), and structural style and orientation in the Permo-Triassic fold belt (Curtis and Hyam, 1998). There is some debate over the relative positions of East Falkland and West Falkland within Gondwana. Thomas et al. (1997) argue that West Falkland once lay up to 300 kilometres further north than East Falkland, although this suggestion is undermined by their subsequent radiometric analysis (Section 2.2) (Jacobs et al., 1999). Hyam (1998; 1997) considers that the two islands were once displaced by up to 60 kilometres, while Curtis and Hyam (1998) concluded that there has been about four kilometres of orthogonal displacement between East and West Falkland, and only about 3.3 kilometres of dextral strike-slip movement. It is argued in Section 4.2 that the amount of dextral displacement along Falkland Sound is likely to have been between five and 10 kilometres.

Irrespective of the amount of displacement between the several crustal blocks which make up the Falklands Microplate, Adie's reconstruction requires that together they have been rotated clockwise by approximately 180° during Gondwana break-up and subsequent continental drift. Mitchell et al. (1986), Taylor and Shaw (1989) and Ben-Avraham et al. (1993) concluded that the microplate on which the Islands lie was translated southwards by about 500 kilometres and rotated by some 100° to 120°. According to Mussett and Taylor (1994) and Marshall (1994b), this occurred during or after the emplacement of dolerite dykes at about 190 Ma in the Early Jurassic and was completed by the Early Cretaceous, when ocean-floor spreading commenced in the South Atlantic. The rest of the rotation occurred passively as the Falkland Plateau was carried westwards as part of the South American Plate. The Ellsworth-Whitmore Mountains crustal block, which originally lay close to the Falkland Islands (Figure 6.2), was rotated anticlockwise, presumably at about the same time. It now lies between East Antarctica and the Antarctic Peninsula (Marshall, 1994b; Curtis and Storey, 1996). However, there is evidence for transcurrent movement on the Gastre Fault System as early as 220 Ma (in the Late Triassic), as well as for displacement of c. 208 Ma old granites and subvolcanic rocks, so southern South America is thought to have started to move away from Gondwana before the intrusion of the Early Jurassic dolerite dykes (Rapela and Pankhurst, 1992). Indeed, Storey et al., (1996) suggest that it is possible and perhaps more likely that the rotation of the Falklands Microplate occurred during Permo-Triassic Gondwanide dextral transpression.

It appears that the Mesozoic sedimentary basins around the Falkland Islands started to form during the break-up of Gondwana, particularly during the separation of East Antarctica from South Africa, and so that it might have taken place at about the same time as the period of 'active' rotation of the Falklands Microplate. Although the mechanisms of rotation and separation are thus of particular relevance to hydrocarbons exploration, the details of how they took place remain unknown. It is likely that strike-slip displacement on the faults bounding the Falklands Microplate played a major role, together with crustal extension (Ben-Avraham et al., 1993; Marshall, 1994b; Storey et al., 1996). Lorenzo and Mutter (1988) suggested that the Falklands Plateau was lengthened by about 400 kilometres before the South Atlantic started to open in the Early Cretaceous, and Barker (in press) deduced that the Falkland Plateau Basin is underlain by oceanic crust. However, although the rotation model requires that there be some tectonic discontinuity between the Falklands Microplate and the rest of South America, interpretation of seismic and gravity data to the west of the Falklands has failed to recognise any major fault-line or suture in that area (Richards et al., 1996a). Richards et al. (1996b) provide detailed arguments for treating the rotation hypothesis with caution. The dichotomy between the geological and palaeomagnetic evidence, which strongly suggests rotation, and the geophysical evidence, which fails to support the idea, implies that much work remains to be done before a satisfactory solution is found.

