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Lithostratigraphy of volcanic and sedimentary sequences in central Livingston Island, South Shetland Islands

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Abstract: Livingston Island contains several, distinctive sedimentary and volcanic sequences, which document the history and evolution of an important part of the South Shetland Islands magmatic arc. The turbiditic, late Palaeozoic–early Mesozoic Miers Bluff Formation (MBF) is divided into the Johnsons Dock and Napier Peak members, which may represent sedimentation in upper and lower mid-fan settings, respectively, prior to pre-late Jurassic polyphase deformation (dominated by *open* folding). The Moores Peak breccias are formed largely of coarse clasts reworked from the MBF. The breccias may be part of the MBF, a separate unit, or part of the Mount Bowles Formation. The structural position is similar to the terrigenous Lower Jurassic Botany Bay Group in the northern Antarctic Peninsula, but the precise stratigraphical relationships and age are unknown. The (?) Cretaceous Mount Bowles Formation is largely volcanic. Detritus in the volcaniclastic rocks was formed mainly during phreatomagmatic eruptions and redeposited by debris flows (lahars), whereas rare sandstone interbeds are arkosic and reflect a local provenance rooted in the MBF. The Pleistocene–Recent Inott Point Formation is dominated by multiple, basaltic tuff cone relicts in which distinctive vent and flank sequences are recognized. The geographical distribution of the Edinburgh Hill Formation is closely associated with faults, which may have been reactivated as dip-slip structures during Late Cenozoic extension (arc splitting).

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Introduction

Livingston Island, in the South Shetland Islands (Fig. 1), contains several major lithological units, which together characterize an important part of the South Shetland Islands magmatic arc. In particular, they include pre-volcanic basement, arc volcanic sequences, plutonic intrusions and post-subduction volcanic rocks. By contrast with much of the Antarctic Peninsula, outcrops on Livingston Island are generally more accessible and are locally well exposed. Previous studies in the central part of the island were typically of a reconnaissance nature only and contacts between the sedimentary and intrusive assemblages were mainly inferred to be vertical faults (Hobbs 1968, Smellie *et al.* 1984). In most cases, interpretation of the original relationships was speculative. Moreover, published details of the individual rock groups are generally limited and scattered throughout the literature (e.g. Smellie *et al.* 1984, Smellie 1990, Muñoz *et al.* 1992, Pallàs *et al.* 1992, Doktor *et al.* 1994) and most of the major lithostratigraphical divisions are undefined.

In this paper, the lithostratigraphy of volcanic and sedimentary rocks in central Livingston Island is revised and formally defined for the first time (Table I), and possible origins for the different rock groups are suggested. Studies such as this one will enable workers to arrive at a common stratigraphy and nomenclature and are an essential prerequisite to a fuller understanding of the origin and evolution of magmatic and sedimentary complexes in the northern Antarctic Peninsula region.

Geological background

The oldest rocks on Livingston island comprise the strongly deformed, turbiditic Miers Bluff Formation (MBF), which is essentially restricted to Hurd Peninsula (Hobbs 1968, Fig. 1). The MBF has been correlated with the lithologically and structurally comparable Trinity Peninsula Group (TPG) in northern Antarctic Peninsula. Its depositional age is unknown but can be constrained, by a combination of field relationships and isotope chronology of clasts and sedimentary beds, to lie between late Carboniferous and early Jurassic (summarized in Smellie *et al.* 1984, Hervé *et al.* 1991, Arche *et al.* 1992b, Willan *et al.* 1994). The tectonic setting of the MBF and TPG is uncertain and contentious. Although a geographical association with a magmatic arc is undisputed, the structural position of the sequences (i.e. whether fore- or back-arc) is unknown (e.g. Dalziel 1984, Storey & Garrett 1985, Smellie 1991).

Most of the outcrops on Livingston Island consist of weakly deformed, volcanic and volcaniclastic rocks. Well-dated sequences are exposed only at Byers Peninsula and Williams Point (Fig. 1). They indicate that, following late Jurassic marine sedimentation, volcanism became widespread in Livingston Island during the Cretaceous (especially mid-

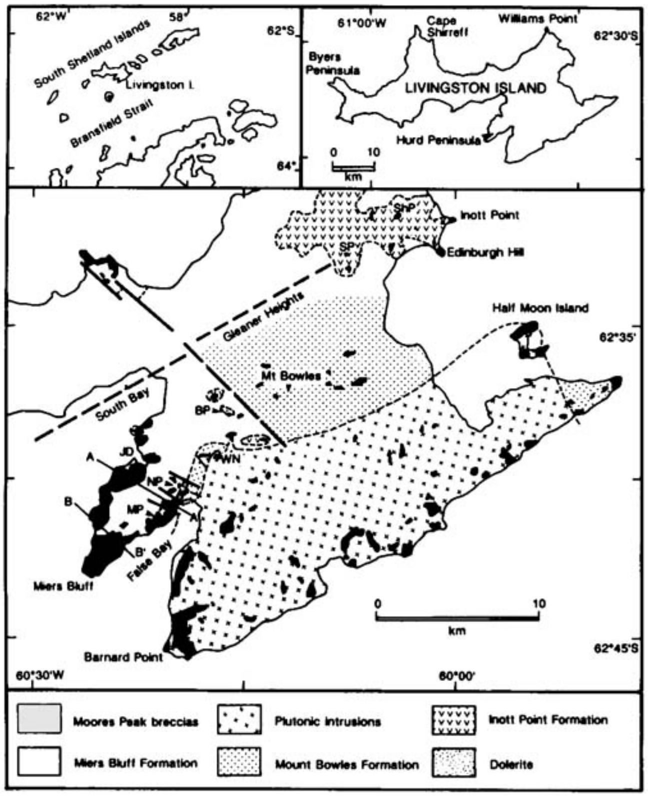

Fig. 1. Geological sketch map of eastern Livingston Island. The locations of composite schematic cross sections shown in Fig. 3 (A-A', B-B') are also indicated. Faults shown are inferred from geomorphological features and satellite images. Rock exposures shown in black. For clarity, some of the smaller outcrops have been exaggerated. Abbreviations: JD = Johnsons Dock; MP = Moores Peak; NP = Napier Peak; WN = Willan Nunatak; BP = Burdick Peak; SP = Samuel Peak; ShP = Sharp Peak.

Cretaceous times: Smellie *et al.* 1984, Chapman & Smellie 1992, Crame *et al.* 1993). Late Cretaceous dolerite sills (74–79 Ma: Smellie *et al.* 1984) are also widespread on the island.

Plutonic rocks related to the magmatic arc comprise several small tonalite stocks on Hurd Peninsula and gabbro intruded by tonalite (Barnard Point tonalite), which form the mountainous terrain in the south-east part of the island. The small stocks on Hurd Peninsula are commonly assumed to be apophyses of the larger Barnard Point tonalite pluton, although the individual outcrops are geographically isolated, texturally distinct and undated. The Barnard Point tonalite has yielded a K-Ar isotopic age of 46 ± 1 Ma (Smellie *et al.* 1984), interpreted as the age of emplacement, whereas an unrelated tonalite on Half Moon Island nearby was dated as 102 Ma (Grikurov *et al.* 1970).

Dykes are present in most outcrops in central Livingston Island. By analogy with dated hypabyssal intrusions at Byers Peninsula (128–74 Ma; Smellie *et al.* 1984), a wide range of ages is possible for the dykes. Although probably mainly Cretaceous, the only dyke dated so far in central Livingston Island yielded an Eocene age (56 Ma: Grikurov *et al.* 1970). Field relationships indicate an Eocene age (*c.* 40–45 Ma) for dykes adjacent to and cutting the Barnard Point tonalite (Smellie 1983). Some dykes may be very young, possibly related to the late Cenozoic opening of Bransfield Strait, although the evidence is equivocal and needs to be substantiated (Santanach *et al.* 1992).

Quaternary volcanic rocks also crop out on north-eastern Livingston Island (Smellie *et al.* 1984, Smellie 1990). By contrast with lavas in the older volcanic outcrops, the Quaternary lavas are very fresh and have unusual compositions with alkaline and tholeiitic characteristics (Smellie 1990).

Table I. Lithostratigraphy of volcanic and sedimentary sequences in central Livingston Island.

