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Del Mesozoic al Present

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Raimon Pallàs i Serra	Francesc Sàbat	Joan Manuel Vilaplana

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

J.L. SMELLIE[1], M. LIESA[2], J.A. MUÑOZ[2], F. SÀBAT[2], R. PALLÀS[2] and R.C.R. WILLAN[1]

[1]*British Antarctic Survey, Natural Environment Research Council, High Cross, Madingley Road, Cambridge CB3 OET*
[2]*Universitat de Barcelona, Facultat de Geologia, 08028 Barcelona, Spain*

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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Key words: Livingston Island, lithostratigraphy, geological evolution, magmatic arc, Antarctica

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-

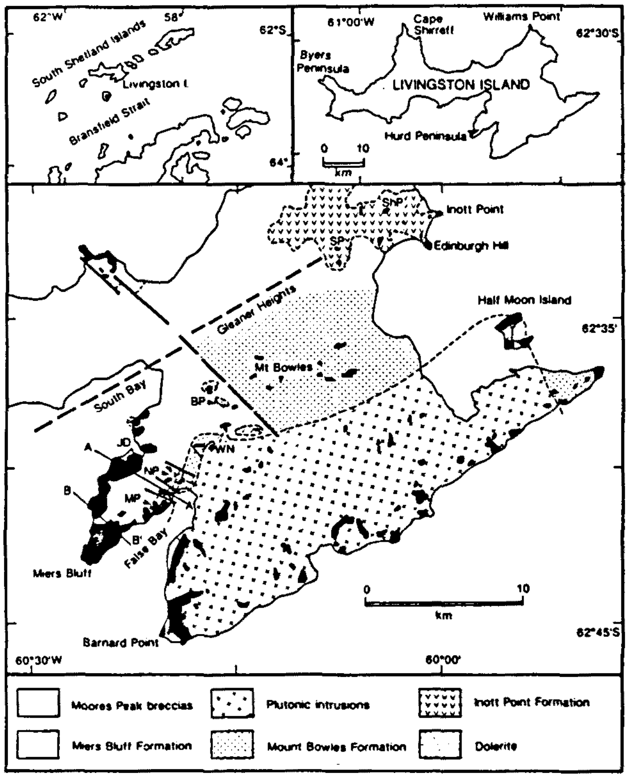

Fig. 1. Geological sketch map of eastern Livingston Island. The locations of composite schematic cross sections shown in Fig. 3 (A-A', B-B') are also indicated. Faults shown are inferred from geomorphological features and satellite images. Rock exposures shown in black. For clarity, some of the smaller 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.

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.

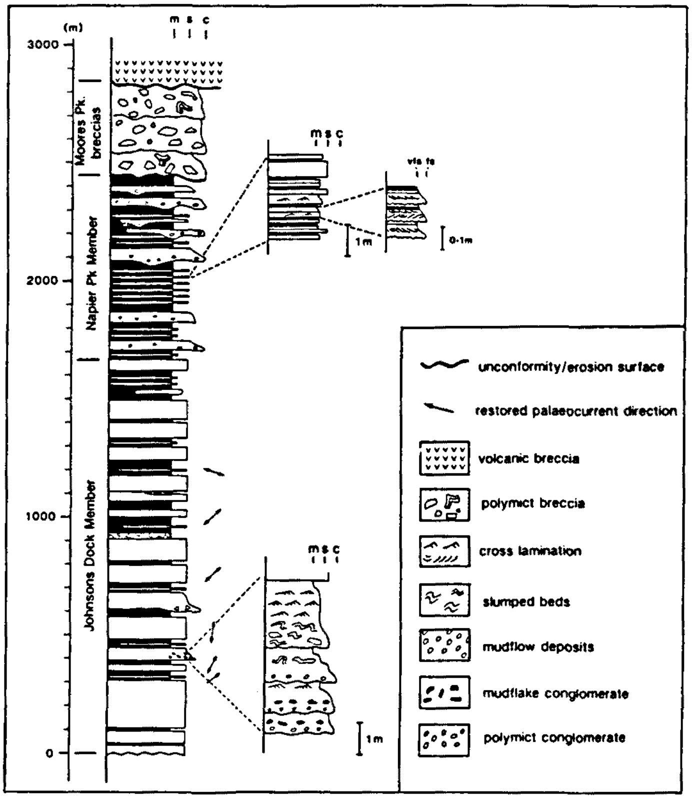

Fig. 2. Composite sedimentary section through the Miers Bluff Formation (Johnsons Dock and Napier Peak members) and Moores Peak breccias, based on sections A-A' and B-B' (Figs 1 & 3). Modified after Pallàs *et al.* (1992). Abbreviations: m - mudstone; s- sandstone; c -conglomerate; vfs - very fine sandstone;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

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 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,

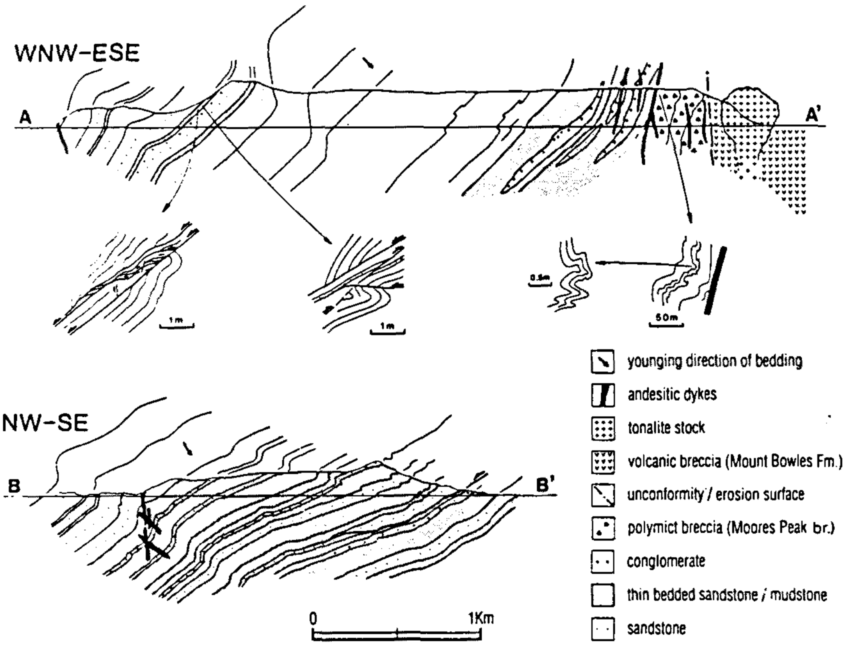

Fig. 3. Schematic geological cross sections through Hurd Peninsula illustrating the principal lithological and structural relationships. No phase 2 folds are intersected by these cross sections. The main folds shown represent phase 3a second order folds. They are located on the short and overturned limb 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.

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).

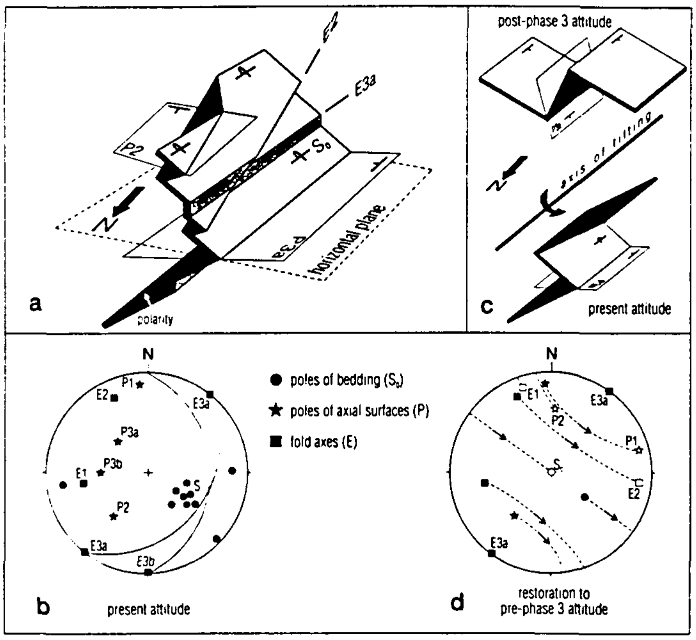

Fig. 4. Structure of the Miers Bluff Formation. **a.** Sketch illustrating the present-day relationships between phase 2 and 3a folds (the folds most widespread on Hurd Peninsula). Note that north points towards the viewer because this perspective most clearly demonstrates all of the involved elements. 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.

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.

Interpretation

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

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 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 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).

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.

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,

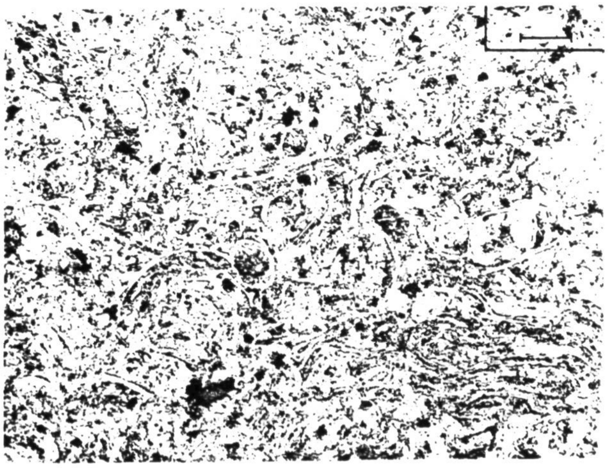

Fig. 7. Photomicrograph of a pale vitric tuff at Moores Peak showing well-preserved vitroclastic texture dominated by bubble-wall shards and compressed pumice. Scale bar is 1.28 mm long.