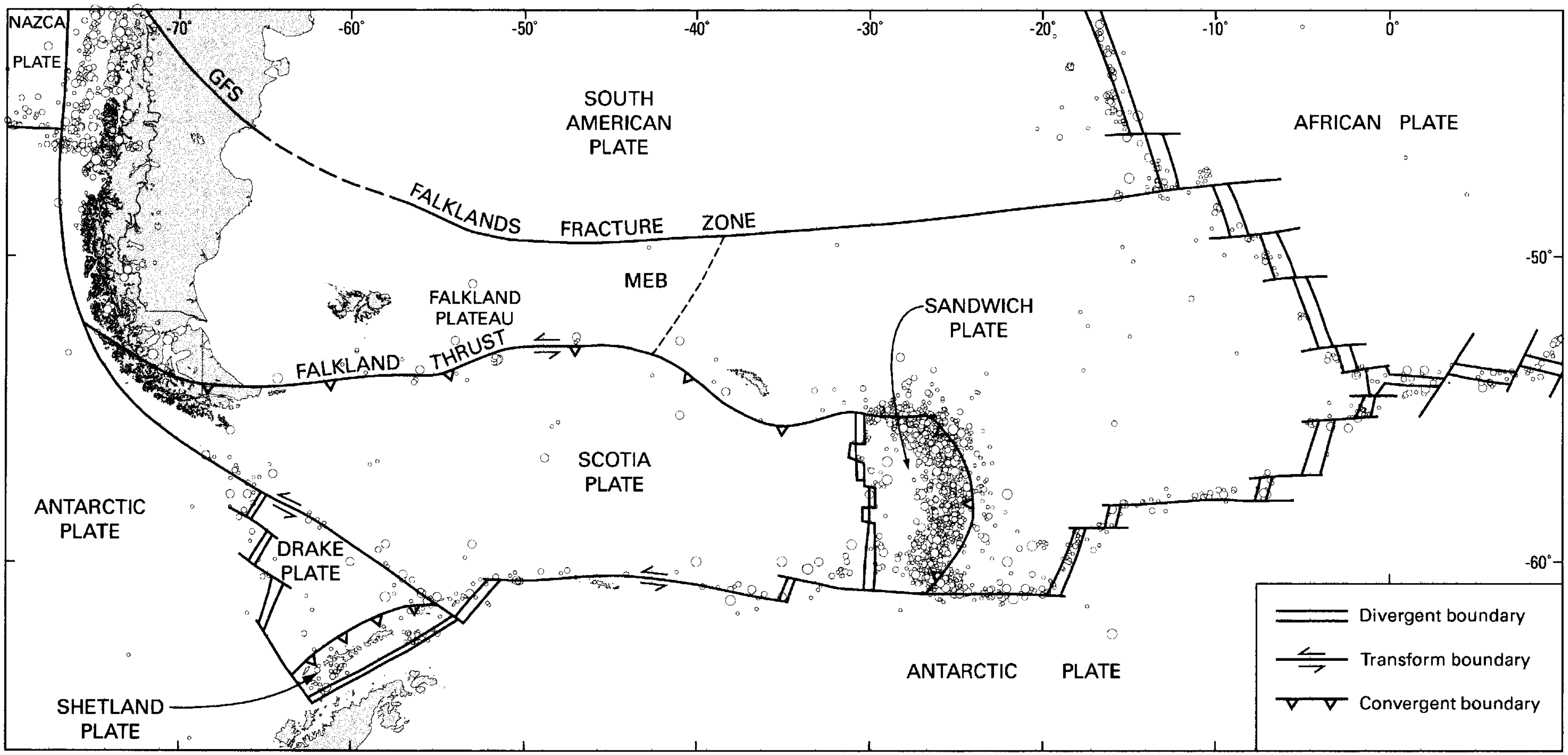


Figure 6.1 Plate tectonic map of the Falklands region

Circles mark epicentres of all recorded earthquakes of known magnitude 4 or greater up to July 1998. Largest circles indicate magnitudes in the range 7 to 7.9. Earthquake data from BGS Global Seismology Database. Other information after Fitzgerald et al. (1990), Marshall (1994a) and Cunningham et al. (1998). Note that the Falklands Fracture Zone and Gastre Fault System (GFS) are not an active plate boundary. MEB: Maurice Ewing Bank.

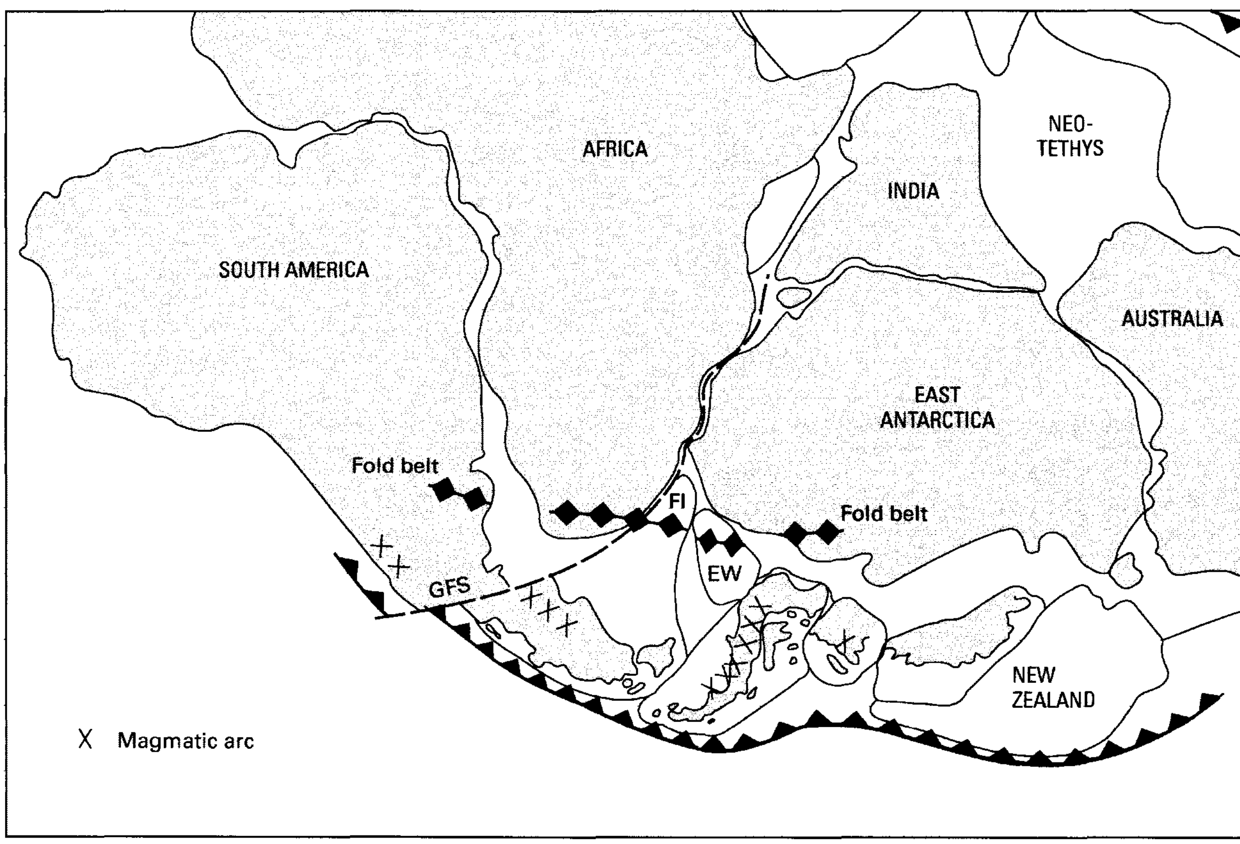


Figure 6.2 Location of Falkland Islands in Gondwana

Simplified reconstruction of Gondwana during the late Triassic/early Jurassic (pre-break-up) after Storey et al. (1992), Rapela and Pankhurst (1992), and Lawver and Scotese (1987). It shows the Falklands (FI) and the Ellsworth-Whitmore Block (EW), with the Gondwanide fold belt, the Gastre Fault System (GFS) and its possible continuation between Africa and East Antarctica as the Agulhas Fault Zone, and the late Triassic to early Jurassic magmatic arc of southern South America and the Antarctic Peninsula.

