GEOLOGY OF A MESOZOIC INTRA-ARC SEQUENCE ON BYERS PENINSULA, LIVINGSTON ISLAND, SOUTH SHETLAND ISLANDS

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ABSTRACT. The late Jurassic (? Lower Tithonian) to Lower Cretaceous (? Barremian) Byers Formation of eastern Livingston Island is a complex sequence of intertonguing volcanic and volcanogenic rocks which are believed to have accumulated within a volcanic arc. It can be related to similar successions in Patagonia and South Georgia, and forms part of the Upper Mesozoic arc of the Scotia arc region. Four constituent members are defined on lithological/facies criteria but they are difficult to delineate geographically because of the poor exposure and complex inter-relationships. Marine, terrestrial and lacustrine environments are all represented. The sequence is interpreted as one that began with open-shelf marine sedimentation but which became progressively more enclosed by the development of surrounding volcanoes. It is possible that the marine basin of the Byers Peninsula area became completely infilled with volcanic detritus and that terrestrial accumulations eventually covered the whole area. Marine faunas of (?) Lower Tithonian, (?) Berriasian and Valanginian age occur in the lower parts of the formation, whereas a Lower Cretaceous flora is present towards the upper part of the succession.

BYERS PENINSULA (Figs 1 and 2) is the largest ice-free area in the South Shetland Islands. Situated at the exposed western end of Livingston Island, it occupies an analogous position to a slightly smaller ice-free area (Fildes Peninsula) on King George Island. However, the geological potential of such an unusually large exposure of rock in the Antarctic Peninsula area is considerably reduced by the large amount of debris cover, formed as a result of extensive periglaciation.

Although the South Shetland Islands were discovered in 1819 (Miers, 1820), no systematic geological exploration was attempted on Livingston Island until the summer of 1957–58, when G. J. Hobbs of the Falkland Islands Dependencies Survey carried out reconnaissance geological observations in conjunction with a topographical survey (Hobbs, 1968). Hobbs covered the greater part of Byers Peninsula in several long day excursions from a camp near the south-eastern extremity of the peninsula. He recorded a mixed sequence of stratified volcanic rocks with interbedded conglomerates and sandstones, which he assigned to the "Younger Volcanic Group" of Miocene age, on the basis of general comparisons with the rocks of King George Island (Barton, 1965). Three volcanic plugs were mapped near Devils Point, Vietor Rock and Negro Hill.

Subsequently, the area received intensive study by Chilean geologists, who discovered late Jurassic and early Cretaceous ammonites in the central western part of the peninsula (González-Ferrán and others, 1970; Tavera, 1970; Covacevich, 1976), and a Lower Cretaceous flora in the eastern part (Hernández and Azcárate, 1971). Thus the Miocene age first suggested by Hobbs was no longer tenable and the geological map was substantially revised by Valenzuela and Hervé (1971). Nonetheless, they considered the agglomerates of Ray Promontory to be a younger unit (of unspecified age), separated from the older rocks by an angular unconformity.

The present study was undertaken in February and March 1975 as part of a geological project to re-map the South Shetland Islands (excluding the Elephant and Clarence Islands group and Deception Island). Important stratigraphical revisions (Table I) are the suggestion that the previously supposed younger agglomerates of Ray Promontory form part of the bedded Mesozoic sequences exposed elsewhere on Byers Peninsula, and the re-interpretation of the "breccia-pipe" deposits, mapped by Valenzuela and Hervé (1971), as sills or lavas intruded or erupted into wet sediments. The bedded sequence of Byers Peninsula is formally described here as the *Byers Formation*.

In conjunction with the authors' field work, a programme of sampling for radiometric dating was carried out by A. D. Saunders and S. D. Weaver of the University of Birmingham.