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 (≤ 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

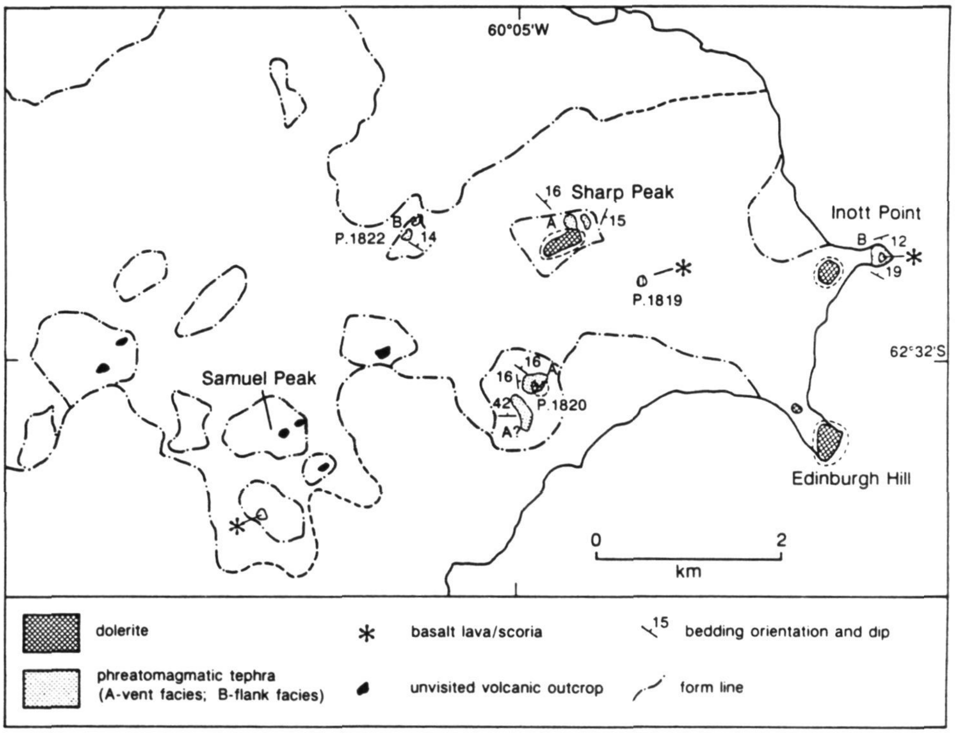

Fig. 8. Geological sketch map of north-eastern Livingston Island, showing the principal outcrop of the Inott Point Formation. Form lines and coastline based on satellite image (Institut Cartogràfic de Catalunya 1992).

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

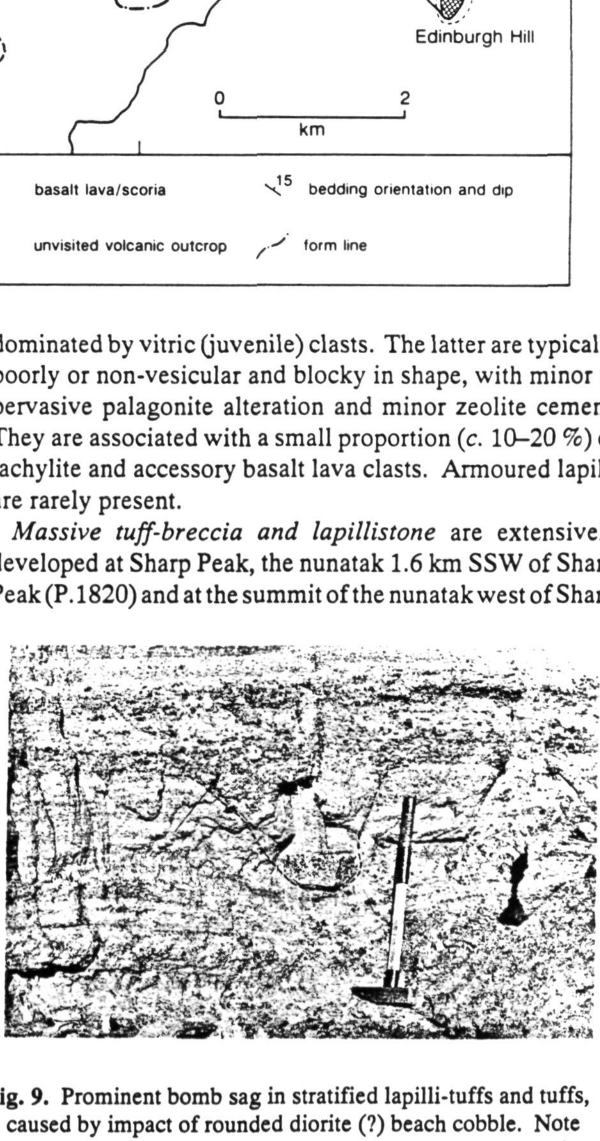

Fig. 9. Prominent bomb sag in stratified lapilli-tuffs and tuffs, caused by impact of rounded diorite (?) beach cobble. Note also the concentration of coarse clasts (indicated by arrow) developed within the sag, on the lee side of up-current lip of the structure. Pyroclastic current (surge) travelled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

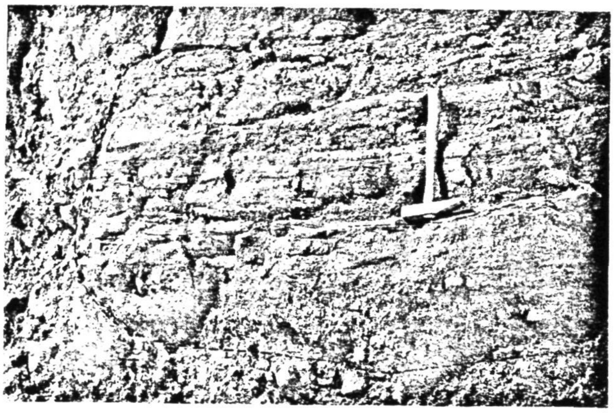

Fig. 10. Massive-graded lapillistone and lapilli-tuff beds, each passing up into wavy-laminated tuff. The stratified beds form a large clast (outlined) within massive tuff-breccia. In the picture, the clast grades down and to the left into massive tuff-breccia. Nunatak 1.5 km south of Sharp Peak. The 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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En aquest article es fa una descripció de la successió volcànica que aflora a la Punta Hannah. En aquesta localitat, s'hi troba el millor aflorament de materials del *Grup Volcànic de la Península Antàrtica* a l'Illa Livingston. Es fa una subdivisió de la successió en diversos trams. Es destaca la presència d'un contacte no concordant i la presència de falles normals. Els diferents trams són interpretats com el producte de diverses fases en l'activitat volcànica.

Estratigrafía de los materiales volcánicos de Punta Hannah (Isla Livingston, Shetland del Sur).

Francesc Sàbat (1), Zheng Xiangshen (1,2), Josep Mª Casas (1) y Raimon Pallàs (1)

1) Facultad de Geología, Universidad de Barcelona
2) Institute of Geology, Academia Sinica, Beijing, China

Resumen

Se ha realizado un corte geológico de Punta Hannah. Las rocas de esta localidad forman parte del Grupo Volcánico de la Península Antártica (GVPA) y constituyen una alternancia de rocas volcanoclásticas y lavas cuya edad es supuestamente cretácico superior. La sucesión está organizada en tres tramos bien diferenciados, cada uno de los cuales se sugiere que corresponde a un acontecimiento eruptivo distinto. El tramo inferior está constituido por lavas dacíticas y brechas volcanoclásticas. Una ligera discordancia separa el tramo inferior y el tramo intermedio. El tramo intermedio registra un acontecimiento volcánico completo; se inicia con rocas piroclásticas andesíticas, seguidas de lavas basáltico-andesíticas y brechas volcanoclásticas riolíticas. En el tercer tramo las lavas son más abundantes y tienen características petrológicas propias. A ambos lados de la discordancia que separa los tramos inferior e intermedio existe una fuerte zeolitización y la parte inferior del segundo tramo está atravesada por algunos diques y sills, sugiriendo que esta zona fue estructuralmente débil y fácilmente alterable. A lo largo de todo el corte, el buzamiento de las capas es regular y suave hacia el NNW, lo que indica que el área ha sido estable con posterioridad a los acontecimientos volcánicos. El interés del excelente afloramiento de Punta Hannah reside en que se puede observar el registro de un acontecimiento eruptivo completo (tramo intermedio) que ilustra el tipo de actividad volcánica mesozoica que originó las Shetland del Sur.

Palabras clave: Livingston, Cretácico, litostratigrafía, petrología, vulcanismo.

Abstract

Hannah Point cross section is built up by volcanic and volcaniclastic rocks belonging to Antarctic Peninsula Volcanic Group. Three cycles corresponding to different volcanic events can be distinguished in the volcanic sequence. The lower cycle is constituet by dacitic lavas and volcaniclastic breccias. An unconformity marks the boundary between the lower and middle cycles. The middle cycle records a whole volcanic event; it stars with andesitic pyroclastic rocks, followed by andesitic-basaltic lavas and ending with riolitic volcaniclastic breccias. The upper cycle is mainly constitued by fresh basaltic lavas. Zone around the above mentioned unconformity was a weak zone where strong zeolite alteration and dykes are present. Bedding attitude is regular and slightly dipping to NNW suggesting this area has been stable after the volcanic events. Hannah Point area yields a good example of the Cretaceous volcanic activity that built up the South Shetland Islands.

Keywords: Livingston, Cretaceous, lithostratigraphy, petrology, vulcanism.

Introducción

La isla Livingston es la segunda isla en extensión de las Shetland del Sur (fig. 1) y posee un gran interés de cara al conocimiento de la evolución geológica de la parte Norte de la Península Antártica por un doble motivo. En primer lugar es la única isla donde aflora la Formación Miers Bluff (FMB) que constituye el zócalo del arco volcánico de las Shetland del Sur, y por otra parte hay algunos afloramientos clave para estudiar la evolución del volcanismo asociado a la construcción de dicho arco volcánico (Smellie et al, 1984)

Las rocas volcánicas y volcanoclásticas pertenecientes al Grupo Volcánico de la Península Antártica (GVPA) afloran en distintas zonas de Livingston, pero su estudio únicamente se ha llevado a cabo en la Península Byers (Grikurov et al., 1970; Pankhurst et al., 1979; Smellie et al., 1984). Debido a que las rocas volcánicas de Byers son cretácicas en contraposición a las que afloran en la Isla del Rey Jorge que son Cenozoicas, Smellie et al. (1984) propusieron que, en las Shetland del Sur y durante la formación del GVPA, la actividad volcánica migró de SW a NE. Para discutir esta hipótesis es necesario realizar estudios detallados en otras localidades, situadas entre la Península Byers y la Isla del Rey Jorge. En cualquier caso, el excelente afloramiento de Punta Hannah ofrece una oportunidad inmejorable de caracterizar la actividad volcánica de las Shetland del Sur durante el Mesozoico.

Punta Hannah está situada en la parte central de la costa Sur de Livingston, en el extremo Este de Bahía Walkers, ofrece una sucesión continua de rocas que probablemente registra más de un acontecimiento eruptivo y puede constituir un buen ejemplo del vulcanismo que formó el GVPA.

Marco geológico

El zócalo de las islas Shetland del Sur está constituido por la FMB que aflora únicamente en la Península Hurd de Livingston (Hobbs, 1968; Smellie et al., 1984 y en prep.; Pallàs et al., 1992). Esta formación, que se considera Carbonífero - Triásica, está constituida por una alternancia de arenitas y lutitas correspondientes a un lóbulo turbidítico (Smellie, 1991; Arche et al., 1992) y está afectada por una deformación polifásica anterior al Cretácico (Dalziel, 1972; Muñoz et al., 1992).