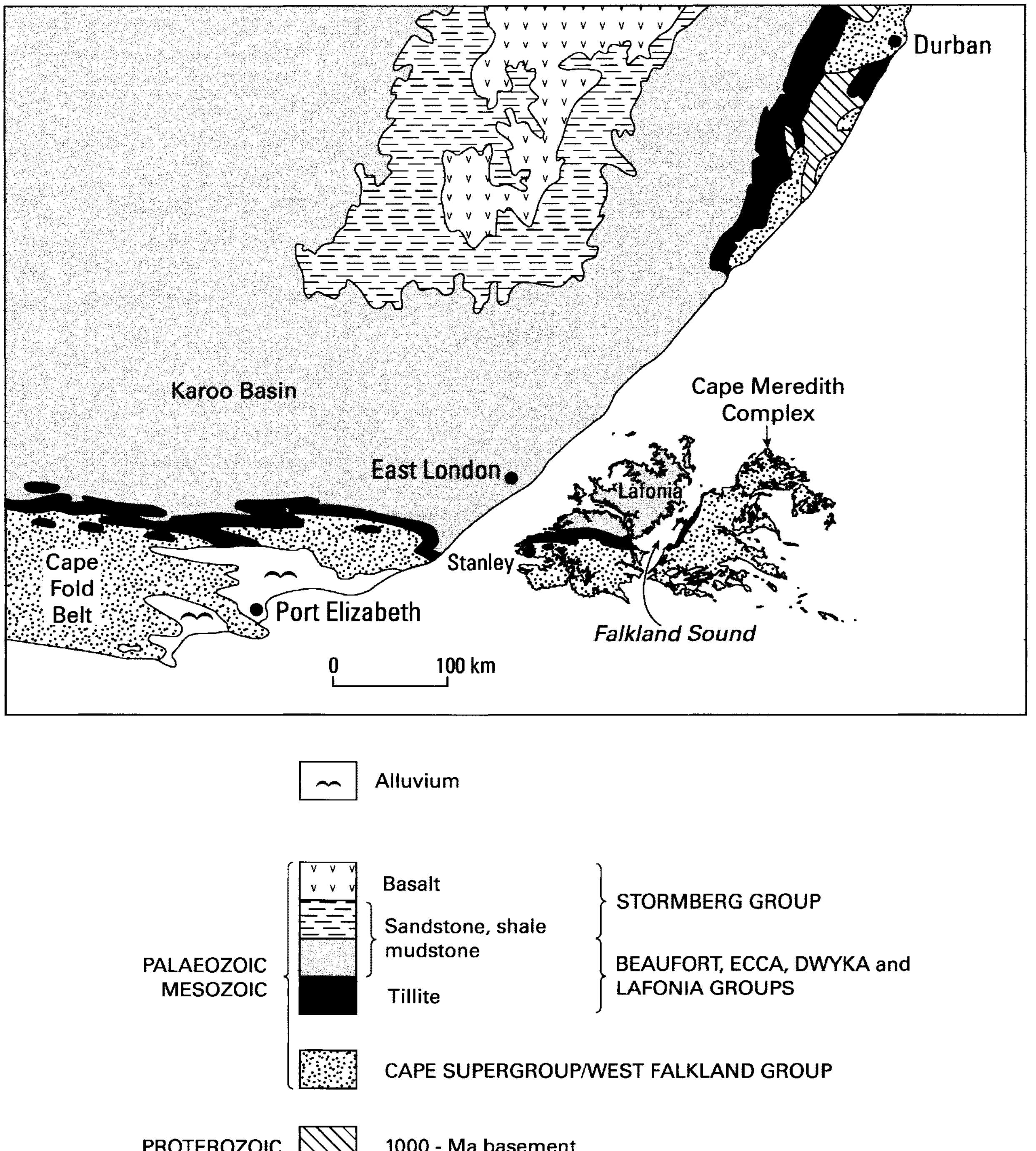


Figure 6.3 Correlation of Falkland Islands with South Africa

Reconstructed position of the Falkland Islands relative to South Africa during the early Jurassic. After Mitchell et al. (1986) and Adie (1952).

7. ECONOMIC GEOLOGY OF THE FALKLAND ISLANDS

7.1 Introduction

The Falkland Islands have long been considered essentially barren of economic mineral deposits. When Baker (1924) undertook a 'comprehensive survey for coal, oil and other minerals', he concluded that the Islands 'appear to be discouragingly deficient in minerals of economic importance'. In 1968, it was 'not considered necessary for a new geological survey to be undertaken since it is unlikely that, even with the use of modern techniques, it would reveal any valuable mineral deposits' (Hansard, 20 Dec 1968, 500). Common sense seems to suggest that if mineralisation is present then some evidence would have been noticed by now. Conversely, in very large areas of the Islands the bedrock is covered by superficial deposits, soil and vegetation. Very few boreholes or deep excavations have been made and until very recently, no survey using modern exploration techniques had been carried out. Reconnaissance geochemical and heavy mineral prospecting commenced in 1997 and continues at the time of writing, but there is still a negligible amount of geophysical data for the onshore area. It cannot be said that the Falkland Islands have been thoroughly prospected for mineral deposits. Until they have, they should not be regarded as barren.

The potential for discovering economic mineral deposits can be approached either by considering the similarities of the geology of the Falkland Islands to that of neighbouring continents, or by appreciating the probable differences. A straightforward comparison with the type and habitat of known mineral showings in South Africa, for example, suggests that there are very few realistic possibilities for economic mineralisation.

On the other hand, while the geology of the Falklands is extremely similar to that in parts of South Africa, it is not identical. The Falklands lay within what was to become a major fault system which eventually divided the continent of Gondwana. Its tectonic development was more complex and protracted than that of adjacent parts of South Africa, and was at times governed by movement on deep-seated fracture systems (Section 4). Moreover, continental break-up was probably driven by the activity of one or more mantle plumes (de Wit and Ransome, 1992; Storey et al., 1996). The volume of dyke intrusion in the Falklands was much greater than in previously adjacent parts of South Africa (Thomas et al., 1997). As a consequence of greater proximity to the axes of rifting between Africa and Antarctica, it is possible that igneous and hydrothermal activity in the Falklands was more extensive than currently suspected. Moreover, variations in facies and sequence thickness, the volume of granitic and metamorphic clasts in the Fitzroy Tillite Formation, and clay maturity analysis suggest that West Falkland was at or close to the eastern margin of the Permo-Carboniferous Karoo Basin. Differences in the underlying basement and in the sedimentary sequence might be conducive to the formation of mineral deposits where none are seen in the south-eastern parts of South Africa.