Group	Formation	Member	Age	Thickness	Principal lithological characteristics
	Inott Point Formation		<1Ma	>100 m (up to 300 m?)	Several small basaltic tuff cone relicts; rocks essentially fresh
Antarctic Peninsula Volcanic Group	Mount Bowles Formation		Cretaceous?	>600 m	Pyroclastic and lahar deposits, basalt-andesite lavas, thin arkosic sandstones and associated hypabyssal intrusions; highly altered
	Moores Peak breccias*		unknown (see text)	>200 m	Thick, massive sedimentary breccias dominated by MBF detritus
Trinity Peninsula Group	Miers Bluff Formation	Napier Peak	late Palaeozoic-early Mesozoic	<1300 m	Even, continuous, thin beds of fine sandstone and mudstone (T_{ace}, T_{ce} turbidites); lenticular conglomerates
		Johnsons Dock		*c.* 1700 m	Massive medium-coarse sandstones (S_3, T_a turbidites); thin-bedded sandstones and mudstones (T_{ace}, T_{ac}, T_{ae} turbidites); slump beds

*Stratigraphical status (formation/member) and affinities unknown.

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Miers Bluff Formation

On Hurd Peninsula, extensive outcrops of turbiditic sandstone, mudstone, conglomerate and sedimentary breccia have been assigned to the Miers Bluff Formation (originally Miers Bluff Series; Hobbs 1968). Three, previously undefined, lithostratigraphical divisions were described by Arche *et al.* (1992a, b) and Pallàs *et al.* (1992, Fig. 2). The lower two divisions (named here as the Johnsons Dock and Napier Peak members) are unambiguously part of the MBF, whereas the correlation of the upper division (Moores Peak breccias) is ambiguous. It may be part of the MBF or represent a younger, unrelated unit.

A type section for the MBF is defined on the south side of Johnsons Dock and was figured and described by Arche *et al.* (1992b). By contrast, Doktor *et al.* (1994) restricted the definition of the MBF to include only the lowermost strata on Hurd Peninsula, equivalent to our Johnsons Dock Member, which they further subdivided into three members. However, they incorrectly quoted Pallàs *et al.* (1992) as saying that units higher in the Hurd Peninsula sequence "comprise facies unknown in the TPG" and hence excluded the higher units from the MBF. The statement by Pallàs *et al.* (1992) referred only to the uppermost, lithologically distinctive unit, described in this paper as the Moores Peak breccias. Doktor *et al.* (1994) acknowledge that they did not examine the upper units nor much of the ground in eastern Hurd Peninsula. Thus, their basis for rejecting much of the outcrop as part of the MBF is not based on personal knowledge and may be suspect. Moreover, the three members described by Doktor *et al.* (1994) are lithologically inhomogeneous (which they acknowledge) and we question the reality of correlating comparatively thin rock units, which may be laterally discontinuous, across the large, central ice cap (Doktor *et al.* 1994, fig. 3). We also observed the lithological variations used by Doktor *et al.* (1994) to define their three members. However, similar lithofacies also crop out in eastern Hurd Peninsula. Further subdivision may yield important clues to the origin of the MBF, but we suggest that a simpler stratigraphical subdivision is all that can be justified by the present level of exposure and knowledge. Thus, we see no reason to subdivide the strata forming the west and south coasts of Hurd Peninsula, and they are grouped together in this paper within our Johnsons Dock Member.

Field relationships

The MBF is generally overturned and dips to the north-west. Thus, the bottom of the section is concealed beneath South Bay and the top is towards False Bay (Fig. 1). It is probably overlain unconformably by (?) Cretaceous volcanic sequences (Hobbs 1968, Smellie *et al.* 1984). Although the outcrop is continuous around the coast, exposures in central Hurd Peninsula are largely obscured by ice. Nevertheless, a simplified, composite stratigraphical section can be

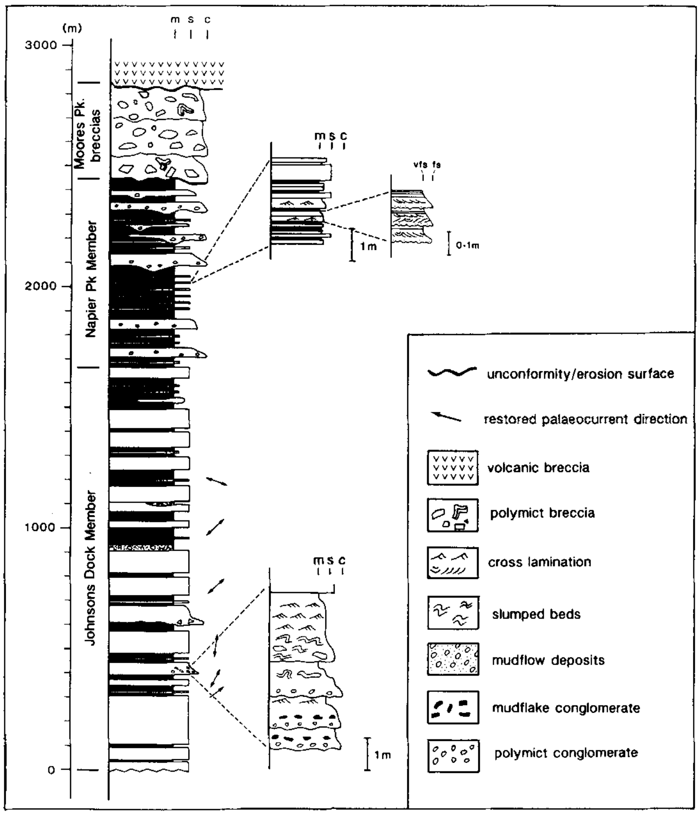

Fig. 2. Composite sedimentary section through the Miers Bluff Formation (Johnsons Dock and Napier Peak members) and Moores Peak breccias, based on sections A-A' and B-B' (Figs 1 & 3). Modified after Pallàs *et al.* (1992). Abbreviations: m - mudstone; s- sandstone; c -conglomerate; vfs - very fine sandstone; fs - fine sandstone.

constructed by projecting information from the southern outcrops (around Miers Bluff) into the central part of Hurd Peninsula (Figs 2 & 3). Interpretation of this section relies on there being no structural complications duplicating or cutting out major sections of strata. There are no major internal unconformities and the sequence has an approximate total thickness of about 3000 m (Hobbs 1968, Arche *et al.* 1992a, b, Muñoz *et al.* 1992). Trace fossils are locally abundant and support a marine origin for the sequence (Hobbs 1968, Doktor *et al.* 1994). Poorly preserved plant fragments, reworked coaly detritus and a single shell fragment are also present but are stratigraphically undiagnostic, although generally favouring an age younger than Carboniferous and not precluding a Mesozoic age (Hobbs 1968, Schopf 1973). Imprecise Rb-Sr isotopic ages of 204 ± 19 and 221 ± 34 Ma for MBF shales were interpreted as representing minimum ages for metamorphism (Pankhurst 1983, Hervé 1992). A more precise Rb-Sr errorchron age of 243 ± 8 Ma was interpreted as dating diagenesis in the early Triassic (Willan *et al.* 1994). MBF sandstones also contain detrital euhedral zircons with an age close to 322 Ma,

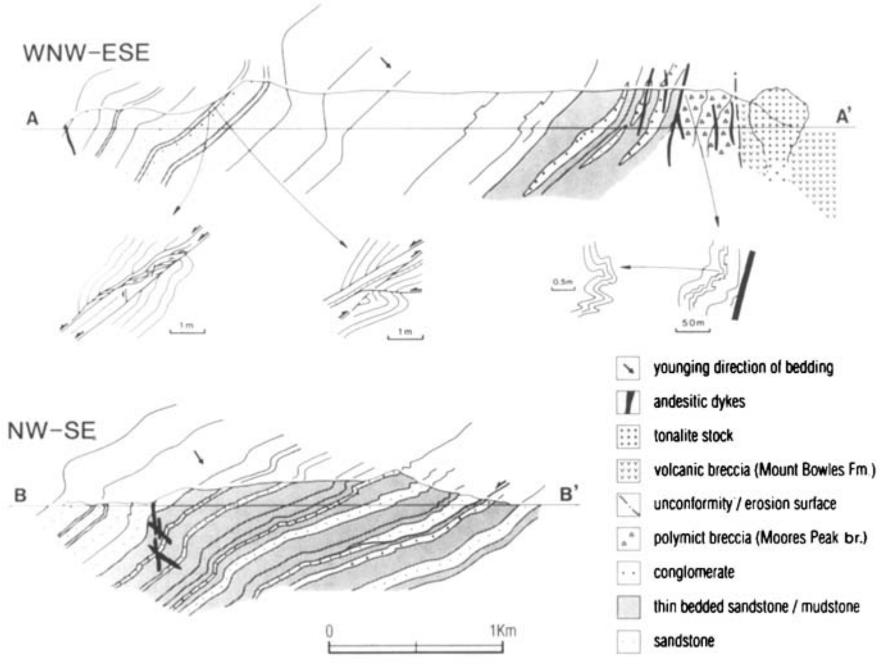

Fig. 3. Schematic geological cross sections through Hurd Peninsula illustrating the principal lithological and structural relationships. No phase 2 folds are intersected by these cross sections. The main folds shown represent phase 3a second order folds. They are located on the short and overturned limb of a first order fold several km in wavelength (also phase 3). The location of these sections is shown in Fig. 1. Modified after Muñoz *et al.* (1992).

interpreted as a provenance age and hence a maximum age of sedimentation (Loske *et al.* 1988).