Las Shetland del Sur están formadas mayoritariamente por rocas volcánicas del GVPA. Estas rocas pertenecen a la serie calcoalcalina, forman parte del arco magmático de la península Antártica y de las Shetland del Sur, y están asociadas a la subducción meso-cenozoica de la litosfera del proto - Pacífico bajo la Península Antártica (Storey y Garret, 1985). La mayor parte de la isla Livingston está constituida por rocas volcánicas cretácicas del GVPA. Hobbs (1968) dividió las rocas de la península Byers en una serie volcánica antigua y otra moderna, pero actualmente se considera que forman una serie continua a la que se denomina formación Byers (Pankhurst, 1982; Pankhurst y Smellie, 1983; Smellie et al., 1984). En la base de la serie volcánica de Byers hay intercalaciones de lutitas marinas con ammonites del Jurásico terminal y Cretácico inferior seguidas de intercalaciones de sedimentos continentales con restos de plantas del Cretácico inferior (González-Ferran et al., 1970; Hernandez y Azcarate, 1971; Covacevich, 1976, Smellie et al, 1984; Thomson, 1992). Las dataciones radiométricas (Smellie et al, 1984) confirman las asignaciones paleontológicas y además ponen de manifiesto la existencia de cuerpos de rocas volcánicas y subvolcánicas correspondientes al Cretácico superior. Las rocas de esta última edad también afloran en otros puntos de Livingston y son abundantes en las islas Greenwich y Robert donde han sido agrupadas en la formación Coppermine (Pankhurst y Smellie, 1983; Smellie et al., 1984). La isla del Rey Jorge también está constituida fundamentalmente por rocas volcánicas del GVPA, pero en este caso la mayoría de ellas son del Paleógeno (Birkenmajer et al., 1983; Smellie et al., 1984; Zheng y Liu, 1989).

La actividad plutónica en las Shetland del Sur es muy reducida. Los plutones intruyen las rocas del GVPA y se considera que forman parte del arco magmático de la península Antártica. En Livingston y sus alrededores hay gabros y tonalitas en la cordillera del Friesland y en la isla de Media Luna; y también existen algunas pequeñas apófisis tonalíticas en la Península Hurd (Livingston). La edad radiométrica de las tonalitas de Media Luna es 120 Ma (Creácico superior) y la de las tonalitas de Punta Barnard (Livingston) es 46 Ma (Eoceno medio) (Grikurov et al., 1970; Dalziel et al., 1973; Smellie et al., 1984)

Numerosos diques basáltico-andesíticos cortan las rocas de la FMB y del GVPA (Caminos et al., 1973; Willan, en prep.). Las relaciones de campo indican que los diques han intruido durante un periodo muy prolongado que abarca buena parte del Cenozoico; por el momento solo se dispone de la datación radiométrica de un dique cercano a la ensenada Johnsons que ha dado una edad de 56 Ma (Paleoceno superior) (Grikurov et al., 1970).

Las rocas volcánicas más modernas de Livingston son basaltos olivínicos frescos que afloran en la parte NE y central de la isla. Estas rocas tienen una edad radiométrica de menos de 1.5 Ma (Smellie et al., 1984 y en prep.), son similares a las que existen en algunos puntos de Greenwich y son geoquimicamente diferenciables de los basaltos de las islas Decepción y Pinguin; de todas formas se considera que ambos tipos están asociados a la extensión responsable de la formación del mar de Bransfield (Smellie et al., 1984 y en prep.).

Descripción de la sección de punta Hannah

La simplicidad de la estructura y la calidad del afloramiento del corte de Punta Hannah hacen que éste sea un lugar ideal para levantar una sucesión estratigráfica detallada que permita estudiar el resultado de la actividad volcánica, así como distintos estadios de su evolución.

Hobbs (1968) describe el afloramiento de Punta Hannah y presenta un corte parcial. Divide la sucesión de rocas volcánicas en tres partes: la parte basal (42.7 m) considera que esta compuesta de andesitas masivas interestratificadas con aglomerados verdes, lavas amigdalares y tobas; la parte central (110 m) está constituida principalmente por aglomerados friables verdosos y azulados, cenizas y lavas amigdalares marrones; la parte superior (195 m) está formada por lavas andesíticas, tanto masivas como amigdalares.

El corte presentado (fig. 2) tiene unos 1800 m de longitud en dirección NNW-SSE y se extiende desde el nivel del mar hasta el punto más elevado de punta Hannah (cota 195). El espesor de la serie volcánica expuesta es de unos 470 m; el buzamiento es relativamente regular y suave (15 - 25°) hacia el NNW; así pues los niveles inferiores están situados en el extremo Sur del corte y los superiores en el extremo Norte. Basándonos en las relaciones de campo y en las descripciones petrológicas es posible dividir la sucesión volcánica en tres tramos.

Primer Tramo.- Aflora en el sector meridional del corte y tiene un espesor visible de unos 120 m. La capa 1 (fig. 2) tiene unos 13 m de espesor y está formada por lava dacítico-andesítica masiva; muy compacta y dura; tiene textura porfírica hialopilítica. Un 3% de su volumen está ocupado por microfenocristales (0.1-0.3 mm) de clinopiroxeno (Cpx) cuya forma varía entre euhedral y subeuhedral; en general, el Cpx es fresco pero ocasionalmente está alterado a anfíbol. Algunos microfenocristales pseudomorfos de plagioclasa (Pl) están rellenos de clorita foliada. La matriz está compuesta de microlitos de Pl (30%), Cpx granular de forma irregular (30%) y cuarzo (10%).

El resto de capas del tramo inferior están constituidas por rocas piroclásticas, principalmente brechas volcanoclásticas y lapillis. En particular la capa 3 es de lapilli más o menos endurecido, compuesto por fragmentos de roca volcánica (55%) y cristales englobados por ceniza; los fragmentos son irregulares, subangulares o subredondeados, con texturas porfirítica, pilotaxítica y hialopilítica segúnlos casos; los cristales son principalmente fragmentos de fenocristales de Pl, euhedrales o subhedrales, alterados; además de ceniza la matriz posee fragmentos de vidrio. En la parte intermedia de la capa 3 hay un nivel de unos 7 cm de composición riolítica con evidencias de una estructura de flujo; este nivel está zeolitizado y está afectado por pequeñas fracturas rellenas de zeolita foliada y, a veces, de calcita. Inmediatamente encima hay un nivel de ceniza discontinuo de pocos centímetros de espesor, afanítico y constituido por vidrio y cuarzo.

Segundo tramo.- *Está formado por una alternacia de tobas, brechas* volcanoclásticas y lavas andesíticas, y puede dividirse en tres subtramos: A, B y C, que totalizan en conjunto unos 200 m de espesor.

El subtramo A tiene un espesor de unos 60 m, está formado por una alternancia de rocas piroclásticas duras y blandas y culmina en una superficie claramente erosiva que lo separa del subtramo B. Las capas duras (6, 8 y 10) son de brechas volcanoclásticas soldadas; los fragmentos de vidrio volcánico constituyen una parte importante de la roca (de 20 a 50%); los fragmentos de rocas volcánicas tienen forma irregular y son subangulares o redondeados; la matriz tiene textura porfírica con fenocristales de Pl y Cpx englobados en vidrio y ceniza. Las capas blandas (7 y 9) tienen una constitución parecida a las capas duras pero poseen un contenido inferior de vidrio.

El subtramo B tiene un espesor de unos 75 m y se inicia con una capa (11) de tobas laminadas en cuya parte inferior localmente se halla un nivel de aglomerados recubriendo las depresiones del techo del subtramo A. Las tobas laminadas contienen ceniza volcánica y ocasionalmente presentan estructuras de flujo; las vacuolas son muy abundantes y están rellenas de clorita. Este subtramo se caracteriza por contener dos capas de lava que se adelgazan hacia el Sur, de forma que su espesor en la parte Norte es 20 m y en la parte Sur 10 m. Estas lavas son compactas, duras y tienen un color gris obscuro; su textura es pilotáxica o microlítica. Poseen algunos fenocristales de Pl y Cpx; los de Pl son euhedrales y están alterados a clorita, mientras que los de Cpx son hipidiomorfos y más frescos; la matriz está compuesta principalmente de Pl acompañada de Cpx. Estas lavas poseen algunas amígdalas rellenas de zeolita y clorita. La lava superior (capa 14) presenta una disyunción columnar muy marcada en su mitad inferior que es reemplazada por una disyunción planar en la parte superior.

El subtramo C tiene un espesor de unos 65 m y está constituido exclusivamente de rocas piroclásticas. Se inicia con un nivel de aglomerado de 2.5 m situado en la base de la capa 15. El resto de esta capa está constituido por brechas volcanoclásticas. Las brechas volcanoclàsticas de la capa 16 están soldadas y son de color gris verdoso. La capa 17 también está constituida de brechas volcanoclásticas y presenta una estratificación evidente. La roca de la parte inferior de cada uno de los lechos es de color negro grisáceo; más del 80% de la roca son fragmentos de rocas volcánicas tales como basaltos, andesitas y dacitas; los cristales de Pl, y también de Cpx; representan el 10% del volumen total; la matriz es fundamentalmente vidrio con estructura fluidal. La roca de la parte superior de los lechos es brecha volcanoclástica soldada de color marrón; los fragmentos de vidrio son el elemento predominante (40%) y poseen un tamaño comprendido entre 2 y 20 mm, tienen forma irregular, son de color verde pálido e incluyen cristales de Pl y Cp; más del 20% de la roca está formada por fragmentos de roca volcánica de forma irregular y subangulares que están distribuidos caóticamente, los fragmentos más grandes tienen 6x3 y 4x1.3 cm; el 5% de la roca son cristales de Pl y Cpx; la matriz de la roca está compuesta de ceniza y pequeños fragmentos de roca y de vidrio. La capa superior del subtramo C (capa 18) está formada por brechas volcanoclásticas soldadas y masivas; la roca tiene un color marrón o gris oscuro y el 60% de su volumen son fragmentos de vidrio con cristales de Pl y Cpx alterados a clorita; estos fragmentos tienen varios cm de tamaño, son subangulosos, irregulares y de color verde oscuro; la matriz está constituida principalmente por vidrio con estructura fluidal que contiene numerosas pequeñas amigdalas rellenas de zeolita y clorita; la matriz también contiene algunos pequeños fragmentos de roca (5%) y cristales de Pl y Cpx (5%).

Tercer tramo.- Tiene 150 m de espesor y está constituido por brechas volcanoclásticas y lavas, no obstante, debido a las condiciones de afloramiento tan solo se ha podido muestrear la capa superior que corresponde a una lava basáltica masiva. Esta capa tiene casi 80 m de espesor y descansa sobre lava con disyunción columnar. El basalto es gris-negro, compacto y duro; los fenocristales representan el 3% del volumen total; la mayoría son microfenocristales de Pl y unos pocos de Cpx; ocasionalmente los microfenocristales presentan textura glomeroporfírica; la matriz está compuesta de microlitos euhedrales y aciculares de Pl, y cristales subhedrales de Cpx.