7.2 Metallic minerals

7.2.1 Gold

Gold has been found in stream sediments in three widely separated areas of the Falklands by Cambridge Mineral Resources plc (Penguin News, 12 March 1999). There are no previous records of its existence in the Islands.

Subeconomic showings of gold have been found in the Cape Fold Belt of South Africa. For example, in the Port Elizabeth area gold occurs with sulphide in quartz veins in the Goudini Formation (equivalent to part of the Port Stephens Formation of the Falklands) (Toerien and Hill, 1989). Similar occurrences might be present in the area of D1 folding and thrusting, especially where metamorphic grades are highest in the centre of East Falkland (Sections 4.2.1, 4.3). No sulphide-bearing quartz veins have been reported, however.

The Devonian sedimentary rocks in the Cape Fold Belt and the Falkland Islands probably include detritus carried from the Kaapvaal Craton of South Africa, which is well known for extensive deposits of gold-bearing sandstones and conglomerates. It is thus possible that concentrations of heavy minerals in the sandstones of the Port Stanley Formation and the Port Stephens Formation include gold.

7.2.2 Iron

Small nodules of iron minerals, either limonite or siderite, occur sporadically at the base of peat or in the soil on clay patches. They represent precipitates from acidic waters in the peat. A sample of iron ore (siderite) found to contain about 59 per cent iron oxide (Anon., 1912) was probably an occurrence of this sort.

Ferruginous sandstones occur locally in the Port Philomel Formation, for example on Saunders Island to the west of The Neck, but are of no economic significance.

7.2.3 Base metals

No traces of other metallic mineralisation have been seen or reported in situ in the Falklands. A sample of copper ore sent to the Imperial Institute for analysis was found to be composed of mixed chrysocolla and copper carbonate and to contain 51.7% CuO, with 0.18% NiO (Anon., 1912). The collection site of this specimen was not stated, and it seems probable that the sample originated from a ship's cargo. Ships carrying cargoes of copper ore from Chile have been wrecked in several locations, including Bull Point.

Some Jurassic dolerite dykes (Section 3.1) contain traces of pyrite, but no other sulphide minerals have been recorded. A nickel-copper sulphide deposit is associated with the Jurassic-aged dolerite intrusion at Insiswa, near Kokstad in South Africa. This is relatively close to the past position of the Falkland Islands, but the Insiswa Intrusion is an enormous differentiated body up to 1000 metres in thickness (Maske and Cawthorn, 1986). In the Falklands, the only dolerite intrusion which might be differentiated is the South Fur Sill.

7.3 Non-metallic minerals

7.3.1 Mineral sands

Minerals such as zircon, garnet and rutile commonly occur in sandstones in the West Falkland Group and to some extent in the Lafonia Group (Knox, 1997; Knox and Aldiss, 1999). Many sandstones in the Port Stanley Formation contain visible concentrations of heavy minerals (Section 2.3.4). The Fitzroy Tillite Formation also contains detrital garnet and other heavy minerals. Concentrations of these heavy minerals eroded from the bedrock formations occur in some beach sands (Adie, in Greenway, 1972). It is conceivable that economically significant heavy mineral deposits might be present on modern beaches or in raised beach deposits. However, although raised marine erosion surfaces are extensive most are covered in solifluction deposits and peat, and the extent and composition of the raised beach deposits is little known (Section 5.2.10)

7.3.2 Glass sands

The sand on some Falklands beaches is notably clean and white. Baker (1924) investigated some for their potential as glass sand, but concluded that their quality was inferior to that of more accessible supplies. Although Adie (in Greenway, 1972) found some beach sands of acceptable purity, he concluded that they would require expensive pre-treatment.

7.3.3 Clay

Smooth lacustrine clay found in some closed ramparted hollows (Section 5.2.4) seems to have potential for small-scale pottery manufacture. Similar clays might be present in more extensive lacustrine deposits elsewhere (Section 5.2.9). Clays in raised marine deposits are known from three widely spaced localities (Section 5.2.10) and could be more voluminous than the lacustrine deposits. Their suitability for pottery was being tested at the time of writing.

7.3.4 Phosphates

Phosphatic concretions occur in the Prince Albert Formation of South Africa, a part equivalent of the Port Sussex Formation (Section 2.4.3). They are of no economic significance (Toerien and Hill, 1989; Stratten, 1986) and so far none have been noted in the Falklands.

7.4 Radioactive minerals

Sub-economic deposits of uranium are known in certain sandstones in the Karoo Supergroup of South Africa, the most important being in the Beaufort Group (Turner, 1985; Le Roux and Toens, 1986). The more significant deposits all occur in the western part of the basin, remote from the previous position of the Falkland Islands.

Nevertheless, the factors thought to be required for the occurrence of uranium mineralisation could be present in the Bay of Harbours Formation part of which is correlated with the lower Beaufort Group (Section 2.4.5). There are potential sources in the volcanic detritus found throughout the Lafonia Group (Section 2.4) or in granitoid rocks of the Cape Meredith Complex which were probably exposed nearby during Lafonia Group deposition (Sections 2.2, 2.4.2, 4.3). Proximity to the Gondwanide fold belt could have provided sufficient heat-flow to stimulate the necessary circulation of fluids. Channel sandstones are present in the Bay of Harbours Formation, although their porosity is probably not great and they are most common in the west of Lafonia, where the metamorphic grade is lowest (Section 4.3). However, there is no evidence that uranium mineralisation is present anywhere in the Falklands.

Xenotime (a mineral of yttrium phosphate) can include proportions of uranium, thorium and the rare earth elements. It has been noted as an accessory mineral in a pegmatite from Cape Meredith (Section 2.2) (Baker, 1924). The rare earth phosphate mineral monazite, which can also contain appreciable amounts of thorium, is widespread in small concentrations in sandstones of the West Falkland Group and to a lesser extent in those of the Lafonia Group (Knox, 1997; Knox and Aldiss, 1999). There is no evidence that either mineral is likely to occur in economically significant concentrations.