Lithological description

The lowest lithostratigraphical unit on Hurd Peninsula (*Johnsons Dock Member*) consists of interbedded thick sandstones and mudstone (Fig. 2; approximate sequence boundaries shown in Pallàs *et al.* 1992, fig. 2). It crops out in the west and south of Hurd Peninsula and has an exposed thickness of 1700 m. Medium to coarse-grained sandstones predominate and are arranged in amalgamated beds often several metres thick (cf. Doktor *et al.* 1994, figs 5 & 10). The sandstones are generally poorly sorted and structureless, lending a massive appearance to many outcrops. Uncommon, erosive-based, matrix-supported conglomerates are also present, up to a few metres thick and containing abundant rounded to angular clasts of quartzose sandstone and mudstone. They show rare grading, comprising basal reverse grading followed by normal grading above. The normal-graded parts begin in pebbly sandstone and pass up through coarse mudflake conglomerate or breccia into cross-laminated sandstone. The basal strata in the Johnsons Dock Member are laterally continuous and are succeeded by coarse and medium-grained sandstones, similar to those stratigraphically below, alternating with thin-bedded sandstones and mudstones. The thin-bedded sections consist of fine-grained sandstone-mudstone beds up to a decimetre in thickness, which generally display incomplete Bouma sequences (T_{ace}, T_{ac}, T_{ae}). Abundant sole structures indicate that the finer beds were deposited by NE–SW-orientated palaeocurrents (calculated after structural restoration). Slumped strata are common, in places resulting in detached fold hinges.

The *Napier Peak Member* is confined to eastern Hurd Peninsula (Pallàs *et al.* 1992, fig. 2). It consists mainly of even, continuous, thin beds of fine-grained sandstone and mudstone. Lenticular beds of conglomerate and coarse sandstone are also present and seem to increase in thickness and number up through the section (Fig. 2). The fine sandstone beds are sharp-based (erosive) and sole marks are common. They are normally graded and cross laminated T_{ace} and T_{ce} turbidites. Some fine sandstones form locally amalgamated thin T_{acd} and T_{ac} beds a few cms thick. Dispersed, small, pebbly clasts are present in the sandstone and mudstone beds. The lenticular conglomerates in the upper parts of the sequence are up to 1 m thick, normally graded and have channelled bases. They contain well rounded, cm-size clasts of sandstone, quartz and granitoid.

Sandstones throughout the MBF are poorly sorted arkosic arenites and (less common) wackes formed of roughly equal proportions of quartz and feldspar (plagioclase and micro- to cryptoperthitic microcline), together with chloritized biotite, muscovite, minor lithic fragments, and accessory sphene, clinozoisite, garnet and zircon (cf. Smellie 1991, Smellie *et al.* 1984, Arche *et al.* 1992a, Doktor *et al.* 1994). Plagioclase is more abundant than microcline. The quartz includes monocrystalline and polycrystalline varieties. All the clasts are angular-subangular. The sparse matrix is clay-rich but locally consists of feldspar, silica and/or sericite. Calcite forms a minor secondary cement or fracture-fill.

Contact metamorphic recrystallization is restricted to areas around intrusions near Johnsons Dock and overlooking False Bay. North of Johnsons Dock, the strata are bleached and cordierite is prominent in the mudstones (Hobbs 1968). Sandstone matrices may also show coarser clay minerals (e.g. muscovite after sericite), chlorite and green-brown biotite.

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Chlorite, clinozoisite and quartz occur in veins. Apart from these local contact metamorphic effects and hydrothermal propylitic alteration (silicification, chloritization, epidotization) near volcanic rocks in eastern Hurd Peninsula, the metamorphic grade of the MBF is very low and corresponds mainly to incipient (anchizone) metamorphism (Arche *et al.* 1992a).

Structure

The structure of the MBF is dominated by open folding (Dalziel 1984, Muñoz *et al.* 1992, Doktor *et al.* 1994). Strata are predominantly overturned and dip at *c.* 45° to the NW or WNW (Fig. 3). The folding is polyphase (Fig. 4a) but cleavage is generally absent, similar to the Trinity Peninsula Group (Aitkenhead 1975), and is only crudely developed in mudstones, in some fold hinges. Most of the folds are open and of a large scale (tens to hundreds of metres in wavelength), whereas mesoscopic folds and kink bands are locally developed, particularly in thin-bedded sandstone-mudstone strata. Four fold systems are identified. They can be grouped into three folding phases, which are summarized here (after Muñoz *et al.* 1992). Two of the systems (2 and 3a) are present at most localities on Hurd Peninsula, whereas two (1 and 3b) are geographically restricted.

Axes of phase 1 folds plunge west (Fig. 4b) and axial surfaces dip steeply to the south. They are only developed in southernmost Hurd Peninsula. Axial surfaces of mesoscopic and larger-scale folds of phase 2 strike NW–SE and dip NE, whereas fold axes plunge NNW (Fig. 4a & b). Phase 3 folds, which represent the "main phase" of folding, consist of two almost coaxial open fold systems with subhorizontal fold axes which trend NE–SW (3a) or N–S (3b). These folds face SE, and their axial planes also dip SE (3a) or E (3b) (Fig. 4a & b). Phase 3 folds generally only produce variations in bedding dip and do not affect strike directions. Variations in the strike of bedding are mainly the result of phase 2 folds (Fig. 4a). Small scale contractional and detachment faults are associated with fold phases 2 and 3 (see details of these small scale structures in Fig. 3).

According to Dalziel (1972), overturning of the MBF is an effect of a tight fold several km in size, facing ESE and with its axial plane dipping WNW. This interpretation conflicts with the virtual absence of cleavage, the open geometry of all the observed folds and the fact that no WNW-dipping axial planes have been observed. According to our data the only folds that may have contributed to the overturning of bedding and resulted in the present regional dip are those of phase 3a. However, the open geometry of these SE-facing folds and the absence of cleavage are inconsistent with the predominance of *overturned* bedding. Accordingly, we tentatively suggest that the *folded* MBF sequence has been tilted to the SE, resulting in the overturning of bedding and the present attitude of the SE-dipping axial surfaces. Because the orientation of the tilting axis is not known, we present it as coaxial with the fold axes of phase 3a in Fig. 4c, because this solution involves the minimum amount of rotation.