Rocas subvolcánicas.- Las rocas intrusivas no son abundantes en Punta Hannah, tan solo algunos diques y sills delgados intruyen las capas inferiores del segundo tramo. La roca de estas intrusiones menores es dura y compacta, su textura es porfirico-dolerítica, tiene composición basáltica y es de color gris obscuro. Los fenocristales son de Pl y Cpx; los de Pl son euhedrales, están alterados a clorita y constituyen el 30% de la roca. Los fenocristales de Cpx son generalmente frescos, de color verde obscuro o negro, tienen formas irregulares, presentan textura glomerocristalina, poseen un tamaño máximo de 1 mm y forman el 5% de la roca. La matriz está constituida principalmente por microlitos euhedrales de Pl, conjuntamente con minerales opacos y Cpx.

Estructura.- Las rocas volcánicas y volcanoclásticas del corte de Punta Hannah casi no están deformadas y, tal como ya se ha apuntado anteriormente, buzan suavemente (15 - 25°) y regularmente hacia el NNW. Las únicas estructuras de deformación observadas son unas pocas fallas normales y un pequeño pliegue de ámbito local y relacionado con la intrusión de un sill. Las fallas normales son poco numerosas pero evidentes, producen desplazamientos desde inapreciables hasta unos 10 m, tienen una dirección ENE-WSW (tienen la misma dirección que las capas) y se agrupan en dos familias, unas que buzan al SSE y otras hacia el NNW. A pesar de que el número de observaciones es pequeño, estas sugieren que en un primer estadio las fallas han actuado como extensionales casi puras y que en un segundo estadio han funcionado como direccionales. Durante al menos el primero de estos movimientos se ha producido la cristalización de fibras de laumontita, las cuales tapizan las superficies de falla y rellenan pequeñas fracturas paralelas a ellas. En la zona estudiada y en sus alrededores inmediatos no aflora ningún cuerpo de rocas plutónicas.

Evolución de la actividad volcánica

Las rocas que componen cada uno de los tramos tienen una disposición particular y características petrológicas ligeramente distintas, por lo que se sugiere que cada tramo podría corresponder a un acontecimiento volcánico distinto.

Las capas del primer tramo buzan unos pocos grados (entre 5 y 12) hacia el NNW, las del segundo tramo tienen un buzamiento más acentuado (alrededor de 15°, aunque muy excepcionalmente pueden llegar a 30°) y, finalmente, las del tercer tramo tienen nuevamente un buzamiento suave. La zona de cambio de buzamiento entre el primer y el segundo tramo es bien visible y corresponde a una ligera discordancia angular; el segundo tramo se depositó sobre una superficie erosiva que afecta al primer tramo, el cual previamente había sido basculado ligeramente. Esta erosión que afectó al primer ciclo con anterioridad a la deposición del segundo sugiere que cada uno de los tramos podría corresponder a un acontecimiento volcánico distinto. Las relaciones geométricas entre el segundo y el tercer tramo no son visibles ya que la zona de contacto está recubierta por derrubios y hielo.

Por otra parte cada tramo presenta características petrológicas propias. El primer tramo está constituido por lavas y rocas volcanoclásticas de composición riolítico-dacítica; es de resaltar la existencia de un nivel lenticular y delgado de composición riolítica (intercalado en la capa 3). Algunos de los niveles de este ciclo (especialmente el de composición riolítica) están muy zeolitizados.

El segundo tramo está compuesto por rocas volcanoclásticas y lavas basáltico-andesíticas. El subtramo A está formado por brechas volcanoclásticas andesíticas que a veces están soldadas; el espesor de las tobas soldadas aumenta desde la base hacia la parte superior del subtramo, conjuntamente con la proporción de fragmentos de vidrio; las rocas de este subtramo están atravesadas por diques de rocas subvolcánicas y han sido muy zeolitizadas. El principal componente del subtramo B son dos coladas masivas y gruesas de lava andesítica con textura microlítica, de todas formas el tramo se inicia con una capa delgada de aglomerado y contiene varios niveles de brechas volcanoclásticas. Las coladas de lava se adelgazan y el contenido en amígdalas aumenta hacia el SSE sugiriendo que el sentido del flujo fue de Norte a Sur. Las rocas del subtramo B también están ligeramente alteradas y el vidrio volcánico y algunos fenocristales de plagioclasa están parcialmente reemplazados por clorita. La gran abundancia de fragmentos de vidrio con estructuras de flujo sugiere que la roca del subtramo C es más ácida; su abundancia aumenta desde la base hacia el techo. Las rocas de este subtramo están poco alteradas, aunque pueden estar débilmente cloritizadas.

El tercer tramo está constituido de brechas volcanoclásticas, y coladas de lava potentes y masivas. Estas lavas son basálticas y son más frescas que las lavas de los tramos anteriores.

Discusión y conclusiones

El corte geológico de Punta Hannah está constituido por rocas volcánicas y volcanoclásticas probablemente del Cretácico superior. Se considera que forman parte del Grupo Volcánico de la Península Antártica, el cual tiene naturaleza calcoalcalina. Los trabajos en curso permitirán conocer en detalle las características químicas de las rocas de Punta Hannah y su edad radiométrica, al mismo tiempo que aportarán datos complementarios para argumentar el significado de cada uno de los tres tramos descritos. Provisionalmente sugerimos que cada tramo corresponde a un episodio de actividad volcánica distinto.

Probablemente el primer tramo registra solo el último estadio de un primer episodio volcánico. El segundo tramo podria registrar un episodio volcánico relativamente completo. Este episodio se hubieses iniciado con una erupción no excesivamente explosiva (subtramo A), seguida de la efusión de lavas andesítico-basálticas (subtramo B) y hubiese terminado con el deposito de material piroclástico ácido con gran cantidad de fragmentos de vidrio (subtramo C). Los limites entre los subtramos están remarcados por niveles de aglomerado que corresponden a fuertes explosiones. Por otra parte, el hecho de que la deposición del segundo tramo esté condicionada por la paleotopografía y la presencia de laminación interna en algunas capas sugiere que parte de los depósitos han sido generados por caída aérea (airfall). Aunque el afloramiento del tercer tramo es incompleto, puede observarse que el volumen de lava emitido durante el último episodio es muy superior al emitido en los episodios anteriores. La lava es basáltica y, por lo tanto, menos diferenciada que la de los episodios anteriores.

La parte inferior del segundo tramo está atravesada por algunos diques y sills, y además las tobas situadas a ambos lados de la discordancia que separa los tramos inferior e intermedio están fuertemente zeolitizadas. Estas dos observaciones parecen indicar que el contacto entre los dos tramos es una zona de debilidad estructural que es fácilmente alterable.

Los buzamientos regulares y suaves hacia el NNW indican que la zona ha sido estable con posterioridad a los acontecimientos volcánicos que formaron las rocas de la sucesión estudiada. Las únicas estructuras de deformación observadas son fallas normales con débil desplazamiento que han sido posteriormente reutilizadas como fallas direccionales.

Agradecimientos

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Pie de figuras

Fig. 1.- El corte geológico esta situado en Punta Hannah, localizada en la costa Sur de la Isla Livingston, correspondiente a las Shetland del Sur, cercanas al extremo septentrional de la península Antártica.

Fig. 2.- Corte geológico de Punta Hannah con indicación de los tramos, subtramos y capas descritas.

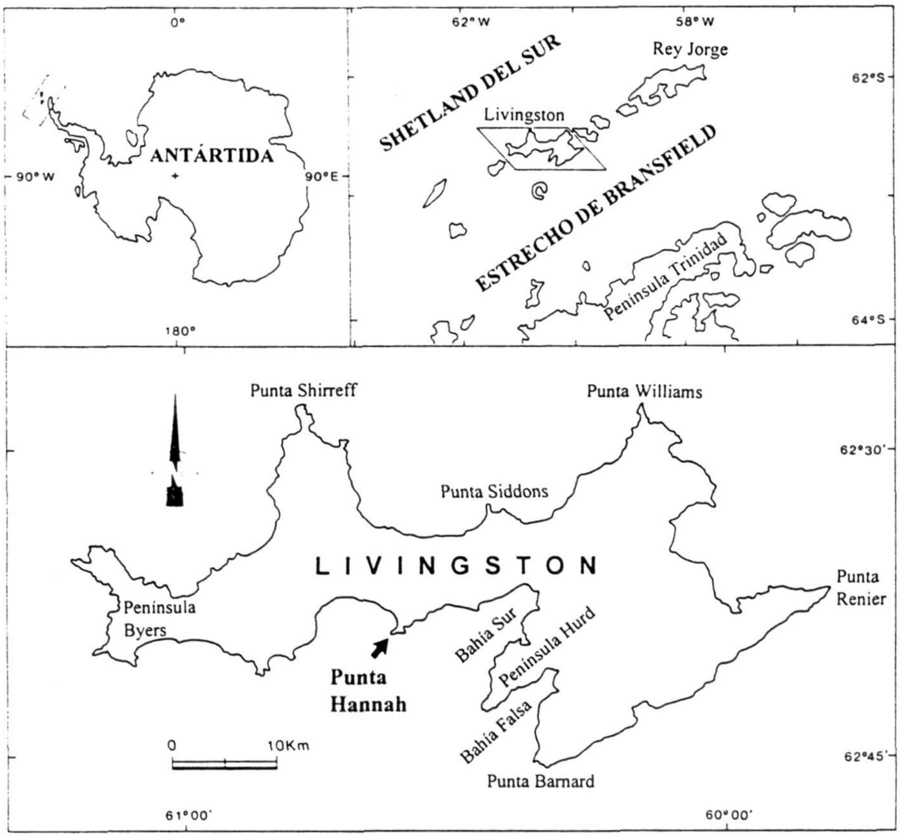
0°
90° W
90°E
180°
ANTÁRTIDA
62° W
58° W
62°S
64°S
SHETLAND DEL SUR
Rey Jorge
Livingston
ESTRECHO DE BRANSFIELD
Peninsula Trinidad
Punta Shirreff
Punta Williams
62°30'
Punta Siddons
LIVINGSTON
Punta Renier
Peninsula Byers
Bahia Sur
Peninsula Hurd
Punta Hannah
Bahia Falsa
0
10Km
62°45'
Punta Barnard
61°00'
60°00'