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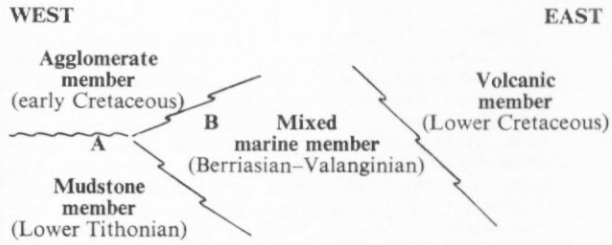
TABLE I. EVOLUTION OF THE STRATIGRAPHICAL INTERPRETATION OF THE BEDDED ROCKS OF BYERS PENINSULA

Hobbs (1968)	*Valenzuela and Hervé (1971)*	*This paper*
Younger Volcanic Group (Miocene)	**Younger unit** Continental deposits (age unspecified) **Older unit** (late Tithonian–Barremian) Marine deposits Continental deposits	

A. North side of New Plymouth.
B. 1.5 km south-east of Richards Cove.

Thirty-seven K-Ar determinations on 18 whole-rock samples and one mineral separate were made from intrusions and lavas on Byers Peninsula (Pankhurst and others, 1980). Because of the alteration which the rocks have suffered, most have been updated, but insofar as none of the dates exceeds the stratigraphical age suggested, they agree with fossil data from the sedimentary rocks. Pankhurst and others placed considerable reliance on dates of about 126 Ma (? Valanginian) obtained from a dyke and a (?) lava in the agglomerates of Ray Promontory, and they suggested a close correlation with the sedimentary and volcanic rocks of Byers Peninsula. This confirms the conclusion reached by the present authors on the basis of field mapping.

BYERS FORMATION

The Byers Formation forms the bedrock to the whole of Byers Peninsula and Elephant Point (Fig. 1). It consists of a mixed sedimentary and volcanic succession, which is mainly marine in the west, terrestrial in the east and has a zone of mixed facies between the two (Table I). Ammonites and belemnites in the marine beds indicate an age range of at least (?) Lower Tithonian to Valanginian, whereas a Lower Cretaceous flora has been obtained from one locality (Negro Hill) in the terrestrial sequence. An overall structural dip to the east suggests that the terrestrial sequence is at least partly younger than the marine sequence and that the youngest exposed parts of the formation are represented by the rocks in the Elephant Point area. No stratigraphical base or top to the formation is exposed. Four main members are defined here on lithological grounds, although, because of poor exposure and intertonguing facies, their boundaries are not always easy to define:

i. *Mudstone member:* deep-water flysch-like sequence of mudstones and thin graded sandstones, probably passing up into (ii).
ii. *Mixed marine member:* a mixed succession of fossiliferous shales and volcaniclastic sandstones and pebble-conglomerates. May include some intertonguing beds of (iii), of which it is a part lateral equivalent.
iii. *Volcanic member:* almost entirely terrestrial pyroclastic rocks with interbedded lavas of basaltic andesite composition.
iv. *Agglomerate member:* partly a lateral equivalent of (iii); locally fossiliferous; rests on (i) and (ii).

A measured thickness of about 1 600 m across the southern part of the peninsula was claimed by Valenzuela and Hervé (1971). However, poor discontinuous exposure, repeated faulting of