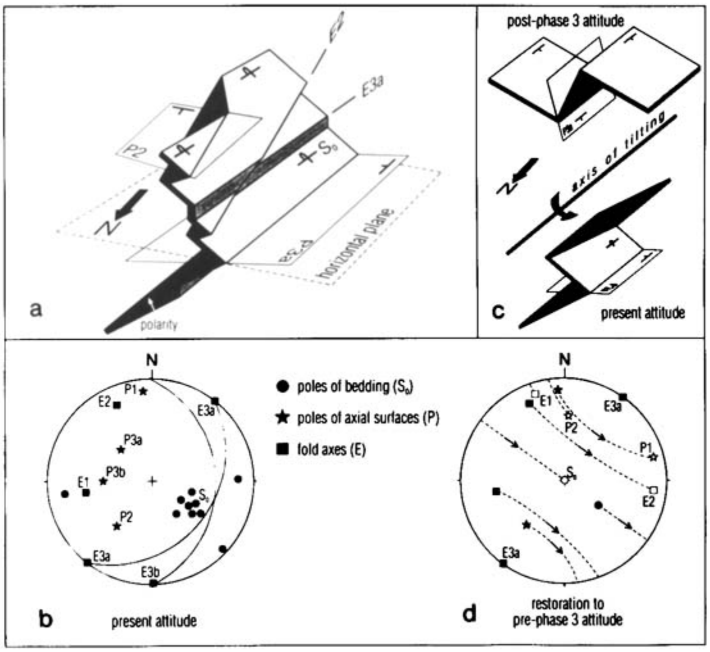

Fig. 4. Structure of the Miers Bluff Formation. **a.** Sketch illustrating the present-day relationships between phase 2 and 3a folds (the folds most widespread on Hurd Peninsula). Note that north points towards the viewer because this perspective most clearly demonstrates all of the involved elements. The small arrow labelled "polarity" shows the way-up direction and the structure is a first order syncline of phase 3a. Most of the strata exposed on Hurd Peninsula form part of the upper, overturned limb only. For clarity, the sizes of the phase 2 folds shown have been enhanced with respect to phase 3 folds. **b.** Lower hemisphere stereoplot showing the present attitude of bedding, axial surfaces and fold axes of phases 1, 2 and 3. The individual points shown represent areas of maximum density of corresponding data for each structure; several poles to bedding are included to show variations in the present-day regional attitude of bedding. **c.** Sketch illustrating a possible original attitude of first order phase 3 folds (several km in wavelength) and the effect of the postulated subsequent tilting south-easterly, which affects the whole of Hurd Peninsula. Note that only the short, overturned limb of this megastructure is present in Hurd Peninsula. **d.** Lower hemisphere stereoplot showing the procedure used to restore phase 1 and 2 structures to their possible pre-phase 3 attitudes. The restoration uses an axis of rotation parallel to phase 3a fold axes, in order to return the overturned bedding to a horizontal upright position. This rotation removes the effects of phase 3a folding and tilting.

Interpretation

Each of the lithostratigraphical units defined probably represents a major depositional episode (or growth *stage*;

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Mutti & Normark 1987), although both members also contain several minor cycles of coarsening-thickening and fining-thinning sequences (Arche *et al.* 1992b, Doktor *et al.* 1994) representing smaller scale fluctuations in sediment supply and local basin dynamics (cf. *sub-stages* of Mutti & Normark 1987). The Johnsons Dock Member is dominated by thick, coarse-grained, structureless or poorly graded beds, which resemble deposits of high-density turbidity currents (Lowe 1982). By contrast, the thinner-bedded sandstone-mudstone lithofacies in this member, and most of the Napier Peak Member, are composed largely of finer-grained, internally better-structured beds, which are classical turbidites and can be described in terms of the Bouma sequence (see also Arche *et al.* 1992b).

The three-dimensional morphology of the Johnsons Dock and Napier Peak members, and of the depositional basin are unknown. Moreover, the vertical sedimentological changes are ambiguous and do not easily compare with models of either aggrading or prograding sequences. An interpretation of the sedimentary environment using fan terminology (e.g. Walker 1984, Arche *et al.* 1992b, Doktor *et al.* 1994) is applied here, but we acknowledge that other depositional models may also be applicable (e.g. Shanmugam *et al.* 1985, Macdonald 1986, Mutti & Normark 1987). The Johnsons Dock Member contains a predominance of amalgamated, massive, coarse sandstones towards the base and thick beds of coarse sandstone interbedded with finer sandstones-mudstones above, suggestive of a very crudely fining-upward succession. It resembles an upper mid-fan sequence, with the massive sandstones and channel-based breccias (lower part of the member) deposited in a series of suprafan lobes and the better-bedded, upper part of the member possibly corresponding to lobe-fringe and/or inter-channel (or overbank) sequences. The Napier Peak Member, dominated by thin, fine-grained sandstone-mudstone turbidites, is characterized by even, continuous bedding probably reflecting unconstrained sediment flow over a smooth palaeotopography. They probably represent lower mid fan sedimentation, whilst the interbedded lenticular conglomerates may be either crevasse-splay deposits (non-erosive beds low in the sequence) or deposits of migrating, shallow channels (channelized beds higher up). Although exposure is poor, there is an apparent upward increase in abundance of conglomerates, consistent with the possible proximity of a major channel distributory system towards the end of the depositional period.

It is likely that the polyphase folding of the MBF was completed before the deposition of the Late Jurassic–Cretaceous Byers Group (Smellie *et al.* 1980, Crame *et al.* 1993) because these rocks and other sequences of Cretaceous age elsewhere on Livingston Island are not as strongly deformed. This interpretation assumes that the MBF and the Jurassic–Cretaceous sequences have not been juxtaposed by later faulting. The field relationships with the Byers Group are unexposed but the disposition of outcrops on Hurd Peninsula suggests that an unconformity exists between the MBF and (?)Cretaceous volcanic sequences (Mount Bowles Formation, see later), consistent with the sparse age evidence. The tilting episode postulated must be younger than the phase 3 folds but its age and origin are poorly constrained. It could be seen either as a continuation of deformation related to phase 3 folding or even as an event younger than the Mount Bowles Formation. In the first case, the present attitude of MBF strata could be seen as an extreme product of back-tilting, such as occurs at the rear of some accretionary complexes and is associated with underplating or back-thrusting (Seely 1977, Dalziel 1984). The second possibility is consistent with the presence of steep bedding in the Mount Bowles Formation, but is impossible to prove because of the lack of adequate exposure. Finally, tilting in the MBF could also be associated with the dip slip (reverse or normal) component of a regional fault system.

To restore the orientation of structures older than phase 3 folds requires a rotation about a horizontal axis of *c.* 140°. This action rotates the present-day, regionally overturned bedding to a horizontal right way-up attitude. As a result, phase 1 fold axes become orientated NNW–SSE, and the folds slightly east-facing, whereas phase 2 fold axes would be orientated E–W and the folds would face north (Fig. 4d).

In summary, the structural evolution of the MBF is tentatively explained as follows:

1) Local generation of phase 1 folds, facing east and with axial surfaces striking NNW–SSE. A synsedimentary origin for these early folds cannot be ruled out.

2) Generation of phase 2 folds, facing north and with subhorizontal E–W orientated axes.

3) Main phase folding, which deforms the folds of phases 1 and 2. Phase 3a deformation consisted of large-scale open folds, facing SE with steeply-dipping short limbs and NNE–SSW-trending subhorizontal axes. Note that phase 3a folds observed in the field are mesoscopic and correspond to second order folds parasitical on the short and overturned limb of a much larger (but still *open*) fold structure (Fig. 4a & c). The N–S-orientated phase 3b folds have only a local distribution.

4) Folding was completed before the Cretaceous period. However, additional tilting to the ESE is postulated in order to overturn bedding and achieve the present regional bedding attitudes. It could have taken place as an extension of the main phase of folding, or much later.

Moores Peak breccias

Field relationships

A lithologically distinctive sequence of unfossiliferous sedimentary breccias crops out at Moores Peak and in scattered, small exposures to the north, in eastern Hurd Peninsula (Fig. 1). Field relationships with both the MBF and volcanic

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breccias of the Mount Bowles Formation are unclear because of intervening ice, tectonic disruption and widespread cross-cutting intrusions with attendant alteration. The breccias have a minimum thickness of about 200 m (estimated on the widest, continous exposure *c.* 400 m NNW of Moores Peak). The total thickness is unknown as the sequence continues north-westwards beneath the central ice cap on Hurd Peninsula. It is bounded to the south-east by a complicated transition zone (described later), which separates undoubted Mooores Peak breccias from the Mount Bowles Formation. The age of the Moores Peak breccias is unknown but it must lie between that of the Napier Peak Member of the MBF and the Mount Bowles Formation.

Lithological characteristics

The deposits consist almost entirely of massive breccia. Bedding is generally ill-defined and most outcrops appear predominantly massive, but erosive bed surfaces are sometimes preserved. In most outcrops, the coarse clasts are 10–90 cm across and are dominated by angular, tabular mudstone and sandstone fragments generally similar to MBF lithologies (Fig. 5). Porphyritic lava clasts are also rarely present and there are conspicuous blocks of deformed sandstone-mudstone strata, some of which contain a weak cleavage. The clasts are predominantly supported in a silty-sandy, epidote-rich matrix but parts of the sequence are clast-supported. Some metre-thick conglomeratic beds, which are tentatively included in this rock group, are graded from conglomerate with well rounded clasts (1–2 cm in diameter) mainly of quartz and granitoid, passing up through breccio-conglomerate dominated by coarse, angular, sandstone-mudstone clasts, into coarse sandstone at the top.