Fig. 1

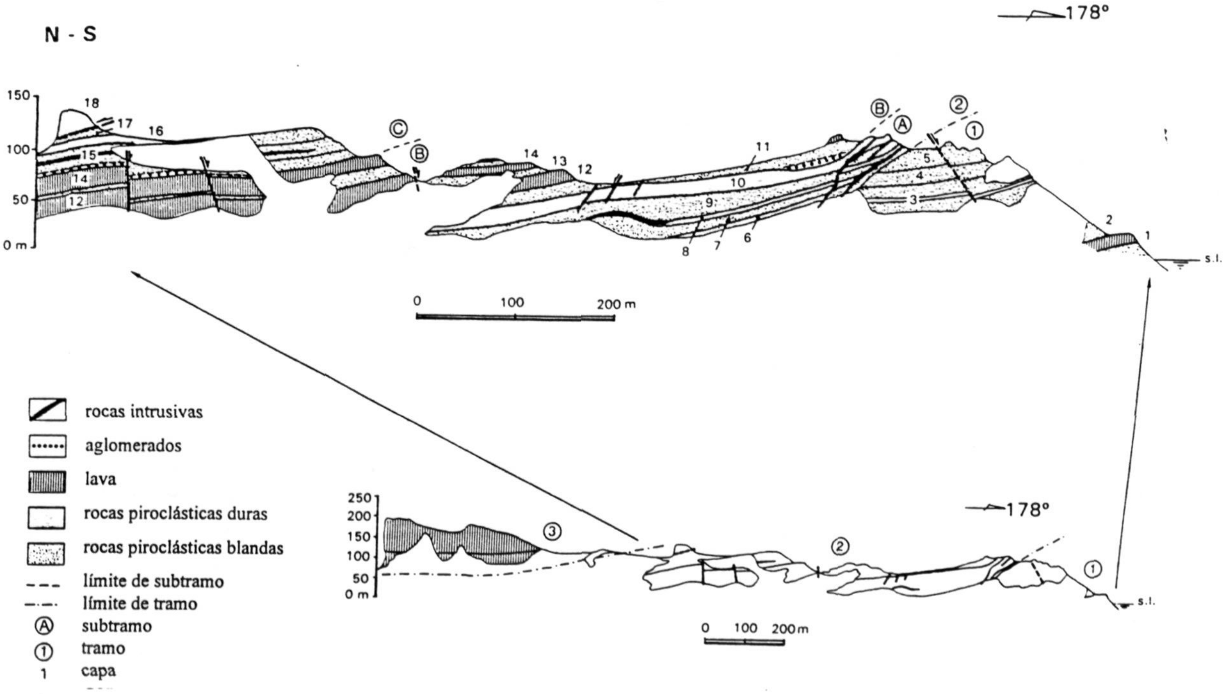
N - S
178°
150
100
50
0 m
18
17
16
15
14
12
Ⓒ
Ⓑ
14
13
12
11
10
9
8
7
6
Ⓑ
Ⓐ
②
①
5
4
3
2
1
s.l.
0
100
200 m
rocas intrusivas
aglomerados
lava
rocas piroclásticas duras
rocas piroclásticas blandas
límite de subtramo
límite de tramo
Ⓐ subtramo
① tramo
1 capa
250
200
150
100
50
0 m
③
②
①
178°
s.l.
0 100 200m

Fig. 2

Smellie,J.L., Sàbat,F., Pallàs,R. & Zheng,X. (in press): Age and correlation of volcanism in central Livingston Island, South Shetland Islands: K-Ar and geochemical constraints. *Journal of South American Earth Sciences,*

Aquest article complementa els dos anteriors tot aportant dades radiomètriques i geoquímiques sobre diverses mostres de roques volcàniques i plutòniques que afloren en les localitats de l'Illa de Livingston que hem estudiat. Les dades radiomètriques estan majoritàriament d'acord amb els esquemes cronoestratigràfics que s'havien proposat anteriorment. Les dades geoquímiques donen informació sobre possibles afinitats entre les successions volcàniques de les diferents localitats.

Age and correlation of volcanism in central Livingston Island, South Shetland Islands: K-Ar and geochemical constraints

J.L. SMELLIE[1], R. PALLÀS[2], F. SÀBAT[2] and X. ZHENG[3]

1. British Antarctic Survey, Natural Environment Research Council, High Cross,
Madingley Road, Cambridge CB3 0ET, U.K.
2. Universitat de Barcelona, Facultat de Geologia, 08028 Barcelona, Spain
3. Institute of Geology, Academia Sinica, Beijing, China

All correspondence should be addressed to J.L. Smellie:
Telephone: (01223) 61188
Fax: (01223) 62616
E-mail: J.Smellie@bas.ac.uk

Abstract —Volcanic sequences in central Livingston Island can be divided into two broad groups. The *older group* consists of basalt-dacite lavas, clastic rocks and associated hypabyssal intrusions. The lavas are lithologically and compositionally similar to other pre-Pliocene, volcanic arc lavas in the South Shetland Islands. The outcrops vary from relatively fresh (at Cape Shirreff, Hannah Point and Siddons Point) to indurated and pervasively altered (at Mount Bowles, Burdick Peak and Hurd Peninsula). Samples from the fresh outcrops yielded Late Cretaceous ages for eruption or intrusion, ranging from 90.2 ± 5.6 Ma at Cape Shirreff, to 73.0 ± 2.3 at Siddons Point. Chemical analyses of the lavas suggest that the sequences at these two outcrops can probably be correlated stratigraphically with the Byers Group and Coppermine Formation, respectively. Two samples from Hannah Point yielded conflicting ages of 87.9 ± 2.6 Ma and 67.5 ± 2.5 Ma from the centre and top of the sequence, respectively. The stratigraphical affinities of the Hannah Point sequence cannot yet be determined unambiguously but it is unlikely to be part of the Byers Group. All of the samples from the altered outcrops (which correspond to the Mount Bowles Formation) yielded Eocene-Oligocene K-Ar ages (44.4 to 35.0 Ma), interpreted as reset ages related to the emplacement and cooling of a nearby Eocene tonalite pluton responsible for much of the alteration, and also dated in this study (43.3 ± 2.8 Ma). A Cretaceous eruptive age (possibly Late Cretaceous) for the altered outcrops is likely but cannot yet be proven. By contrast, the *younger group* consists of degraded basalt lava flows, tuff cone and tuff ring remnants, which are part of the Inott Point Formation. The lavas are very fresh and Pleistocene or Recent in age (≤ 1 Ma). They are compositionally distinctive and are indistinguishable from supra-subduction alkali basalts preserved elsewhere in Livingston, Greenwich and Penguin islands.

Resumen—Las secuencias volcánicas de la zona central de la isla Livingston se pueden dividir en dos grandes grupos. El *grupo antiguo* está formado por lavas basáltico-dacíticas y rocas clásticas, con intrusiones hipoabisales asociadas. La litología y composición de las lavas son similares a las otras lavas de arco volcánico, pre-Pliocenas, existentes en las islas Shetland del Sur. Algunos afloramientos son relativamente frescos (como los de cabo Shirreff, punta Hannah y punta Siddons) mientras que otros estan completamente alterados (como los del monte Bowles, pico Burdick y península Hurd). La edad radiométrica de las rocas de los afloramientos frescos es Cretácico superior, variando desde 90.2 ± 5.6 Ma para el cabo Shirreff a 73.0 ± 2.3 Ma para la punta Siddons. Se considera que esta es la edad de la erupción o intrusión. Los análisis químicos de las lavas sugieren que la secuencia del cabo Shirreff puede correlacionarse con el Grupo Byers, mientras que la de punta Siddons se correlacionaría con la Formación Coppermine. Las edades radiométricas de las muestras de punta Hannah son 87.9 ± 2.6 Ma y 67.5 ± 2.5 Ma, del centro y la cumbre de la secuencia, respectivamente. Por el momento no se han podido determinar de forma precisa las afinidades estratigráficas de la secuencia de punta Hannah, pero es improbable que esta forme parte del Grupo Byers. Las muestras de los afloramientos alterados (que corresponden a la Formación del monte Bowles) tienen edades K-Ar correspondientes al Eoceno-Oligoceno (de 44.4 a 35.0 Ma). Se ha interpretado que estas edades corresponden principalmente al emplazamiento y enfriamiento de un plutón tonalítico Eoceno (43.3 ± 2.8 Ma) que aflora en las proximidades y al que se considera responsable de la alteración de los afloramientos de la zona del monte Bowles. Aunque no se ha podido demostrar, es probable que la erupción de las rocas que constituyen estos afloramientos alterados también se produjese durante el Cretácico, quizá el Cretácico superior. Por otra parte, el *grupo moderno* está constituido por coladas de basalto parcialmente erosionadas y por restos de conos y acumulaciones anulares de escoria que forman parte de la Formación punta Inott. Estas lavas son muy frescas y su edad máxima es Pleistoceno (≤ 1 Ma). Tienen una composición peculiar y son muy parecidas a los basaltos alcalinos de supra-subducción que afloran en otros puntos de las islas Livingston, Greenwich y Pinguino.

INTRODUCTION

VOLCANIC OUTCROPS in central Livingston Island were revisited and mapped during three austral summers, between 1989 and 1992 by successive Spanish Antarctic expeditions. The outcrops occur between Cape Shirreff and Mount Bowles and samples for dating were obtained at ten localities (Fig. 1). They are not continuous because of extensive snow and ice cover, and age relationships are generally indeterminate due to lack of exposure. The outcrops comprise weakly deformed and indurated, variably altered basalt-andesite lavas and pyroclastic rocks locally interbedded with sedimentary rocks, and a group of undeformed, fresh, friable, basalt lavas and tephra (Hobbs, 1968; Smellie et al., 1984; Smellie et al., in press). All of the volcanic sequences are unfossiliferous and they lack published isotopic ages. Ages were previously assigned on the basis of somewhat speculative regional comparisons and range from Late Jurassic to late Cenozoic (Hobbs, 1968; Smellie et al., 1984). In this paper, we report the results of the first K-Ar isotopic study of the volcanic sequences in central Livingston Island, in an attempt to constrain their ages more closely.

SEQUENCE AND SAMPLE DETAILS

Older volcanic sequence

Cape Shirreff consists of about 450 m of porphyritic basaltic lavas and minor volcanic breccias. The rocks are deformed into open folds with NW-SE-trending axes and subvertical axial surfaces intruded by numerous dykes. A sample of fresh olivine basalt (325c3) was obtained from a lava or sill on the southern side of the outcrop. It is holocrystalline, with c. 4 % olivine and 10 % plagioclase phenocrysts in a groundmass of plagioclase, clinopyroxene and opaque oxide. Alteration is restricted to

minor smectite and partial replacement of olivine by serpentine in fractures. By contrast, Hannah Point consists of a homoclinal, NNW-dipping sequence with a thickness of about 500 m, comprising basaltic-andesite to dacite lavas interbedded with a variety of pyroclastic and thin sedimentary beds. Two lava samples were obtained: both are basaltic andesites, obtained from the top (434) and just above the centre (407) of the sequence. The samples are very similar petrographically. They are essentially aphyric, consisting of pilotaxitic plagioclase and intergranular clinopyroxene and opaque oxide. Alteration consists mainly of finely crystalline smectite, mainly affecting plagioclase, and traces of (?) zeolite. Sample 407 is the more pervasively altered of the two samples and it also contains diffuse veinlets of carbonate. A thick (c. 90 m) sill was sampled at Siddons Point. It shows gently dipping, layer-like structures similar to dolerite sills elsewhere in the South Shetland Islands, which are spectacularly developed in the Coppermine Formation (cf. Smellie et al., 1984). The sample obtained (278) is a fresh olivine dolerite formed of coarse interlocking plagioclase, clinopyroxene, olivine and opaque oxide. Orthopyroxene is a rare constituent. Alteration is restricted to minor replacement of olivine by opaque oxide, talc and (?) iddingsite.