7.5 Mineral fuels

7.5.1 Hydrocarbons

There is a chance that worthwhile reservoirs of oil or natural gas occur offshore, deep beneath the sea-floor. The prospects are good enough to have attracted investment in offshore exploration by several international oil companies from 1996 to 1998 (Richards and Fannin, 1994; Richards et al., 1996a; Richards and Fannin, 1997).

As concluded by Baker (1924) and Adie (in Greenway, 1972), there seems to be no significant chance that oil or gas will be found onshore the Falkland Islands. A licence to prospect for hydrocarbons in the Douglas area was issued to Firstland Oil and Gas plc in 1984. A report on the geology of the area was prepared for them by Hunting Geology and Geophysics but remains confidential. No further work was done. A sample of bitumen from the Falklands (Anon., 1912) is probably not of local origin. Baker (1924) concluded that all reported occurrences of bituminous material on the Falklands stemmed from oil shale being shipped from Australia.

The only potential hydrocarbon source-rock exposed onshore is the Black Rock Member. This can be correlated with the carbonaceous mudstones of the Whitehill Formation of South Africa, and the Irati Shale Formation of South America, which is an economic oil shale (Marshall, 1994a). The Black Rock mudstones are carbonaceous, although more so in East Falkland than in West Falkland (Section 2.4.3.2). However, a large part of the Islands is over-mature with respect to hydrocarbon development, although the west is mostly in the gas window (Section 4.3). The portion of the Black Rock Member buried beneath the younger Lafonia Group between the Goose Green Axis and the Falkland Sound Fault (Section 4.2.1) probably retains the most potential as a hydrocarbons source: it can be expected to relatively carbonaceous and it lies in an area of late diagenetic zone metamorphism. However, field and microscopic assessment suggests that porosity is generally poor in sandstones throughout the sequence.

Observation during the present survey suggests that many phenomena which might be suspected to be onshore oil seepages are in fact seepages of iron-rich water from within the peat.

7.5.2 Coal

Reports of coal in the Falklands date back at least as far the Challenger expedition of the 1870's, but none have been substantiated. Most reports stem from appearance of carbonaceous mudstones in the Black Rock Member (Section 2.4.3.2), particularly where they have been tectonically sheared (as at Port Sussex). Some of the Black Rock mudstones contain up to 40 per cent carbon, but none are combustible.

Permo-Carboniferous coals are widespread in Gondwana. Baker (1924) examined the prospects for coal in the Lafonia Group with particular care. He concluded that it correlates with a part of the South African Karoo Basin in which there are no economic coals. Adie (in Greenway, 1972) reached the same conclusion. At present coal is only exploited in the north-eastern part of the Karoo Basin, or in the Triassic Molteno Formation, which is younger than any part of the Lafonia Group (Smith and Whittaker, 1986) (Johnson et al., 1996). Although plant fossils are common in the Bay of Harbours Formation, no trace of coal was found during the present survey.

Small lenses of coal occur in the Port Philomel Formation (Section 2.3.3) but are of no economic significance.

Fragments of coal which can sometimes be found on beaches are therefore most likely to have come from ships. For example, Wreck Bay in San Carlos Water is named for a coaler which was abandoned there. Pieces of good quality coal can still be found in the vicinity. The Clarence S Bennet, loaded with coal, burnt out in Fox Bay in 1903 (Southby-Tailyour, 1985).

7.5.3 Peat

Peat (Section 5.2.7) continues to be used as a domestic fuel in many places throughout the Islands, although to a lesser extent than in the past.

Early investigations (Anon., 1907) concluded from an analysis of four peat samples that they would probably yield a compressed fuel of good quality. According to figures quoted by Adie (in Greenway, 1972), the calorific value of the Falklands peats compares well with those of Ireland.

7.6 Constructional materials

7.6.1 Bulk rock

In the past, crushed rock has been supplied for construction in the Stanley area from small quarries in the Stanley Quartzite (Section 2.3.4) near the present power station, on Sapper Hill, in the Moody Valley, on Navy Point and at Mary Hill near Stanley Airport (Rosenbaum, 1985; Rosenbaum, 1984). In recent years, local needs have been supplied by the Stanley Quartzite quarry at Pony's Pass [VC 341 697] and the Fitzroy Tillite quarry at the Frying Pan [VC 086 589].

Several quarries were opened to serve the Mount Pleasant Airport construction project. These are the Pleasant Peak Quarry [UC 97 61] (Stanley Quartzite), the Mare Harbour Quarry [VC 00 49] (Brenton Loch Formation sandstone) and the Comoda Ditch Quarry [UC 97 57] (Fitzroy Tillite) (Kenrick, 1987).

Materials for construction of the rural road system were mostly taken from chains of borrow-pits along the line of the roads. Most pits were subsequently landscaped but some remain open, notably at Canada Runde [UC 83 57], between MPA and Darwin. Local experience has shown that the Fox Bay Formation tends to provide the most suitable material. Other formations can be of uneven quality, necessitating more extensive trial-pitting. For example, some parts of the Port Stephens Formation comprise alternations of thin hard sandstone beds and friable, clayey weathered rock. When excavated and spread, this combination can give a readily compactible mixture which forms a smooth durable road surface. Elsewhere, the same formation can yield a sandy mix which reverts to a slurry under winter conditions. The Port Stanley Formation generally yields the least suitable material for road construction.

7.6.2 Sand and aggregates

Sand for building purposes in the Islands is usually taken from aeolian deposits. These are most common behind west-facing beaches (Section 5.2.6) and so are not widespread in eastern East Falkland, where demand is greatest. In the Stanley area, sand is taken from the relict sand dunes near Cape Yorke (Rosenbaum, 1985). Wind-blown sand and beach sand are not ideal materials for concrete and mortar mixes, as they tend to be composed of well-rounded grains of fairly uniform size, and to contain varying amounts of organic material.

Fine aggregates produced from crushed quartzite were used in the construction of Mount Pleasant Airport (1983-86), but were found to produce a harsh mix. Sand from Bertha's Beach was also used in limited quantities. In addition, 10 000 tonnes of Leighland sand was imported from the UK for blending with the locally available fine materials (Skene and Brice, 1987).