Interpretation

Although the characteristics of the Moores Peak breccias are sufficiently distinctive to designate it as a separate lithostratigraphical unit, the poor and scattered nature of most of the exposures make an unambiguous sedimentological interpretation difficult at present. However, the thick, sandy, channel-based and amalgamated breccias broadly resemble resedimented conglomerates described by Walker (1975). The disorganized, ungraded nature of the breccias suggests deposition from debris flows, although the absence of muddy matrices suggests high-concentration, cohesionless flows (Postma 1986). By contrast, the less common graded beds suggest direct suspension sedimentation of gravel from high density turbidity currents (Lowe 1982) or possibly low-stage flow conditions in a braided stream (Allen 1981).

The origin and affinities of the Moores Peak breccias are enigmatic and cannot yet be resolved. They could be part of the MBF, a basal unit in the Mount Bowles Formation, or a deposit entirely unrelated to either the MBF or the Mount Bowles Formation and deposited during an intervening

Fig. 5. Heterolithic sedimentary breccia of the Moores Peak breccias at Moores Peak. Note the very variable angularity of the dark (mudstone) cobbles. Much of the pale rock is formed of angular cobbles and pebbles of sandstone. The ice-axe head is 30 cm across.

period. The coarse clast population is overwhelmingly dominated by MBF-derived fragments, suggestive of an origin as intraformational breccias derived by a combination of contemporaneous reworking (e.g. mudflake conglomerates) and/or slumping of the MBF sequence. Superficially similar units interpreted as slump deposits occur within the lower part of the MBF (Arche *et al.* 1992b, Doktor *et al.* 1994). In this interpretation, the breccias could have formed during a major channel migration into the mid-fan area previously occupied by the Napier Peak Member. Thus, the breccias would represent a major influx of very coarse detritus, possibly derived from the degradation and collapse of a tectonically active slope (e.g. sediments associated with some accretionary prism complexes: e.g. Doubleday *et al.* 1993).

However, most of the deposits generally do not resemble classical mudflake conglomerates and it is unclear whether the slump deposits in the MBF are precisely comparable in lithology and origin with the heterolithic breccia deposits of the Moores Peak breccias: the former contain pervasively sheared, mud-rich matrices and large, chaotic slabs of deformed mudstone-sandstone beds whereas the latter have essentially undeformed, silty-sandy matrices and are dominated by sandstone cobbles and boulders. Moreover, the angularity of many of the sandstone clasts, presence of stratified blocks and deformed, cleaved mudstone clasts suggests a high degree of lithification in the clasts prior to deposition, and the epidote-rich matrix suggests a greater influence from a volcanic provenance than is characteristic of the MBF (cf. Smellie 1991, Arche *et al.* 1992a). In addition, the abundant hydrothermal alteration evident throughout the Moores Peak breccias has more in common with that observed in the Mount Bowles Formation than the MBF. These characteristics suggest an age significantly younger than the MBF.

Furthermore, there are stratigraphical and lithological similarities with basal conglomerates of the Early Jurassic Botany Bay Group in northern Antarctic Peninsula (Farquharson 1984, Rees 1993). Like the breccias at Moores Peak, the Botany Bay Group deposits were derived largely from Trinity Peninsula Group sandstones and mudstones and they crop out in a similar structural position, between the Trinity Peninsula Group and the APVG. The conglomerates were interpreted by Farquharson (1984) as non-marine deposits of debris flows and braided streams on alluvial fans. Most of the conglomerates in the Botany Bay Group are clast-supported, forming flat, sharply defined, generally non-erosive beds with angular–rounded clasts up to 1.7 m in diameter. However, matrix-supported and/or graded conglomerates with erosive basal surfaces are also present in the Botany Bay Group and resemble parts of the Moores Peak breccias, although a closer sedimentological comparison is not yet possible. Thus, the age and affinities of the Moores Peak breccias are ambiguous at present, although the weight of the evidence probably favours an age younger than the MBF.

Mount Bowles Formation

Field relationships

Outcrops in the central part of the study area consist mainly of altered volcanic sequences collectively assigned here to the Mount Bowles Formation. They resemble volcanic sequences mapped elsewhere on Livingston and Greenwich islands, which are Cretaceous in age (Smellie *et al.* 1984). However, the central Livingston Island rocks have generally proved unsuitable for geochronological analysis, and their eruptive age is uncertain but assumed to be Cretaceous by analogy. They are correlated here with the Antarctic Peninsula Volcanic Group (APVG; cf. Thomson & Pankhurst 1983). The total thickness of the formation is unknown but a minimum thickness of 650 m is inferred (assuming no major tectonic complications) from homoclinal sections exposed in cliffs overlooking False Bay.

The very thick bedding and its poor definition, because of pervasive alteration and jointing, are characteristics of all the volcanic outcrops and preclude the logging of a type section. The descriptions in this paper are based mainly on sections exposed at the unnamed nunatak 1 km west of Willan Nunatak and on the south side of the Mount Bowles massif. Volcaniclastic rocks are predominant in outcrops south of Burdick Peak, whereas lavas form much of the Mount Bowles massif. Contact relationships with the MBF are obscure except at Burdick Peak, where vertical MBF beds are overlain by a sequence of (?) near-horizontal altered andesite lavas, suggesting a possible angular unconformity. The sinuous nature of the contact traced north-eastwards from Moores Peak towards Mount Bowles is also consistent with an unconformity rather than a structural contact (Fig. 1). The contact between the Moores Peak breccias and volcanic breccias of the Mount Bowles Formation is placed at the western end of a complicated "transition zone" about 200 m wide on the north side of Moores Peak. The contact is steeply dipping, strikes NNE–SSW and is not deformed by the polyphase folding observed in the Miers Bluff Formation. There are structural dislocations within the transition zone, and interpretation is complicated by alteration and numerous dykes. Within the transition zone, the rocks consist of breccias with MBF and volcanic clasts, together with numerous andesitic to microtonalitic hypabyssal intrusions. The proportion of MBF clasts in the breccias decreases eastwards. Because of the presence of abundant volcanic clasts, we interpret the transition zone sequence as part of the Mount Bowles Formation. Intrusive breccia also rarely occurs within the Moores Peak breccias.

Lithological characteristics

The outcrops south of Burdick Peak are dominated by massive, dark-coloured volcaniclastic rocks, mainly coarse lapilli-tuffs and breccias with outsize clasts up to 35 cm across. They are interbedded with a few lavas and green volcaniclastic sandstones and intruded by dykes and sills. All the rocks are pervasively altered and jointed. Bedding surfaces are rarely preserved but variable bedding attitudes are evident judged from changes in the gross lithology. Beds are typically several metres thick (up to 20 m), but may be only a few dm thick in the sandstone interbeds. Exposures around the Mount Bowles massif contain similar lithologies but, by contrast, are dominated by vesicular, andesitic lavas with rare volcaniclastic interbeds.

The clastic rocks around Mount Bowles consist of tuff and lapillistone beds *c.* 10–30 cm thick, some showing possible normal and reverse grading. Hobbs (1968) and Smellie *et al.* (1984) described fragments of MBF sandstone in lapilli-tuffs from Mount Bowles and near Willan Nunatak, but in general, the clastic sequences have not been examined closely. The more accessible clastic rocks south of Burdick Peak are very poorly sorted, heteromict lapilli-tuffs, lapillistones and volcanic breccias predominantly andesitic in composition (Fig. 6). The clasts are matrix-supported and angular. Feldspar is present as phenoclasts. Accidental clasts consist of fragments of MBF sandstones, siltstones, mudstones and plutonic rocks. They are ubiquitous but generally form only a small proportion (*c.* 5%) of most beds, except in the "transition zone" sequence, in which they may be abundant (> 50%). Despite the advanced alteration, the coarse tuff and lapilli-size juvenile clasts are relatively well preserved texturally. Most are blocky and non-vesicular but a few are highly vesicular and cuspate; the original relative proportion of vesicular and non-vesicular fragments is unknown. In most rocks, the tuff-rich groundmass is completely recrystallized and the original textures destroyed. However, rare vitroclastic textures remain, consisting of dispersed cuspate glass shards. There are also uncommon, palecoloured tuffs (e.g. at Moores Peak), which are monomict and formed entirely of cuspate shards and highly vesicular glass (Fig. 7). Some of these beds show poorly defined fiammé-like textures.

The rare sandstone interbeds are mainly poorly sorted, matrix-poor, fine sandstones (arkosic arenites and wackes) and siltstones. Plagioclase forms up to 75% of the clast population, and is associated with quartz and minor lithic fragments (polycrystalline quartz and micaceous phyllite). Clast shapes vary from angular to subrounded.