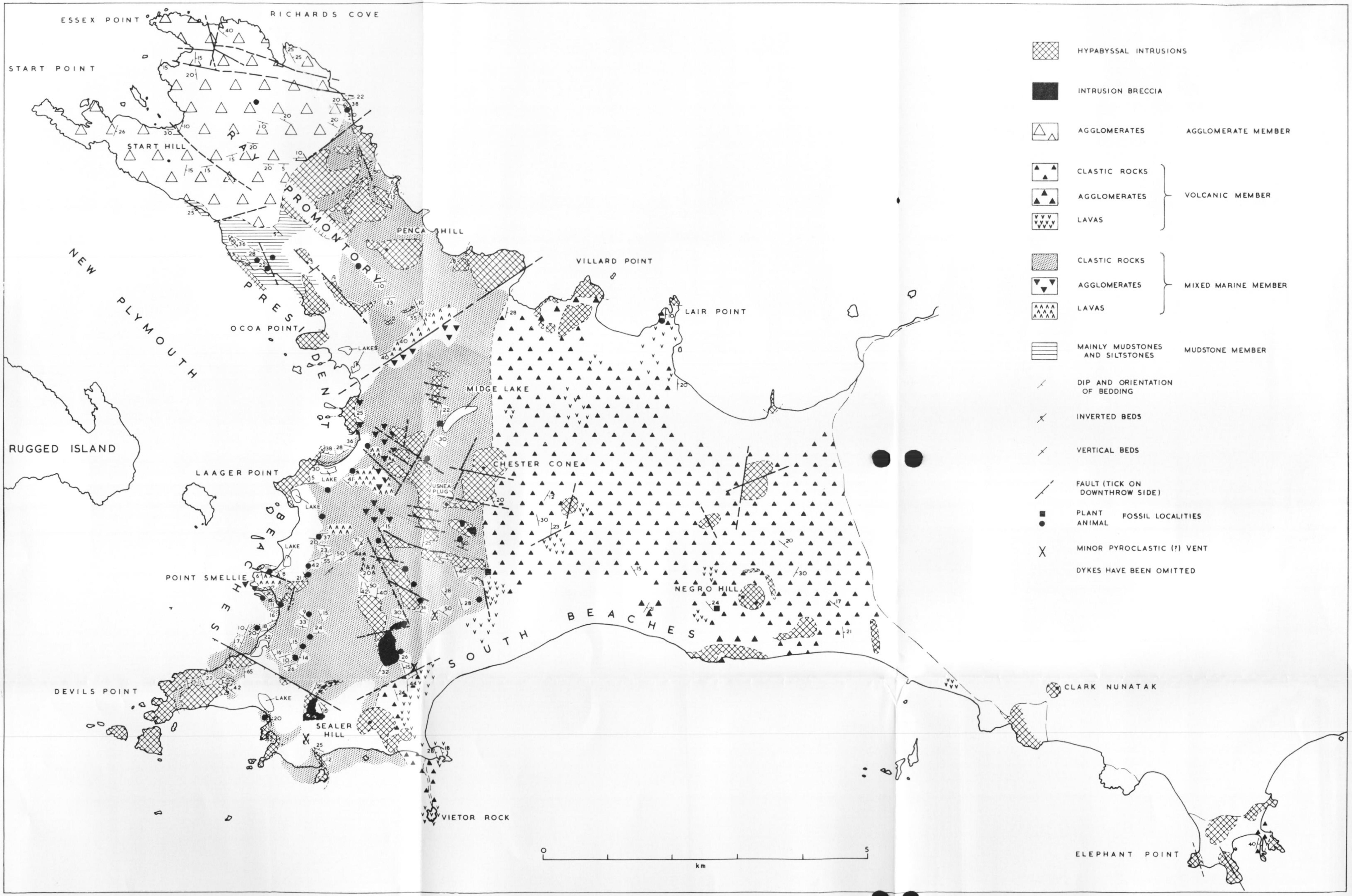


Fig. 2. Simplified geological sketch map of Byers Peninsula.