The numerous, isolated volcanic outcrops near Mount Bowles and between Burdick Peak and False Bay consist of basalt-andesite lavas, volcanic breccias and thin arkosic sandstones, which are included in the Mount Bowles Formation (Smellie et al., in press). They are noticeably more indurated, closely jointed and veined than the Cape Shirreff, Hannah Point and Siddons Point outcrops and bedding is typically indistinct (Smellie et al., in press). Samples for dating were obtained near Willan Nunatak (45D, andesite lava), east of Mount Bowles (262, basaltic andesite lava) and near Moores Peak (105E, basalt lava). Although original, igneous textures are preserved, the primary minerals are partly (45D, 262) to completely (105E) recrystallised to a secondary assemblage comprising albite-quartz-chlorite-epidote-actinolite-sphene. Biotite is conspicuous in samples from near Willan Nunatak and Moores Peak but absent from Mount Bowles samples. Anastomosing microfractures, some marked by abundant actinolite and epidote, are common in sample 105E. Veins of epidote-quartz-plagioclase±actinolite occur more sparsely in other samples, and there are traces of carbonate in sample 80A. The alteration assemblages were interpreted by Smellie et al. (1984, in press) as predominantly an effect of widespread thermal metamorphism and hydrothermal activity caused by plutonic intrusions on the island, although some may be related to Cretaceous volcanism (Willan et al., 1994). All the samples are lavas. Sample 45D is highly porphyritic, containing abundant plagioclase (up to 60 % by volume), altered pyroxene (10-15 %) and opaque oxide (5%) phenocrysts. Samples 105E and 262 are virtually aphyric.

Younger volcanic sequence

The small outcrops SE of Burdick Peak (230) and Samuel Peak (260) are very similar, consisting of very fresh basalt lava interbedded with black or red scoria (Smellie et al., in press). In lithology, the lavas closely resemble the fresh lava from north-west of Burdick Peak (Fig. 1). (This is the so-called "Gleaner Heights lava" sampled by Hobbs (1968) and analysed by Smellie et al. (1984)). The lavas dated here contain abundant (c. 10-12 %) unaltered olivine, minor clinopyroxene (c. 5 %) and rare plagioclase phenocrysts in a fine mesostasis of plagioclase, clinopyroxene, opaque oxide and olivine.

K-AR RESULTS

All samples were ground in a jaw crusher and sieved to 125-180 µm, washed, then analysed at the NERC Isotope Geosciences Laboratory at Keyworth, Nottingham, England, using standard K-Ar analytical methods (summarised in Pankhurst & Smellie, 1983). The results are presented in Table 1.

Older volcanic sequences

On Livingston Island, volcanism began in earliest Cretaceous times (from c. 130 Ma) at Byers Peninsula, and continued until Campanian times (75-80 Ma), at least, in north-eastern outcrops (Pankhurst et al., 1979; Smellie et al., 1984, Rees & Smellie, 1989, Chapman & Smellie, 1992). The Late Cretaceous ages obtained from the samples from Cape Shirreff (90.2 ± 5.6 Ma) and middle Hannah Point (87.9 ± 2.6 Ma) fall within this range and agree with the model of easterly younging of volcanism within the South Shetland Islands developed by Pankhurst and Smellie (1983). Thus, these two ages may reflect eruption ages for the volcanic sequences at the two localities. However, the Cape Shirreff sample may be a sill, thus providing only a minimum age for the locality, and dating of a second, stratigraphically higher sample from Hannah Point yielded a much younger age (67.5 ± 2.5 Ma). If the entire sequence at Hannah Point is coeval, the simplest interpretation is that the upper lava has lost argon and gives an apparent age which is too young. The age of the upper lava is within the range of dated dolerite intrusions on Robert, Greenwich and north-eastern Livingston islands (Coppermine Formation). The volcanism at these localities probably reached a climax around 85-70 Ma and was associated with widespread intrusion of voluminous dolerite sills, some as young as 55-60 Ma (Pankhurst & Smellie, 1983; Fensterseifer et al., 1992). Thus, the age of the upper lava may have been reset by local thermal effects caused by dolerite intrusions not presently exposed at Hannah Point. However, the middle lava is more altered than the upper lava and contains traces of carbonate, possibly rendering the age unreliable. For example, secondary carbonate may contain excess ^{40}Ar (Barker et al., 1982). Thus, the interpretation of the Hannah Point samples is ambiguous and has not yet been resolved, although a general, Late Cretaceous age is indicated. The dolerite sill at Siddons Point (73.0 ± 2.3 Ma) probably belongs to the Late Cretaceous episode of dolerite intrusion associated with the Coppermine Formation.

By contrast, the three dated samples from the pervasively altered sequences in eastern-central Livingston Island (from east of Mount Bowles, near Willan Nunatak and near Moores Peak) yielded Eocene ages (39.8 ± 1.6, 44.4 ± 1.3 and 35.0 ± 3.9 Ma, respectively). Although statistically different, the separation in apparent ages is not great and two are indistinguishable within error. Given the evidence for pervasive alteration, all three ages probably reflect variable post-eruptive Ar-loss caused by a thermal event, most probably related to the large plutonic intrusions mapped in south-eastern Livingston Island. To test this possibility, two samples of tonalite (258a and b) from Willan Nunatak were also dated. They yielded ages of 41.4 ± 1.3 Ma (whole rock) and 43.3 ± 2.8 Ma (hornblende separate), which are indistinguishable within error of each other. The mineral separate age is also indistinguishable from the emplacement age of the Barnard Point tonalite (46±1 Ma; Smellie et al., 1984). Thus, the tonalite at Willan Nunatak may be regarded as an extension of the Barnard Point pluton, which it also resembles in many petrographical and compositional characteristics. The age of lava sample 45D (44.4 ± 1.3 Ma) is indistinguishable from the emplacement age of the Barnard Point tonalite. Conversely, the ages of lava samples 262 (39.8 ± 1.8 Ma) and 105E (35.0 ± 3.9 Ma) are essentially indistinguishable from that obtained on biotite from the Barnard Point tonalite (40 ± 1 Ma). Therefore, the ages of samples 45D, 262 and 105E may reflect either reheating (by the tonalite, or possibly associated with dyke intrusion; cf. Smellie, 1983), or uplift and first cooling of the tonalite pluton to c. 200° C (Smellie et al., 1984). However, the Barnard Point pluton also contains evidence of multiple hydrothermal alterations (Armstrong & Willan,1994). They include at least two post-magmatic episodes of mid- to late Cenozoic age, which could also be responsible for the slightly younger ages obtained on samples 262 and 105E.

Thus, the ages obtained on the pervasively altered, older volcanic sequence in eastern-central Livingston Island merely record the age of the youngest post-eruptive alteration. They give no clear indication of the possible eruptive age(s) of the sequences. In the model of north-easterly migration of volcanism developed by Pankhurst & Smellie (1983), based on a regional study and about 70 K-Ar ages, the age of volcanism in central Livingston Island should fall between that of Byers Peninsula and Williams Point. In consideration of that model, which is possibly supported by the new ages from Cape Shirreff, Hannah Point and Siddons Point reported here, we favour a Cretaceous (possibly Late Cretaceous) eruption age for the older volcanic sequence in eastern-central Livingston Island. This is broadly supported by a poorly constrained, mid-Cretaceous Rb-Sr age of about 113 Ma obtained by Willan et al. (1994) for metasomatism accompanying hydrothermal alteration of mudstones from Hurd Peninsula; the alteration was believed to be coeval with volcanism on Livingston Island. Furthermore, altered volcanic rocks at Half Moon Island (Fig. 1) are intruded by a tonalite with an age of 102 Ma (Grikurov et al., 1970; age recalculated by Smellie et al., 1984), which also suggests the possible presence of pre-Late Cretaceous volcanism in the area. Finally, although the Eocene tonalite at Barnard Point and locally abundant basic-intermediate dykes, some of which are also of Eocene age (Grikurov et al., 1970; Smellie, 1983), may have fed Eocene volcanism, there is currently no evidence of Eocene volcanic sequences on Livingston Island.

Younger volcanic sequence

Samples 230 and 260 are much younger (0.04 ± 0.35 and 0.70 ± 0.30 Ma, respectively) than any other samples dated, consistent with their very fresh appearance. However, the measured volume of radiogenic ^{40}Ar in sample 230 is below the limit of detection (as indicated by the calculated error) and an age of ≤ c. 1 Ma for eruption is probably all that is constrained by the data for both samples (i.e. essentially too young to date). In freshness, age and geochemistry, samples 230 and 260 are clearly distinguished from the Cretaceous-Neogene arc volcanic rocks which form the bulk of the South Shetland Islands volcanic arc (see below). They are correlated unambiguously with the numerous, isolated sequences of very young, fresh lava, tuff and cinder cone relicts, which are common in north-eastern Livingston and northern Greenwich islands (Inott Point Formation: Smellie et al., in press).

GEOCHEMISTRY

All the dated samples were analysed on an ARL 8420 automatic XRF spectrometer at the University of Keele (UK) using standard XRF procedures, together with other samples representative of the volcanic outcrops visited (Table 2). The geochemistry was used to test for any petrological affinities within the outcrops visited and to look for compositional similarities with better-defined, dated volcanic sequences elsewhere on Livingston and Greenwich islands (Byers Group and Coppermine Formation). In the absence of other evidence, a compositional affinity between groups of samples may be used to infer stratigraphical affinities and a similar eruptive age. It is not the intention of this paper to give a detailed petrogenetic appraisal of these sequences.

Older volcanic sequence

All of the samples from the older sequences show the high LIL/HFS (large ion lithophile/high field strength) element ratios and Nb depletion of arc magmas, with the moderate Zr/Y ratios characteristic of continental arcs (Fig. 2). TiO_2 and P_2O_5 correlate positively with SiO_2 and Zr, and Rb/Sr and Ce_n/Y_n ratios are relatively low, which are all features of tholeiitic arc magmas, and this is also indicated on various calc-alkaline:tholeiitic discriminant plots (e.g. FeO*/MgO *vs* SiO_2 (Miyashiro, 1974), MnO-TiO_2-P_2O_5 (Mullen, 1983)). Conversely, most measured element abundances are higher than in typical arc tholeiites and these and some other element ratios (e.g. Na_2O/K_2O, K/Rb) are more comparable with calc-alkaline rock series. Similar, conflicting, tholeiitic-calc-alkaline affinities are a characteristic, distinctive feature of arc volcanism in the South Shetland Islands between Cretaceous and Miocene times and have been ascribed to the trench-proximal location of the volcanism, astride the thin, leading edge of the Antarctic Peninsula magmatic arc (Smellie, 1979).