The lavas, dykes and sills are predominantly andesites, basalts and dolerites with varied textures, many of them similar to textures present as fragments in the pyroclastic rocks. They are mainly porphyritic or glomeroporphyritic, and trachytic textures are common. Phenocrysts include plagioclase (andesine-oligoclase), augite and opaque oxide. Possible altered olivine occurs rarely, and apatite microphenocrysts were observed within lava clasts in the pyroclastic rocks.

Alteration is widespread and pervasive, particularly in the volcaniclastic rocks where primary groundmass textures are largely destroyed. Plagioclase may be little-altered but it is generally replaced mimetically by albite, or by some combination of sericite and epidote (including clinozoisite and zoisite) and magnetite; actinolite, chlorite or biotite may occur along cleavage planes. Pyroxene is rimmed or replaced by amphibole (actinolite and hornblende; electron microprobe analyses also suggest cummingtonite or possibly gedrite), chlorite and minor carbonate. The groundmass is extensively replaced by chlorite, biotite, epidote, actinolite, sulphide and oxide minerals, rutile and sphene (leucoxene), and irregular "pools" of quartz are common. Biotite is restricted to outcrops south of Burdick Peak. Amygdales and veins of all of these secondary minerals are common at all localities.

The alteration mineral parageneses, generally consist of some combination of albite-actinolite-chlorite-epidote-sphene-quartz ± biotite ± sulphide ± oxide; the highest grade parageneses encountered include hornblende-cummingtonite-biotite ± chlorite ± quartz ± oxide. Cummingtonite seems only to occur in outcrops close to the Barnard Point tonalite pluton. Alteration textures in the volcanic rocks are decussate and mimetic, with no signs of a foliation or schistosity, consistent with recrystallization under conditions of no stress. We suggest that the alteration assemblages and textures in the volcanic rocks were caused predominantly by contact metamorphism (albite-epidote-hornfels facies) and hydrothermal solutions associated with either subvolcanic intrusions (?Cretaceous) or later plutons (Eocene) (Smellie *et al.* 1984, Willan 1992).

Structure

Because of alteration, the volcanic outcrops appear massive. However, where identified, bedding is homoclinal within individual exposures, with orientations varying widely between exposures: the Mount Bowles outcrops dip homoclinally E and SE at *c.* 30–40°, whereas at Burdick Peak beds dip W at 70°, and the strata west of Willan Nunatak dip SW at 55°. There is no cleavage and mesoscopic folding has not been observed. Structural relationships between exposures are also unknown.

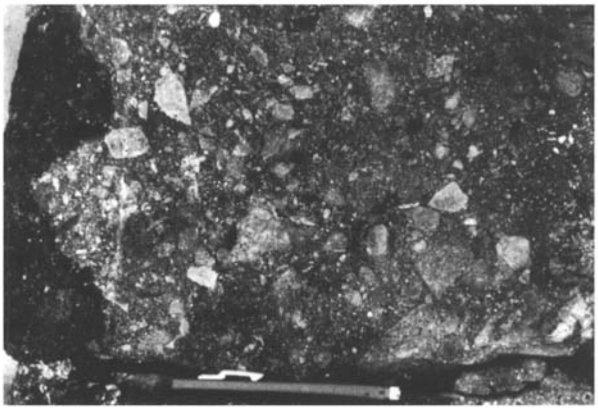

Fig. 6. Coarse volcanic breccia of the Mount Bowles Formation at the nunatak 1 km west of Willan Nunatak. Note the abundance of polymict, angular lithic clasts and fine, dark, volcanic matrix. The pen is 12 cm long.

Interpretation

There is no evidence (e.g. fossils, pillow lavas, hyaloclastite, turbidites) to suggest that the Mount Bowles Formation was deposited in a predominantly subaqueous setting. The origin of the dominant clastic lithofacies is hard to interpret and the evidence is ambiguous. The thick, massive beds of fines-rich,

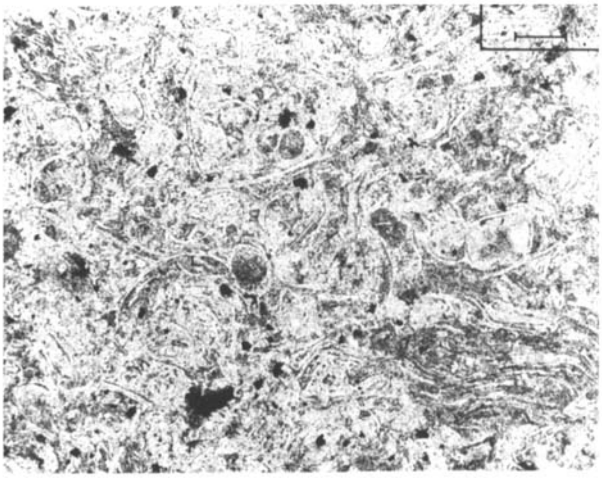

Fig. 7. Photomicrograph of a pale vitric tuff at Moores Peak showing well-preserved vitroclastic texture dominated by bubble-wall shards and compressed pumice. Scale bar is 1.28 mm long.

polymict debris most resemble debris flow (lahar) deposits. Vitroclastic textures are preserved in the abundant fine tuff matrix and were probably ubiquitous prior to alteration, consistent with a high volcanic input. The presence of abundant, angular, polymict (mainly lava) clasts, and juvenile fragments with blocky, poorly vesicular shapes, are characteristics of tephra from phreatomagmatic eruptions, and coeval volcanism is clearly indicated by the interbedded lavas. The very high proportion of accessory and accidental clasts is also a characteristic feature of eruptions from maars, although maar sequences are characteristically thin and have a poor preservation potential (Lorenz 1986). Moreover, phreatomagmatic deposits are generally well stratified, with thinner beds than observed on Livingston Island, although the original fine bedding details may have been masked by the pervasive alteration and jointing. The monomict vitric tuffs with highly vesicular vitroclasts, bubble-wall shards and rare fiammé-like textures are probably airfall products of drier, magmatic eruptions.

The interbedded arkosic sandstones lack sedimentary structures. The transport and depositional mechanisms, and depositional environment are unknown. They contain a clast population comparable to that of MBF sandstones. However, the MBF sandstones have a lower proportion of plagioclase, contain detrital potassium feldspar and have a more variable accessory mineral population than the Mount Bowles Formation arkoses. With their high feldspar content, the Mount Bowles Formation arkoses are classed as immature sandstones. Moreover, the similarity in the size of quartz and feldspar grains suggests that the sediment did not undergo prolonged transport prior to deposition. The arkosic sandstones were, therefore, probably derived from a provenance area composed of uplifted, exposed MBF (or possibly Moores Peak breccias) and volcanic strata.

Some of the bedding orientations in the Mount Bowles Formation may reflect original, variable bedding attitudes, as in overlapping vent complexes (e.g. Smellie *et al.* 1980, fig. 4). However, bedding inclinations are locally much steeper than expected in an undisturbed volcanic succession. Thus, the sequences could also be affected by folding and/or faulting, similar to Cretaceous sequences in western Livingston Island (Byers Peninsula and Cape Shirreff), and/or by forceful emplacement of plutonic or hypabyssal intrusions. Deformation in the volcanic sequence differs in orientation and structural style from that observed in the MBF and is characterized by open folds with subvertical axial planes (Smellie *et al.* 1980); fold axes trend NW–SE at Cape Shirreff (observations by the authors).

Inott Point Formation

Field relationships

Fresh volcanic rocks were described previously from only four localities (Hobbs 1968, Smellie *et al.* 1984, Smellie 1990). Several additional outcrops were discovered during our investigations and the total outcrop area is now known to be considerably larger than formerly thought (Figs 1 & 8). The two small outcrops of volcanic rocks near Burdick Peak are geographically isolated from the larger volcanic outcrop around Sharp Peak. They do not satisfy the criterion of original physical continuity of formations (Whittaker *et al.* 1991). However, they are indistinguishable in lithology, age and origin from the larger outcrop and they are provisionally included in the Inott Point Formation in this paper. One of these small outcrops was previously attributed to Gleaner Heights (Smellie *et al.* 1984) but is here correctly repositioned 1.5 km north-west of Burdick Peak (Fig. 1). The excellent cliff exposures at Inott Point are defined as the type locality. By contrast with volcanic rocks of the (?) Cretaceous Mount Bowles Formation, all of the Inott Point Formation outcrops are essentially fresh, texturally unmodified and undeformed. Although no basal contact is exposed, an unconformable relationship with the Mount Bowles Formation is inferred from the presence of rare fragments of highly altered volcanic rocks in the Inott Point Formation. K-Ar isotopic dating at four, widely separated localities has yielded Pleistocene–Recent ages ($\leq$ 1 Ma: Smellie *et al.* 1984, and unpublished information of the authors).