Small exposures of sand and gravel can be seen in alluvial deposits, although they are generally obscured by a surface layer of peat (Section 5.2.8). For example, sand occurs in the south bank of the Murrell River at Lower Pass [VC 3589 7728]. It is possible that trial pits or shallow boreholes in the alluvial deposits beside the larger rivers would reveal sand and gravel suitable for construction. As noted in Section 5.2.10, sea level was once as much as 50 metres lower than at present. It is therefore probable that alluvial sands and gravels also occur offshore.

7.6.3 Building stones

Several of the older buildings in Stanley have been built using Stanley Quartzite (Section 2.3.4) although it is considered difficult to work. One of the most robust examples is the old smithy in the Public Works Department dockyard area. The quartzite has been used in combination with brick in some other older buildings. It is also widely used for garden walls or between the load-bearing piers beneath pre-fabricated houses. Slabs of weathered rock, quarried blocks and smooth rounded beach boulders have each been used. A particularly fine example of the use of Stanley Quartzite is seen in a low wall enclosing the frontage of the Tabernacle in Barrack Street. The blocks are mostly of square or rectangular outline, with neatly dressed faces. They are composed of an unusual variety of the rock which displays thin planar bedding and plane lamination, enabling it to be split into regular slabs as little as three centimetres thick. This rock is reported to have been taken from Charles Point on the northern shore of Port William.

Stone lintels and mullions used in the construction of Government House are said to have come from sandstone of the Fox Bay Formation in the Port Louis area, in at least one instance including the casts of marine invertebrate fossils.

Rough-hewn kerbstones of a coarse-grained pink granite, believed to have been ships' ballast, have been incorporated in a wall on the south of John Street, Stanley, near its junction with Dean Street. Similar rock has been seen at Goose Green and on Carcass Island (M Keenleyside, oral communication, 1998).

Many of the other settlements have older houses built partly or mainly of stone. A house immediately east of the jetty at Goose Green is mostly built of sandstone from the local Bay of Harbours Formation. There is a small stone quarry in a massive fine-grained sandstone on the coast between Goose Green and Darwin. Similar sandstones were used in buildings in North Arm and in stone corrals at Darwin and Kelp Harbour. None show any susceptibility to frost damage.

The church and old house on Keppel Island are made of sandstone which appears to be of local derivation in the Port Philomel Formation. Very few beach boulders have been used, suggesting that a small quarry was opened for the purpose, perhaps in the cliff nearby.

7.6.4 Flagstone

Plane-bedded fine-grained sandstone (flagstone) from the Fox Bay Formation (Fox Bay stone) has been much used for paving, and also as a decorative facing stone. The stone quarry at Fox Bay is situated on the cliff top about one kilometre south of the main settlement [TC 888 387]. It is mentioned by Halle (1912). Similar flagstones are widely distributed in the Fox Bay Formation, occurring in many of the borrow pits beside the road between Fox Bay and Little Chartres, on Pebble Island due south of Middle Peak, and locally at San Carlos. Flagstones also occur in the Egg Harbour Member, as exposed on the west coast of Lafonia. The size and quality of these occurrences is unknown, but they would be worth further investigation if only for local use.

7.6.5 Ornamental stone

Ornamental granite has been used in the construction of two monuments in Stanley. A pale grey granite used in the monument which commemorates the Battle of the Falklands in 1914 was described in the Falkland Islands Church Magazine of March 1927 as 'Cornish Granite'. This stone does resemble some parts of the Bodmin Granite. Pink granite from Merrivale Quarry, Dartmoor, was used to build the 1982 Liberation Monument. Petrographically it is of note for the sporadic occurrence of rapakivi-textured megacrysts: crystals up to three centimetres in diameter of pinkish potassium feldspar enclosed in a narrow rim of grey plagioclase feldspar. The 1982 monument is surrounded by York stone paving (very similar to Fox Bay stone), also from the United Kingdom. The adjacent wall is built of Stanley Quartzite taken from battle areas.

The Cross of Sacrifice, which commemorates the Falklands dead of the 1914-1918 war, is built of rather undistinguished medium to coarse-grained subarkosic sandstone. The Church Magazine of May 1925 describes it as 'Stancliffe stone'. This refers to sandstones from the Carboniferous Ashover Grit (Millstone Grit Group) from Darley Dale near Matlock, Derbyshire, UK.

7.7 Agricultural minerals

7.7.1 Limestone

Indigenous deposits of lime-rich material could be a valuable aid to land improvement in the Falkland Islands. Although the Devonian Fox Bay Formation was deposited in a marine shelf environment, it includes negligible amounts of limestone (Section 2.3.2). No limestones are known from any of the other bedrock formations.

Occurrences of limy material, taken to be detritus of the calcareous bryozoan *Lithothamion,* are widespread on the shores of the Falkland Islands. This material is found as encrustations on shoreline rock exposures and on loose rock fragments, as limestone cobbles up to about 15 centimetres in diameter (which may or may not have a rock nucleus), and as irregular brittle flakes ranging up to about one centimetre in diameter. Small accumulations of these flakes, here referred to as 'lime gravel' and 'lime sand' are reasonably common on sheltered beaches, but most are far too small to have any practical use. Calcareous detritus from molluscs, brachiopods and foraminifera also occurs.

Lithothamion detritus forms significant sub-tidal banks off Cornwall (south-west United Kingdom) and Brittany (western France) which have been dredged for agricultural purposes. Up to 1974, some 600 000 tons had been dredged from the Brittany coast. The bryozoan typically lives in water depths of about 20 metres and eroded detritus forms small banks resulting from the action of waves and tidal currents. In south-west England this detritus is known as 'maerl' (note that the geological material known as 'marl' is something different).

Relatively large deposits of unconsolidated lime-rich detritus form beach bars on the north side of Findlay Harbour near Chicho Point [UC 228 361]; at 'The Sandbed' on the south side of Wreck Rincon [UC 223 326], and at the mouth of Ruggles Arroyo, close to The Sandbed [UC 2265 3225]. These sites are on sheltered inlets leading from the eastern side of Falkland Sound, up to 25 kilometres north-west of North Arm. Reconnaissance survey indicates that at least 10 000 tonnes of lime gravel exists on these three beaches above low water mark. Greater quantities of lime-rich material exist below the tidal zone: aerial photographs (November 1956) suggest that drifts of lime gravel or sand of unknown thickness could extend about 100 metres offshore at Chicho Beach and 1000 metres at the

Sandbed. Significant deposits of similar material probably exist on beaches elsewhere. The processes giving rise to these deposits are evidently active at the present time, so some degree of replenishment can be expected.