The samples from Cape Shirreff, Hannah Point and Siddons Point are essentially fresh and their analyses should be a reliable guide to the compositions of the original magmas from which they crystallised. The compositions of the three samples are displayed as multi-element spidergrams in Fig. 2a. The two Hannah Point samples are basaltic andesites showing very similar, parallel patterns of smoothly diminishing element concentrations passing from the LIL to the HFS elements. Incompatible element concentrations are enriched and compatible elements depleted in sample 407 relative to 434, suggesting a cogenetic relationship consistent with fractionation of phases such as calcic plagioclase, pyroxene and olivine. Notwithstanding small inconsistent variations (e.g. crossing tie lines for Ba, P), the spidergram for the Siddons Point sample is very similar to those for the Hannah Point samples, although it is not possible to demonstrate a cogenetic relationship. By contrast, the olivine basalt from Cape Shirreff is substantially different in composition from all three samples from Hannah and Siddons points. In particular, the Cape Shirreff spidergram is much flatter, with a more rapid increase in HFS element concentrations passing from Y to Nb, and relatively low LIL element concentrations relative to the other samples. The crossing tie lines and lower LIL/HFS element ratios compared with samples from Hannah Point and Siddons Point are inconsistent with a cogenetic relationship between the three localities by either simple crystal fractionation (or accumulation) or variable partial melting of a similar source.

Because of pervasive alteration, the compositions of samples from eastern-central Livingston Island are likely to have undergone post-depositional element migration. The spidergrams shown in Figs. 2b and c are unlikely to reflect accurately the original, magmatic abundances of the LIL elements (particularly K, Rb, Ba), at least, although the precise effects of the alteration(s) are unquantified and cannot be addressed in this study. Samples 43, 44 and 262 (Mount Bowles) are basaltic andesites, whereas sample 105E (Moores Peak) is a relatively primitive basalt with high MgO (9 Wt %), Ni (160 ppm) and Cr (488 ppm) contents and high Mg-number (61.9). The crossing tie lines for the LIL elements are probably a function of variable element mobility, whereas the spidergram patterns for the HFS elements are essentially parallel and could suggest a cogenetic relationship between the four samples. Samples 45D and 80A (near Willan Nunatak) are compositionally very similar low-silica andesitic rocks. LIL/HFS element ratios are higher than in the Mount Bowles and Moores Peak samples but they are undoubtedly affected by alteration. Furthermore, although the HFS element patterns are also quite distinct, this is probably caused by preferential extraction by Ti- and P-bearing phases (probably Ti-magnetite and apatite; cf. Smellie et al., 1984). The affinities of samples 45D and 80A are not constrained by the geochemical data available and are unknown.

Also shown in Fig. 2d are spidergrams for representative basalts from the Byers Group and Coppermine Formation (data from Smellie et al., 1984 and unpublished), the outcrops of which flank the area discussed in this paper. There are no LIL/LIL or HFS/HFS element ratios diagnostic of either rock group, but a most striking feature of the data is the marked contrast in LIL abundances and LIL/HFS element ratios, which are higher in the Coppermine Formation compared with the Byers Group. When compared with these data, the Cape Shirreff sample is indistinguishable from basalts in the Byers Group. Cape Shirreff is also geographically close to Byers Peninsula and we suggest that a stratigraphical correlation with the Byers Group is likely. Samples from Siddons Point and Hannah Point can be more closely matched with those of the Coppermine Formation than the Byers Group, except in respect of Y-Zr and Ce/Y ratios. In lithology and field appearance as well as geochemical similarity, the Siddons Point outcrop closely resembles widespread, thick and voluminous sills associated with the Coppermine Formation described by Smellie et al. (1984), and we suggest that a stratigraphical correlation is likely. Conversely, the stratigraphical affinities of the Hannah Point sequence are uncertain, although it is unlikely to be part of the Byers Group despite its geographical proximity to Byers Peninsula. Because of the unquantified effects of alteration on original, magmatic, LIL element abundances in lavas from the Mount Bowles Formation in eastern-central Livingston Island, any geochemical-based correlation with either the Byers Group or Coppermine Formation is unjustified at present.

Younger volcanic sequence

In composition, the fresh lava samples obtained near Samuel Peak (260) and Burdick Peak (230) are primitive olivine basalts with enhanced LILE concentrations relative to HFSE, Nb-"depletion" and low $(Ce/Y)_n$ ratios (1.8-3.7) comparable with island arc tholeiites, but high total alkalis and Na/K ratios and moderate ne contents (3.5-3.7) similar to sodic alkali basalts. There is no compositional similarity to any of the pre-Pliocene arc lavas in the South Shetland Islands, nor to any of the Recent Bransfield Strait volcanoes (Deception and Bridgeman Islands, and seamounts SW of Bridgeman Island; Weaver et al., 1979; Smellie et al., 1984; Smellie, 1987; Keller et al., 1992). However, these distinctive characteristics are found in compositionally unusual, alkalic lavas in Pleistocene-Recent outcrops in Livingston, Greenwich and Penguin islands (Weaver et al., 1979; Smellie, 1990). Although spreading in Drake Passage essentially ceased about 4 Ma ago (Barker, 1982), the subducted slab is present beneath the South Shetland Islands and is still being consumed, albeit very slowly (Larter 1991). Thus, the Pleistocene-Recent volcanism can be considered as supra-subduction. These lavas are unique in the South Shetland Islands and were erupted as a consequence of regional extension and splitting of the South Shetland arc during the latest stages of (very slow) subduction and formation of the Bransfield Strait marginal basin (Smellie, 1987, 1990; Smellie et al., in press).

CONCLUSIONS

Recent mapping in central Livingston Island has confirmed the presence there of two broad groups of unfossiliferous volcanic sequences. The older group includes relatively fresh basalt and andesite lavas, which are dated at 90.2 ± 5.6 Ma (Cape

Shirreff), 87.9 ± 2.6 or 67.5 ± 2.5 Ma (Hannah Point) and 73.0 ± 2.3 Ma (Siddons Point). The new ages, together with chemical analyses, provide the first evidence for correlating the sequences at these three localities with the Byers Group and Coppermine Formation and the ages are consistent with a published model of north-easterly migration of volcanism in the South Shetland Islands. By contrast, the volcanic rocks in eastern-central Livingston Island (Mount Bowles Formation) are indurated and pervasively altered due to the proximity of large plutons. New K-Ar ages of 44.4±1.3, 39.8 ± 1.6 and 35.0±3.9 Ma were obtained on altered lava samples and are interpreted as reset ages related to the emplacement and cooling of the Barnard Point tonalite (46 ± 1 Ma). They provide no guide to the age(s) of eruption of the volcanic sequence(s). A Late Cretaceous age is likely for the altered volcanic sequence(s), but it remains to be proven. If true, some of the eastern-central outcrops may represent the source of the pyroclastic flow deposits, airfall tuffs and abundant very coarse volcanic detritus in the Williams Point Beds (Cenomanian-early Campanian) in north-eastern Livingston Island, for which a local provenance (possibly < 20 km) has been postulated but whose location was unknown (Chapman & Smellie, 1992).

The younger volcanic group (part of the Inott Point Formation) contains unaltered olivine basalt lavas, which yielded very young ages (≤ 1 Ma). The lavas are compositionally distinctive and are unequivocally correlated with Quaternary, supra-subduction, alkalic lavas described elsewhere from Livingston, Greenwich and Penguin islands.

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FIGURE CAPTIONS

Figure 1. Geological sketch map of Livingston Island and its location. Also shown are the locations of samples dated and analysed in this study. Modified after Smellie et al. (in press). Abbreviations: ByP - Byers Peninsula; WN - Willan Nunatak; BP - Burdick Peak; MP - Moores Peak; SP - Samuel Peak; GH - Gleaner Heights.

Figure 2. Multi-element spider diagrams illustrating the compositional characteristics of samples from Livingston Island analysed in this study (a-c) and representative analyses of basalts from the Byers Group (P.726.5) and Coppermine Formation (P.841.1) (from Smellie et al., 1984), for comparison (d). Normalising values are those of Sun & McDonough (1989).

Table 1. New K-Ar ages of rock samples from Livingston Island

Sample number	Rock type	Sample location	Wt% K	^{40}Ar rad (nl/g)	% atmos Ar	Age (Ma)[1]	Interpretation
230	basalt lava	1.6 km SE of Burdick Pk.	0.247	0.0004	99.98	0.04±0.35	eruption age ≤ 1 Ma
260	basalt lava	1 km S of Samuel Pk.	0.432	0.0117	95.37	0.70±0.35	eruption age ≤ 1 Ma
258a	tonalite (WR)	Willan Nunatak	0.410	0.6676	40.54	41.4±1.3	reset age
258b	onalite (HBL)	Willan Nunatak	0.815	1.3891	27.91	43.3±2.8	emplacement age
105E	basalt lava	0.6 km NE of Moores Pk.	0.267	0.3660	83.95	35.0±3.9	reset age
45D	andesite lava	0.9 km W of Willan Nun.	0.835	1.4590	32.78	44.4±1.3	reset age
262	basaltic andesite lava	5.5 km E of Mt Bowles	0.436	0.6826	59.61	39.8±1.6	reset age
278	dolerite sill	Siddons Point	0.363	1.0515	42.63	73.0±2.3	emplacement age
434	basaltic andesite lava	Hannah Point (top)	0.539	1.4411	55.26	67.5±2.5	eruption/reset age
407	basaltic andesite lava	Hannah Point (middle)	1.410	4.9352	36.44	87.9±2.6	eruption/reset age
325c3	olivine basalt lava/sill	Cape Shirreff	0.267	0.9601	73.74	90.2±5.6	eruption/emplacement age