Smellie *et al.* (1984) postulated a correlation between the Late Cenozoic volcanic outcrops on Livingston and Greenwich islands and a major, NE-trending fault in South Bay. In addition, a Late Cenozoic outcrop discovered during our investigations, situated 1.5 km SE of Burdick Peak, also lies approximately on the trace of a NW-SE-orientated structure identified from satellite imagery (Santanach *et al.* 1992; Fig. 1).

Lithological characteristics

The outcrops in the Edinburgh Hill area consist of isolated dolerite plugs, plugs encased in basaltic pyroclastic rocks, and basalt lavas associated with black and red scoria. The pyroclastic deposits mainly consist of basaltic tuff-breccia, lapilli-tuff and tuff.

Stratified lapilli-tuff and tuff units are monomict, yellow- or buff-coloured deposits composed of fine to coarse tuff with dispersed (10–15 %) buff and dark grey fine to coarse lapilli, and a small proportion of grey lava blocks up to *c.* 15 cm in diameter. At Inott Point, sub-rounded pebbles, cobbles and boulders of diorite and strongly altered APVG lavas and volcaniclastic rocks, up to 25 cm in diameter, are a minor (*c.* 1–2 %) but conspicuous component. Small-displacement (few dm) faults are common at Inott Point and also occur in the sequence west of Sharp Peak (locality P.1822; Fig. 8). Three distinctive types of deposits are distinguished by their stratification:

1) The commonest type consists of alternating lapilli-rich and lapilli-poor, discontinuous layers a few cm to dm

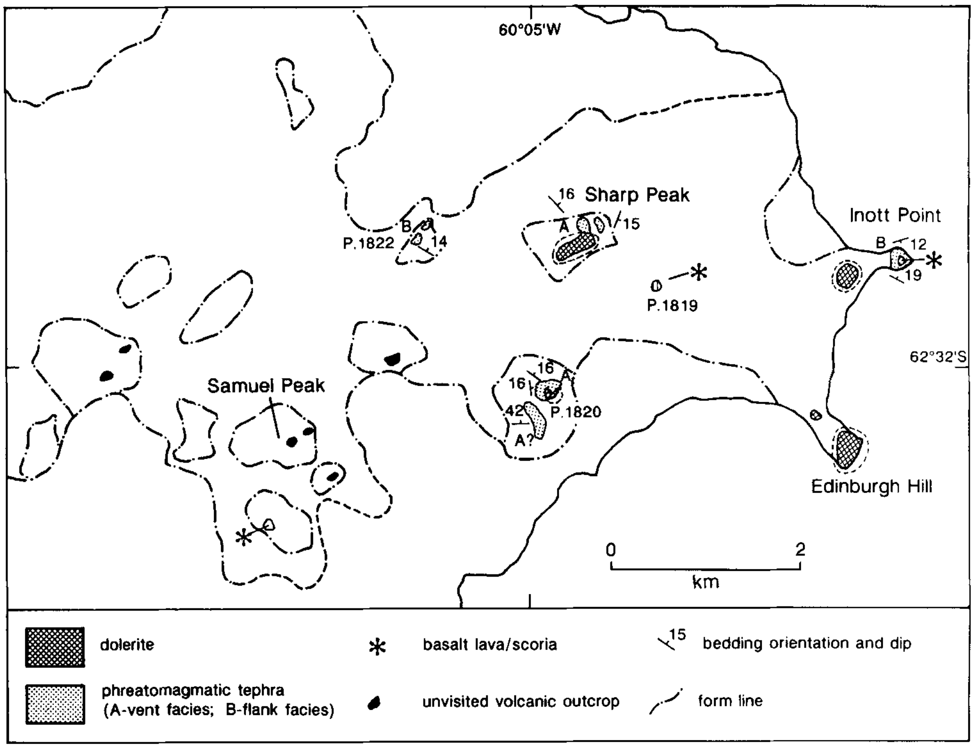

Fig. 8. Geological sketch map of north-eastern Livingston Island, showing the principal outcrop of the Inott Point Formation. Form lines and coastline based on satellite image (Institut Cartogràfic de Catalunya 1992).

thick, with indistinct to rarely distinct bedding planes (Fig. 9). The lapilli form discontinuous, often grain-supported trails or more laterally persistent, diffuse bands of lapilli. These deposits form thick, monotonous successions, which are particularly well exposed at Inott Point. Sag structures are present beneath some outsize clasts at that locality (Fig. 9), but outsize clasts in most outcrops lack sag structures. Lee-side, coarse-clast concentrations may be present in depressions associated with sag structures (block impact craters; Fig. 9).

2) Some poorly sorted, crudely stratified lapilli-tuff forms undulating, wavy bedforms, *c.* 30–40 cm thick and up to 1.5 m long, with gently dipping stoss and lee sides. The wave crests are smoothly rounded, convex-upward and have poorly developed brinkpoints. These bedforms are very uncommon and were observed only in the outcrops SSW of Sharp Peak (locality P.1820).
3) Lapilli-tuff also forms discrete beds 0.15–1 m thick, with sharp, planar bedding surfaces. The thickest development of these beds is a 25 m-thick sequence in the outcrop west of Sharp Peak (locality P.1822). The individual beds consist largely of poorly sorted lapillistone, which is either massive or normally graded and may be faintly planar laminated to top (Fig. 10). The lapillistone passes up into about a dm of wavy-planar laminated, fine to coarse tuff with dispersed lapilli and rare blocks up to 12 cm in diameter. Truncated laminations occur rarely.

The lapilli-tuffs and tuffs are petrographically similar. They are very poorly sorted, rich in fine tuff-size detritus and dominated by vitric (juvenile) clasts. The latter are typically poorly or non-vesicular and blocky in shape, with minor to pervasive palagonite alteration and minor zeolite cement. They are associated with a small proportion (*c.* 10–20 %) of tachylite and accessory basalt lava clasts. Armoured lapilli are rarely present.

Massive tuff-breccia and lapillistone are extensively developed at Sharp Peak, the nunatak 1.6 km SSW of Sharp Peak (P.1820) and at the summit of the nunatak west of Sharp

Fig. 9. Prominent bomb sag in stratified lapilli-tuffs and tuffs, caused by impact of rounded diorite (?) beach cobble. Note also the concentration of coarse clasts (indicated by arrow) developed within the sag, on the lee side of up-current lip of the structure. Pyroclastic current (surge) travelled left to right. Inott Point. The hammer shaft is about 60 cm in length.

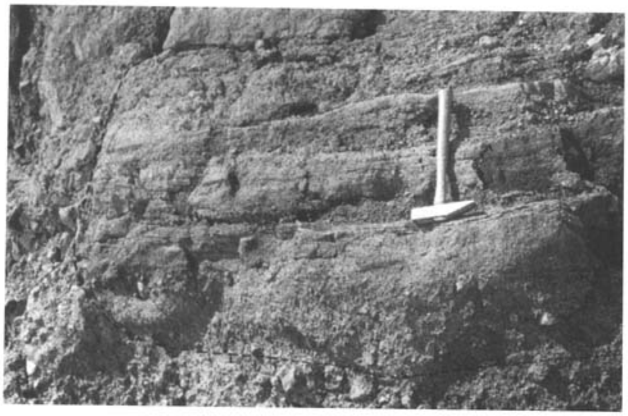

Fig. 10. Massive-graded lapillistone and lapilli-tuff beds, each passing up into wavy-laminated tuff. The stratified beds form a large clast (outlined) within massive tuff-breccia. In the picture, the clast grades down and to the left into massive tuff-breccia. Nunatak 1.5 km south of Sharp Peak. The hammer shaft is 40 cm long.

Peak (P.1822). Together, they form unsorted, mainly matrix-supported deposits generally many metres thick (up to 12 m at P.1822), sometimes bed-like and with erosive bases, but more often with diffuse, very irregular shapes. Relicts and or megaclasts of stratified lapilli-tuff and tuff dm to m-thick are common (Fig. 10). A "ghost"-like, relict stratification is rarely discernible within some otherwise massive deposits. Contacts with the stratified rocks may be sharp or diffuse. A distinctive type of sharp contact consists of curviplanar (convex-down), sometimes splayed surfaces coated with yellow fine–medium tuff. The coarse clast population in the tuff-breccia consists largely of angular, blocky, poorly or non-vesicular accessory and juvenile lava fragments up to 40 cm in diameter, set in fine to coarse tuff or fine lapillistone. Clasts within the lapillistones are identical to those found in the stratified deposits.