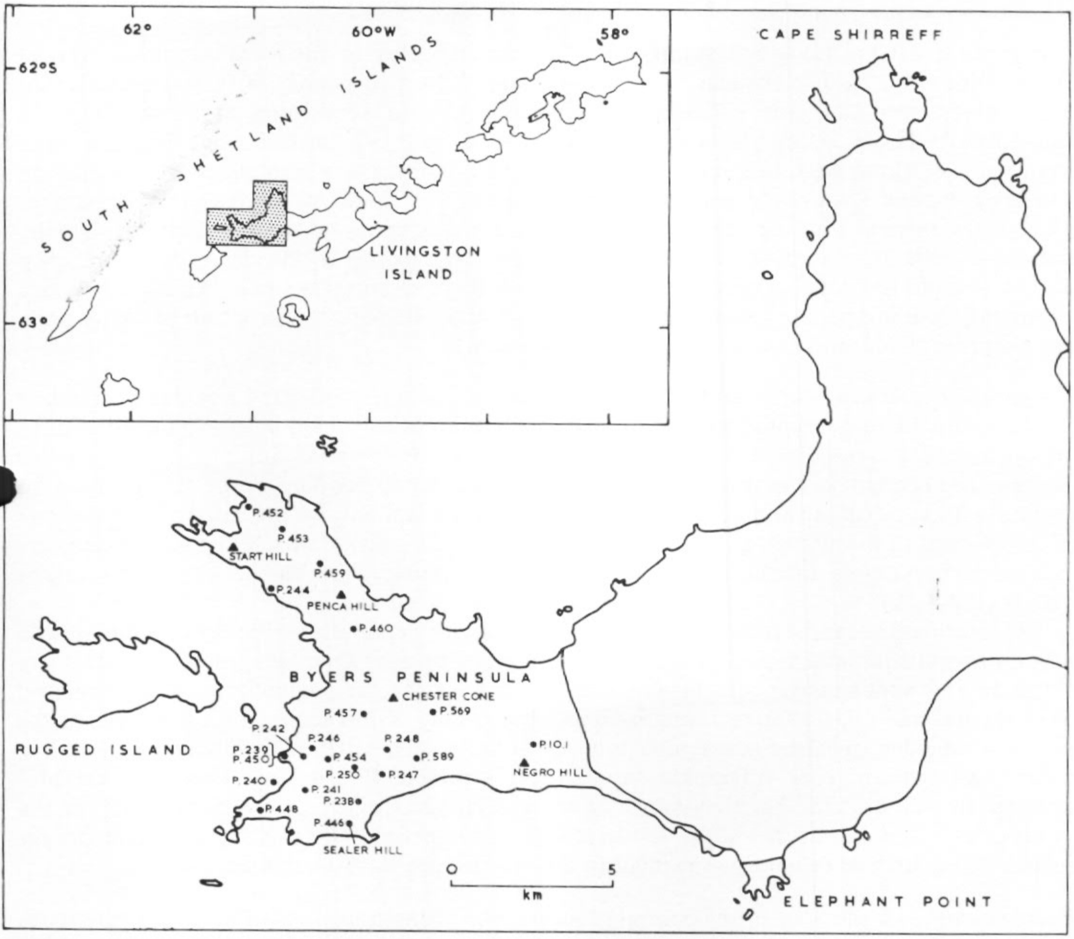


Fig. 1. Sketch maps to show the locations of Byers Peninsula and Cape Shirreff in the South Shetland Islands. Station numbers used in the text are also indicated.

uncertain magnitude and direction, the interruption of the sedimentary sequence by intrusions, and the recognition of a lateral grading from marine to "continental" deposits make their diagrammatic section (Valenzuela and Hervé, 1971, fig. 2) representative rather than real. Consequently, no thickness measurements are given here. The agglomerate member must be at least as thick as the height of Start Hill (268 m) and the figure of about 1 600 m for the remaining members, given by Valenzuela and Hervé (1971), is not an unreasonably high one. Measurements from the geological map suggest that it could be greater but, in view of the unknown structural complications, it would be better to look to a minimum figure for the present.

The mixed marine member, in particular, is intruded by many hypabyssal sheets (Fig. 2). Although these are considered separately here, they show features which suggest that they were intruded before the sediments were properly consolidated and still contained much interstitial water. They were thus more or less penecontemporaneous with the sedimentation and were probably closely allied to the eruptive volcanic activity of the time.

Mudstone member

Exposures of the mudstone member are largely restricted to the foreshore north-west of Ocoa Point (Fig. 2). The sequence is flysch-like and is formed mainly of poorly fossiliferous black mudstones and grey siltstones with occasional thin sandstones and tuffs (Fig. 3). Individual beds are laterally continuous and may be only 1–2 cm thick but they are more commonly 5–10 cm and some tuffs are up to 40 cm. Sedimentary structures are uncommon and they include small-scale contemporaneous faults and flame structures. The sandstones often show normal grading and a few tuffs contain fragments of black mudstone. Oblate, sometimes fossiliferous, rust-stained calcareous concretions, up to 20 cm in diameter, crop out at several horizons, sometimes coalescing to form a complete bed. Trace fossils are abundant, bedding-planar *Chondrites* in particular, and extensive bioturbation of many units has occurred. Compressed wood fragments also occur.

Petrography. The mudstones and siltstones contain silt to fine sand-sized angular plagioclase grains scattered in a cryptocrystalline matrix, which includes much calcite, micro-granular opaque ore, flakes of chlorite or clay and rare cubic crystals of pyrite. Solitary and vermicular burrows are present, orientated parallel and perpendicular to bedding, respectively; they are generally filled by calcite and/or chlorite. Millimetre-scale laminations are caused by variations in the amount of bioturbation and differences in the clast : matrix ratio. Some laminations are rich in carbonaceous matter and contain *Radiolaria* replaced by calcite or polycrystalline quartz (P.244.19).