[1] errors quoted are 2σ

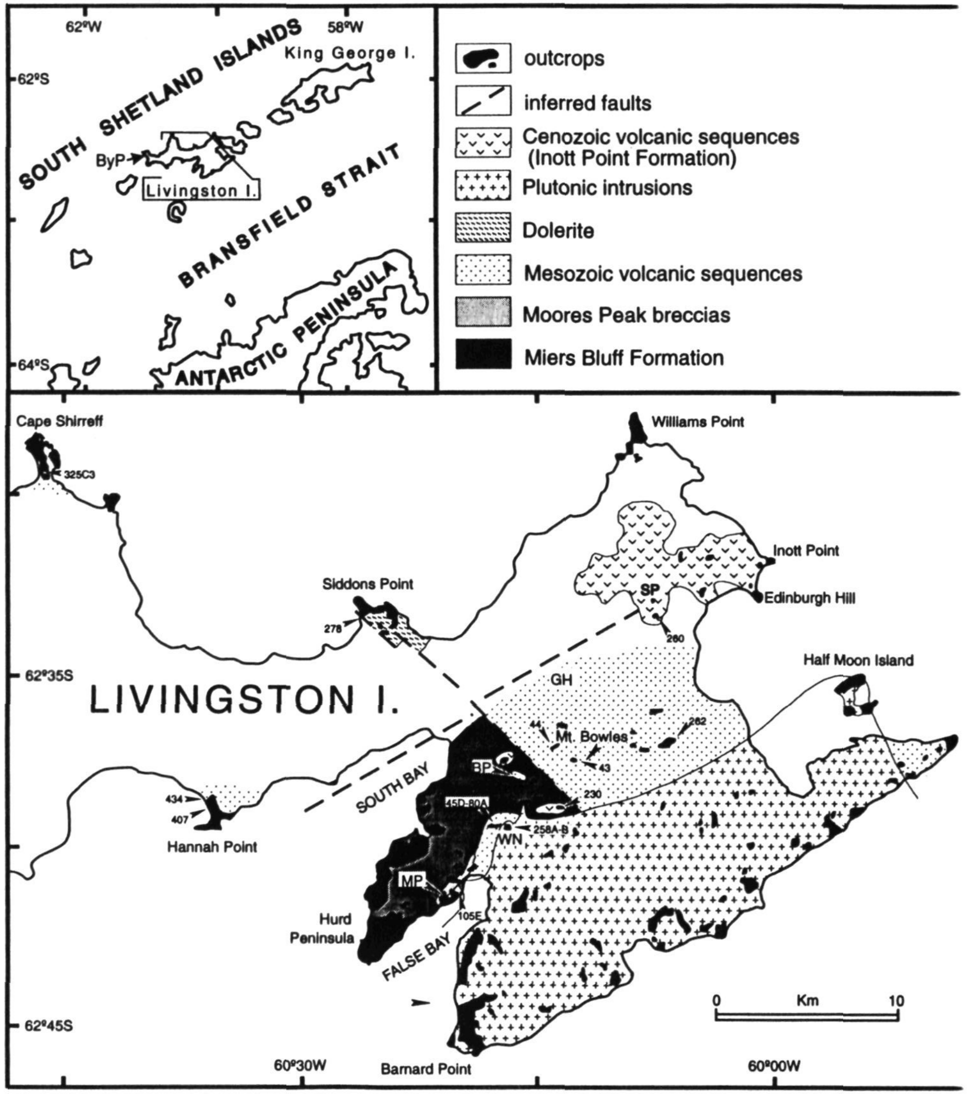
Abbreviations: WR - whole rock; HBL - hornblende mineral separate

Table 2. New geochemical analyses of volcanic and plutonic rocks from Livingston Island

Sample	230	260	43	44	262	45D	80A	105E
Locality	SE Burdick	S Samuel	Mt Bowles	Mt Bowles	E Mt Bowles	W Willan	W Willan	E Hurd P
Rock type	lava	lava	lava	lava	lava	lava	lava	lava clast
SiO2	46.64	47.4	53.84	54.05	53.13	57.49	55.87	47.22
TiO2	0.99	1.15	1.29	1.73	1.27	0.96	1.05	1.08
Al2O3	16.21	16.16	17.09	15.35	16.04	17.29	17.04	16.96
Fe2O3T	10.31	10.51	9.72	11.35	10.93	8.25	9.11	10.98
MnO	0.17	0.17	0.19	0.21	0.18	0.14	0.15	0.18
MgO	11.42	10.96	3.59	3.6	4.89	3.45	3.76	9
CaO	10.7	10.37	8.73	5.68	5.92	6.91	7.38	9.98
Na2O	2.92	3.2	3.54	4.35	4.67	3.67	3.81	1.83
K2O	0.34	0.46	0.32	1.01	0.49	1.29	1.07	0.3
P2O5	0.18	0.21	0.23	0.27	0.21	0.19	0.18	0.17
LOI	-0.38	-0.5	1.66	2.2	2.14	0.73	0.89	2.38
Total	99.5	100.09	100.2	99.79	99.89	100.38	100.31	100.09
Cr	594	443	31	15	92	20	32	488
Ni	216	193	9	11	18	12	16	160
Rb	7	10	10	22	16	42	28	13
Ba	85	126	143	298	218	297	309	67
Sr	418	536	245	214	771	428	453	305
Nb	2	6	4	7	4	4	3	4
Zr	53	57	101	153	96	148	132	79
Y	14	17	24	32	26	27	24	17
V	287	288	265	416	365	216	256	247
La	0	1	7	7	4	4	13	3
Ce	11	27	28	34	21	32	29	28
Nd	14	27	14	19	24	28	34	23
Th	0	2	3	4	1	6	1	3

All samples analysed by XRF
LOI - loss on ignition

407	434	325c3	278	258a	258b
Hannah	Hannah	Shirreff	Siddons	Willan	Willan
lava	lava	lava	sill	pluton	pluton
51.76	52.6	48.55	50.42	59.75	58.32
1.39	1.21	1.24	0.97	0.59	0.62
15.92	16.51	18.12	17.86	18.04	18.46
11.43	11.27	10.85	9.68	6.22	6.57
0.2	0.18	0.17	0.16	0.12	0.11
4.63	4.52	6.52	6.74	3.06	3.17
7.59	8.71	10.61	10.74	6.68	7.18
3.64	3.18	2.82	2.56	3.77	3.77
1.59	0.62	0.31	0.5	1.29	1.09
0.26	0.13	0.29	0.16	0.13	0.14
1.56	1.04	0.81	0.4	0.38	0.46
99.97	99.96	100.29	100.19	100.03	99.89
43	60	128	126	19	18
27	9	55	38	12	11
34	15	6	11	38	35
272	115	81	186	336	277
434	498	483	622	575	597
6	4	5	3	3	3
142	86	100	64	99	95
36	25	22	16	23	27
309	395	255	259	142	144
10	0	3	1	1	4
35	19	39	13	27	25
16	11	22	17	14	10
2	1	0	1	2	3



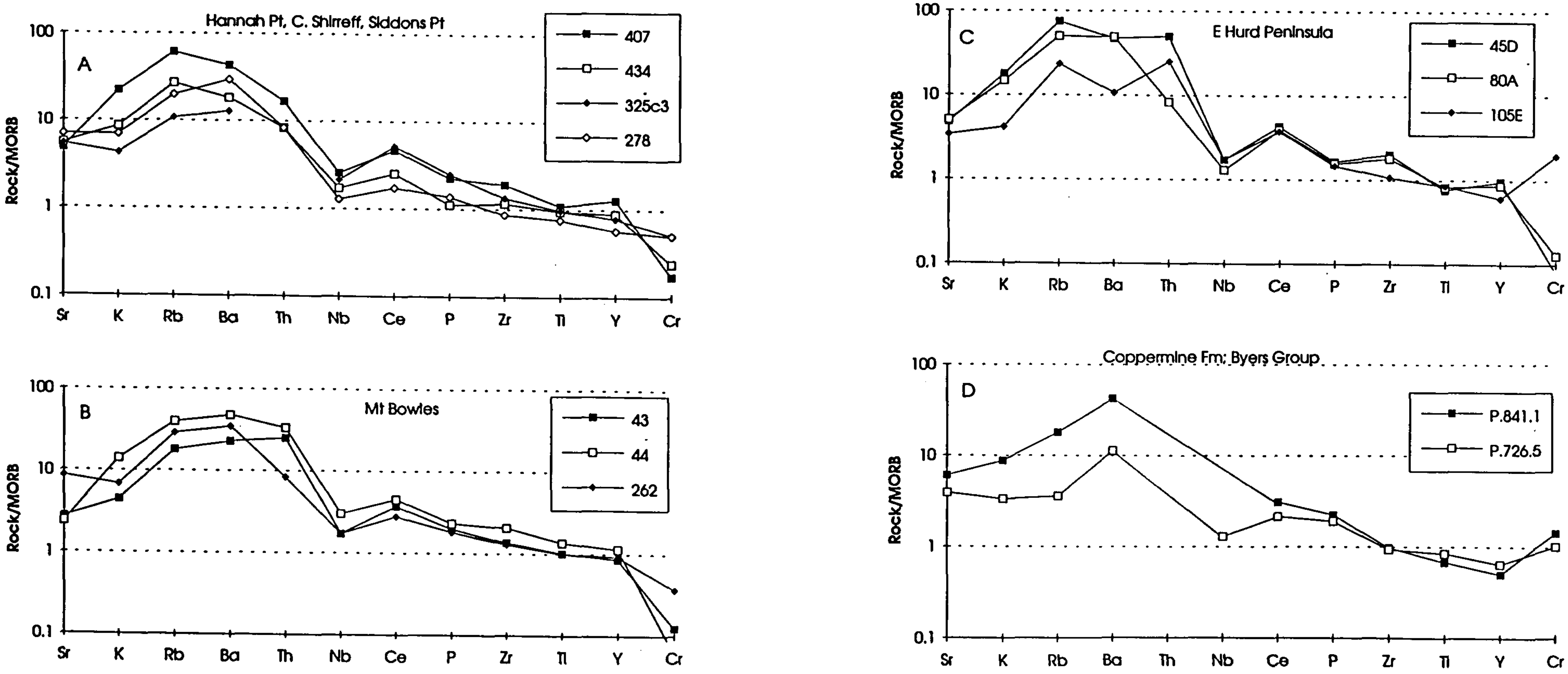

Hannah Pt, C. Shirreff, Siddons Pt
A
100
10
1
0.1
Rock/MORB
407
434
325c3
278
Sr
K
Rb
Ba
Th
Nb
Ce
P
Zr
Tl
Y
Cr
E Hurd Peninsula
C
45D
80A
105E
Mt Bowles
B
43
44
262
Coppermine Fm; Byers Group
D
P.841.1
P.726.5

Fig.2

López-Martínez,J., Vilaplana,J.M., Martínez de Pisón,E., Calvet,J., Arche,A., Serrat,D. & Pallàs,R. (1992): Geomorphology of selected areas in Livingston Island, South Shetland Islands. *In: Geologia de la Antártida Occidental. (López-Martínez, J. ed.) Simposios T3, pp.271-281. III Congreso Latinoamericano de Geología. Salamanca, España. 1992*

Aquest article és principalment un recull de mapes i esquemes geomorfològics d'algunes localitats de l'Illa de Livingston. Cada un dels mapes s'acompanya amb una descripció breu dels trets geomorfològics més importants, conjuntament amb diversos perfils topogràfics de les platges aixecades. S'hi destaquen l'aixecament recent de l'illa i la retracció dels fronts glacials.