Edinburgh Hill is a spectacular columnar-jointed dolerite plug *c.* 250 m in diameter and rising to *c.* 110 m. The description by Hobbs (1968) is generally accurate, although the "reddish flow lenses" in the upper parts, which he described, may simply be accentuated weathering in the coarser (inaccessible) centre of the plug exposed towards the top of the outcrop. Columnar jointed plugs are also present at Sharp Peak and near Inott Point. An irregular, locally vesicular, basaltic intrusion occurs in lapilli-tuffs south of Sharp Peak (P.1820), and a dolerite sill is present *in situ* and as abundant, near *in situ* platy clasts on the tombolo facing Edinburgh Hill.

Basalt lavas are present on the inaccessible summit of Inott Point, as isolated exposures south-east of Sharp Peak (P.1819) and north-west of Burdick Peak, and interbedded with coarse, red and black scoria beds south-east of Burdick and Samuel peaks. The lavas at the latter locality are clastogenic and occur as densely welded lenses up to 2.5 m thick.

Interpretation

The lavas and interbedded reddish- or black-coloured scoria deposits with welded lenses are relicts of cinder cones, the products of essentially "dry" eruptions. By contrast, the more abundant stratified pyroclastic rocks are tuff cone deposits produced by phreatomagmatic eruptions. This is indicated by the abundant vitric clasts and their vesicularity, morphology and pervasive palagonite alteration, the high proportion of tuff-size clasts, and bedding characteristics (Wohletz 1983, Sohn & Chough 1989). The presence of sag structures indicates that some clasts were emplaced ballistically into coeval deposits that were already relatively cohesive, whereas the more common *lack* of sags beneath other outsize clasts, together with lee-side coarse-clast concentrations in block-impact depressions, indicate that most beds were probably emplaced by lateral transport, probably within pyroclastic currents (base surges). The characteristics of the predominant stratified deposits ("type 1", above) suggest deposition from suspension and by traction sedimentation from a turbulent surge, which fluctuated in velocity and particle concentration. The crude, poorly sorted, thicker stratification suggests rather rapid fall-out from suspension, whereas the thinner stratification with lapilli trails and lenses suggests slower fall-out and grain segregation. "Type 2" beds showing low-angle dipping, undulatory bedforms are probably transitional between traction current (surge) deposits and deposits produced by relatively higher-concentration surges with high shear stress. The normally graded basal parts of the "type 3" stratified beds suggest suspension sedimentation from the "body" of a high-density, turbulent surge, whereas the laminated upper parts suggest deposition from the turbulent, low-concentration surge "tail", in which traction sedimentation was more important.

The field relationships of the massive tuff-breccias and associated lapillistones indicate two principal modes of occurrence: 1) As discrete, unsorted, matrix-supported beds with the appearance of debris-flow or proximal high-concentration surge deposits; and 2) ill-defined masses with stratified lapilli-tuff megaclasts and/or "ghost-like" relict stratification. Contacts of the latter with *in situ* stratified deposits are typically steeply transgressive (e.g. Fig. 10) and we suggest that the tuff-breccias and lapillistones were formed by the collapse and partial to complete disaggregation of large sections or "rafts" of stratified deposits, and/or possible pervasive fluidization and *in situ* disruption of bedding. The massive deposits occur only in outcrops containing a sub-central plug, and we suggest that they represent a vent facies association. This is also consistent with the presence of curviplanar, splayed surfaces interpreted here as within-vent slip planes. By contrast, the stratified deposits are interpreted as (?)outward-dipping flank sequences, which lack massive tuff-breccia/lapillistone except as rare interbeds (e.g. summit of P.1822).

Accidental clasts are absent from all but one outcrop (Inott

Point), indicating that the explosive eruptions were extremely shallow and unlikely to have quarried extensively into the underlying bedrock, unlike Surtseyan and maar eruptions (Moore 1985, Lorenz 1986). This conclusion may also apply to the Inott Point outcrop where accidental clasts show relatively good rounding consistent with an origin as beach material. It is perhaps no coincidence that the Inott Point outcrop is at present-day sea level, whereas the other outcrops of pyroclastic rocks are all above the limit for marine geomorphological features (*c.* 200 m above sea level; John & Sugden 1971). An origin by subglacial eruption is likely (but unproven) for all of the north-eastern outcrops (Smellie *et al.* 1984), but the contemporaneous ice would have been relatively thin (<*c.* 100 m) to yield the observed sequences (cf. Smellie *et al.* 1993).

Synthesis

The Antarctic Peninsula and South Shetland Islands form part of a magmatic arc, which was formerly continuous with southern South America. The magmatic and sedimentary sequences in the region and associated tectonic processes are related, ultimately, to the subduction of proto-Pacific oceanic crust beneath the Antarctic continental margin since late Palaeozoic times, at least (e.g. Storey & Garrett 1985).

The lithological characteristics and stratigraphical and structural relationships of the volcanic and sedimentary formations in Livingston Island closely resemble those of tectonostratigraphic units elsewhere in the northern Antarctic Peninsula. This implies that similar geological processes were involved in the evolution of the region. The oldest rocks in the central part of the island consist of the late Palaeozoic–early Mesozoic MBF, which forms the local basement. The depositional environment of the MBF comprised upper mid-fan suprafan lobes and lobe-fringe or interchannel sedimentation (Johnsons Dock Member), and a lower mid-fan setting with thin channel sequences (Napier Peak Member). Deformation of the MBF is dominated by three phases of *open* folding. Our interpretation of the style of the "main" (phase 3) folding episode requires subsequent tilting to create the large-scale overturning observed.

A thick sequence of thick-bedded mass flow deposits overlies the MBF (Moores Peak breccias). The breccias occupy a structural position similar to the Lower Jurassic Botany Bay Group in northern Antarctic Peninsula (i.e. may be unconformable on the MBF), but the Livingston Island deposits are of unknown stratigraphical age and are possibly an upper unit of the MBF. Like the Botany Bay Group, the principal provenance for the breccias was a sandstone-mudstone sequence (the MBF). The MBF clasts were relatively lithified and tectonically deformed prior to redeposition, implying uplift and exposure of part of the MBF during the period of breccia deposition.

The (?)Cretaceous Mount Bowles Formation largely consists of interbedded lavas and volcaniclastic rocks, which unconformably overlie the MBF. Volcanic detritus in the clastic rocks was probably produced mainly during phreatomagmatic eruptions, as a result of interaction of magma and groundwater. The coarse volcaniclastic lithofacies most resemble deposits of debris flows (lahars), but there is no evidence that the Mount Bowles Formation was formed in a predominantly subaqueous (marine or lake) setting. Passive lava effusion was prevalent around the Mount Bowles massif. Thin beds of arkosic sandstone are also present, particularly in outcrops closest to the MBF, indicating exposure of the MBF and/or Moores Peak breccias during the Cretaceous. MBF clasts are extremely rare in outcrops further away from the MBF outcrop (Mount Bowles massif), suggesting that the MBF palaeotopography may have been more subdued or obscured by volcanic rocks when the Mount Bowles sequences were erupted. The moderate dips of the volcanic strata reflect palaeotopography and/or a late Mesozoic/Cenozoic deformation, possibly analogous to that more clearly shown by Cretaceous volcanic and sedimentary formations in Byers Peninsula and Cape Shirreff.

The youngest rocks exposed on Livingston Island crop out most extensively in the north-east and patchily elsewhere, and are of Pleistocene–Recent age (Inott Point Formation). The north-eastern outcrop is the largest outcrop of Late Cenozoic volcanic rocks in the South Shetland Islands excluding Bransfield Strait. The exposures are mainly relicts of numerous small, probably overlapping basaltic tuff cones, but some outcrops consist solely of basalt lava and/or scoria erupted under essentially dry, subaerial conditions. The fault trending NE through South Bay and Edinburgh Hill is probably a Cenozoic strike-slip structure (Santanach *et al.* 1992). Rejuvenation as a dip-slip structure, by extension during arc-splitting and formation of the Bransfield Strait marginal basin, may explain the close geographical association between some faults and the young volcanism.

Acknowledgements

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