The limy detritus is relatively soft and tends to break down during transport into fine-grained material, possibly including lime mud. Such mud would settle only in deep water, below wavebase and where tidal currents are slack, or in very sheltered back waters. This may well be the origin of a lime-rich 'blue clay' (reported to contain some 39% lime) discovered in Stanley Harbour by the Challenger expedition.

The occurrence of partly consolidated shell limestone at Shell Point, Fitzroy, was noted by Adie (1953). This is reported to be largely composed of molluscan debris. Together with overlying calcareous sand this limestone nowhere exceeds two metres in thickness. This deposit is now the site of a small market garden.

7.7.2 Guano

Although the Falkland Islands is noted for its populations of seabirds, no worthwhile accumulations of guano are known to occur. Strange (1983) notes that attempts in the 19th century to export guano from New Island and elsewhere failed, the guano being deficient in nitrogen and in quantity. This is corroborated by analyses of five specimens of penguin guano from Cochon Island and Kidney Island (Anon., 1914).

7.8 Hydrogeology

At present, water supplies throughout the Falklands are taken from streams or springs, with no use being made of groundwater pumped from boreholes. Although small lakes and boggy ground are commonplace in the Falklands, the annual rainfall is not high. Moreover, wind tends to be greatest during the dryer summer months. Thus some settlements experience periodic water shortages. These could probably be alleviated through the development of supplies from groundwater.

The primary porosity of all the bedrock formations can be expected to be low to negligible. Conversely, joints and fractures usually seem to be well-developed, especially in the areas of folding and faulting (Section 4). In eastern East Falkland a widely developed cleavage may also contribute to secondary porosity, especially in the Fitzroy Tillite Formation and the Black Rock Member.

Many of the dolerite dykes are very deeply weathered (Section 3.1). They can thus be expected to form effective aquicludes. However, where they remain unweathered, fracturing within and adjacent to dolerite dykes can promote groundwater flow. Small fresh-water seepages have been observed at the margins of dolerite dykes in several areas.

As noted in Section 5.2.9, many ponds are 'perched' above an impermeable layer of solifluction debris, and are not in hydraulic continuity with their neighbours. Such ponds are poor candidates as perennial water sources, as there will be very little local recharge to replace water pumped or drained from the pond.

The pH of surface water varies considerably. Weller (1975) found that ponds in peat are very acid (pH 4.0 to 5.3), whereas a sand-bottomed pond had a pH of 6.0. Clark et al. (1994) confirm that the majority of lakes and rivers have low pH (as low as 3.1), but state that some ponds (including Lake Sulivan) and water courses have pH close to neutral (6.5 to 7.5). They note that some streams emerging from beneath stone runs or other rocky substrates varied between neutral and slightly alkaline. They also found that ponds separated by a strip of ground as little as one metre across, with or without different water levels, often had significant differences in pH. Slightly alkaline spring water has also been noted at Spring Point, and at Chartres.

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9. APPENDIX 1: PROJECT METHODOLOGY

The Falkland Islands Geological Mapping Project commenced in early 1996 and mapping was completed in 1998. The prime objectives were to revise the geological map of the Falkland Islands at 1:250 000 scale and to produce an up-to-date report on the geology. This work was carried out on behalf of the Falkland Islands Government by the authors of this report: Don Aldiss (British Geological Survey) and Emma Edwards (Falkland Islands Government), based in Stanley. Geological field observations were made in selected areas throughout the Islands during March-June 1996, November 1996 to May 1997, December 1997 and January 1998. Project methodology and rationale are described in more detail by Aldiss (1997).

Given the size of the area, and that much of it is relatively difficult of access, the geological map that could be produced within the scope of the project is only of a reconnaissance nature. It was recognised at the start that complete aerial photographic cover, preferably together with satellite images, would be required to make this task possible, let alone to improve significantly the existing reconnaissance geological map of Greenway (1972). In the event, the necessary coverage was provided by the same series of aerial photographs as used by Greenway, flown in 1956 at nominal 1:25 000 scale. Very little cloud appears on these photographs and obscures only small areas in the Smoko Mountain area of the Wickham Heights. To facilitate annotation, an overlay of Frisk 'K-trace' was mounted on alternate prints. In the office, photogeological interpretation was carried out using a Wild ST4 mirror stereoscope mounted together with a fluorescent strip light on a parallel motion trolley, and equipped with fold-down telescopic eyepieces giving 3x magnification. Small portable stereoscopes were used in the field.

Geological mapping relies on the fact that the form of many topographic features and vegetation patterns reflect variations in the underlying geological formations, even where very little bedrock is visible at the surface. Viewed stereoscopically, aerial photographs reveal the topographic features of an area, which can then be interpreted in terms of the local geology. However, although aerial photographs are an indispensable tool for regional geological surveys, their full potential is realised only when they are used in conjunction with fieldwork.

The present project set out to produce a geological map by closely integrating photogeological interpretation with ground observation. This required identification on the aerial photographs of contrasts in the expression of each of the geological units which have been recognised on the ground. In practice, it was also found that some previously unrecognised subdivisions of the sedimentary rock sequence could be recognised on the basis of photogeological criteria, rather than from ground observations. The aerial photographs were also used a base-map on which to record field observations. Field observation and photo-interpretation were then used together, enabling geological boundaries to be traced across the photographs in much the same way as they would be traced across country during a geological field survey.

The revised geological map combines field observations with detailed re-interpretation of 1:25 000 scale aerial photographs, together with interpretation of satellite images (Landsat TM and SAR) at 1:100 000 and at 1:250 000 scales. Satellite image processing was carried out by Andrew McDonald (BGS Geospatial Information Systems). Data from channels 4, 5 and 7 of two Landsat TM scenes (acquired May 1987) were processed according to standard techniques of edge enhancement and contrast stretching, in order to enhance the images for the purposes of geological mapping. In addition, the processed images were geo-corrected to coincide with the DOS 1:250 000 topographic map of the Islands. Colour photographic prints were then provided at 1:250 000 and 1:100 000 scales for geologists' interpretation. Repeated searches failed to find any Landsat TM scenes in which the northern part of East Falkland is not obscured by cloud. Therefore synthetic aperture radar (SAR) data acquired from the FUYO-1 satellite was used to provide an overview of East Falkland.

Map compilation was at 1:50 000 scale (Figure 9.1). Detail interpreted on aerial photographs (at 1:25 000 scale) was transferred to the 1:50 000 scale topographic maps by eye, for the sake of speed. As the maps were made from the same series of aerial photographs, the agreement of detail is generally excellent. However, loss of accuracy can be expected on steep mountain slopes, where there are few drainage channels and little medium-scale topographic variation expressed in the contours. Furthermore, minor changes in the position of a line in steep ground can have a large effect on the plotted height of the line and therefore on the local geological structure which can be inferred from the outcrop patterns. The geologists' compilations were captured digitally at 1:50 000 scale by the British

Geological Survey Cartographic Unit. Reduction to the presentation scale of 1:250 000 was carried out in the digital environment.

A reconnaissance map of the superficial deposits was compiled at 1:250 000 scale. The duration of the project was insufficient to survey the individual deposits, and in any case few are extensive enough to be portrayed at such a small scale. On the other hand, there are extensive areas with uniform *assemblages* of superficial deposits. For example, most of the Wickham Heights has a superficial cover of solifluction deposits of various types (including several forms of stone run) together with minor blockfield, scree, and alluvium. Conversely, most of Lafonia has a negligible superficial cover except for valley-bottom deposits (alluvium with some peat and minor solifluction deposits). The common assemblages of superficial deposits were defined on the basis on field observation. They were then delineated at 1:250 000 scale by inspection of aerial photographs and Landsat TM satellite images.

The map of the solid geology has been published in two sheets at 1:250 000 scale. The 1:50 000 scale geological maps are available in digital form on CD-ROM, as is the map of the superficial geology, which remains unpublished. The original geologists' compilations and the annotated overlays from the aerial photographs are held by the Department of Mineral Resources, Stanley.

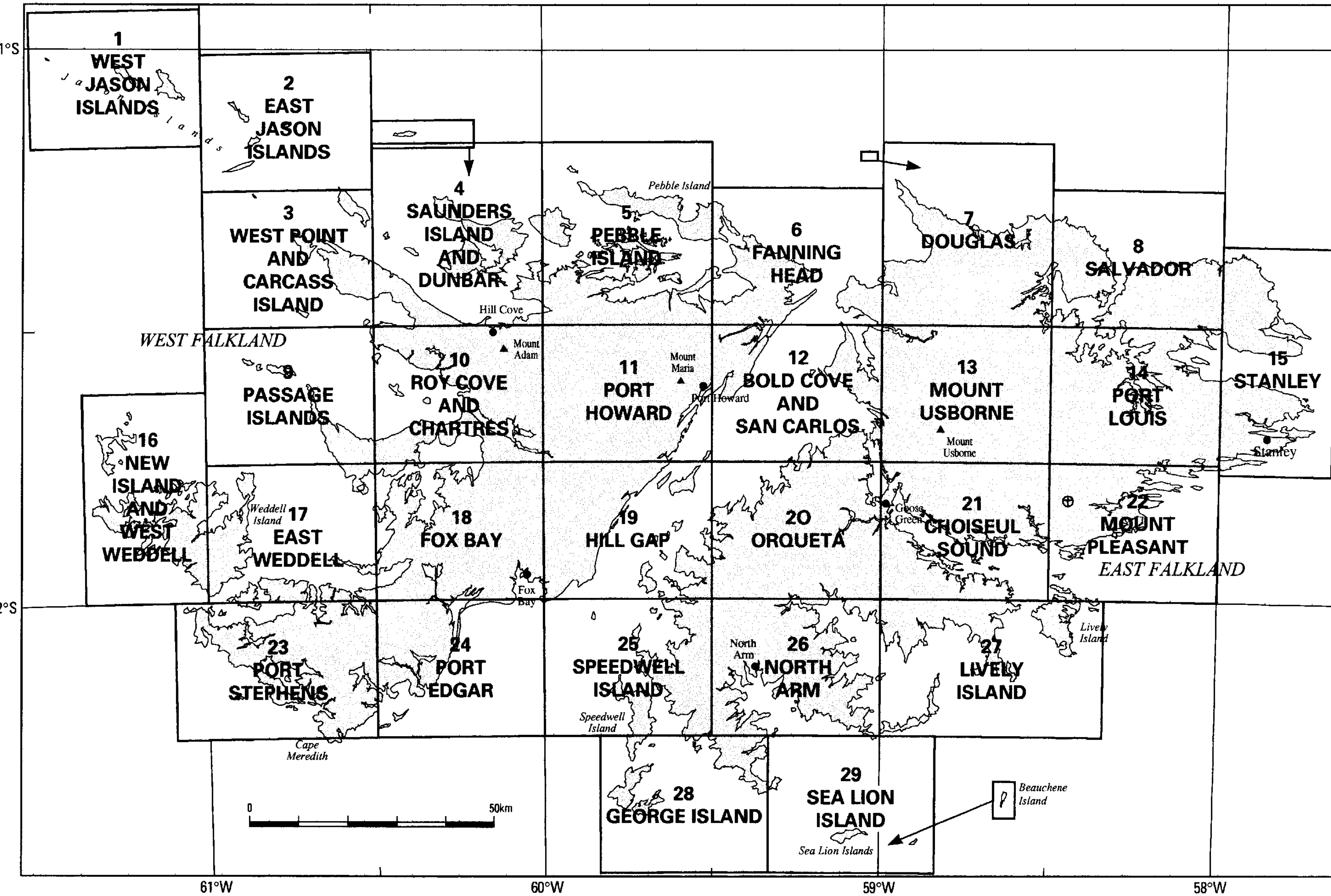

Figure 9.1 Names of the 1:50 000 scale topographic sheets