The interbedded sandstones and tuffs consist mainly of dense pilotaxitic lava fragments and other fine-grained lava types with altered intergranular pyroxene. Plagioclase crystals are common and minor quartz is invariably present; in the tuffs, the plagioclase is often euhedral to subhedral and quartz shows rare lobate margins. Many of the dense (juvenile) clasts in the tuffs are vesicular and there is a notable amount of flattened devitrified pumice; polycrystalline quartz and plutonic rock fragments are also rarely present (P.244.24a). The clasts are predominantly sub-angular but they range from angular (in the tuffs) to sub-rounded (in the sandstones). The sandstones show moderate to good sorting, whereas the tuffs are poorly sorted. No matrix or cement was evident in any of the specimens examined.

Palaeontology. In addition to the common occurrence of the trace fossil *Chondrites*, the mudstone member has also yielded sporadic examples of (?) brachiopods, poorly preserved bivalves, belemnites and ammonites. Two species of belemnite are present. The first closely resembles *Belemnopsis stoleyi* Stevens (1965*b*) from the (?) Kimmeridgian of Indonesia and displays the distinctive, secondarily deepened median ventral groove of that species. The second is closely related to *Hibolithes marwicki marwicki* Stevens (1965*a*) from the Puaroan (Lower Tithonian) of New Zealand. The ammonites include indeterminate oppeliids and fragments of several perisphinctids. Most of the latter are too poorly preserved to identify with any confidence but at least two can probably be referred to the Middle Kimmeridgian–Lower Tithonian genus *Subplanites* Spath. They show early stages with bifurcate ribbing, changing to prominent virgatotome ribbing on the body chamber of the adult, features which are typical of the genus. A single arcuate whorl fragment has the sharp prominent bifurcate ribbing seen on some Argentine forms referred to Burckhardt's (1912) *Berriasella behrendseni*. Although the presence of this species would indicate an Upper Tithonian age, the occurrence of *H. m. marwicki* would favour a Lower Tithonian age, whereas the forms closely related to *B. stoleyi* and *Subplanites* could even favour a Kimmeridgian age. On balance, an early Tithonian age is suggested for these beds.

According to Covacevich (1976, p. 27, caption to fig. 1), Tavera (1970) identified *Berriasella* (*?*) sp., *Pseudolissoceras* (*?*) sp. and *Belemnites* (*Hibolithes*) *jaculum* (Phillips) Feruglio [*sic*] from this locality, but none of these was illustrated or described by Tavera, only listed. Stevens

(1965*a*, footnote on p. 159) has pointed out that Feruglio's (1936) *Hibolithes jaculum* "may be similar to the stouter forms of *H. compressus* and *H. m. marwicki*", thus possibly bringing Tavera's interpretation of the fauna closer to that suggested here.

Mixed marine member

This member forms much of Byers Peninsula west of a north–south line through Chester Cone (Fig. 2). It is lithologically transitional with the volcanic member to the east and intertongues with it. Because of a high shale content, the rocks weather easily and exposures are generally small and scattered. Exposure is particularly poor in the area between Penca Hill and Chester Cone, and the best albeit limited exposures occur in the south-western part of the peninsula. Several lithologically distinctive sequences are present but in the field it was generally difficult to establish their outcrop extent. Nevertheless, there is a general tendency for the rocks to become coarser-grained and younger in an easterly direction.

Dark grey or greenish shales, poorly lithified and interbedded with green and grey sandstone (some with angular clasts of shale), crop out in a large area of very poor exposure surrounding Penca Hill and in parts of the low coastal escarpments on President Beaches (Fig. 2). Calcite beef is common on joint surfaces in certain areas and bioturbation is locally present. This shale-dominated sequence passes up into thin-bedded fine-grained green sandstones, grey siltstones, brown carbonaceous mudstones and a few coarse sandstones, which are well exposed on the foreshore south of Point Smellie; thin conglomerates occur in a similar sequence on the foreshore north-east of Laager Point. Sedimentary structures include thin, often wispy parallel and cross laminations, normal grading, load structures and rare ripple marks.

In the area south and immediately west of Chester Cone, the mudstone–siltstone–sandstone

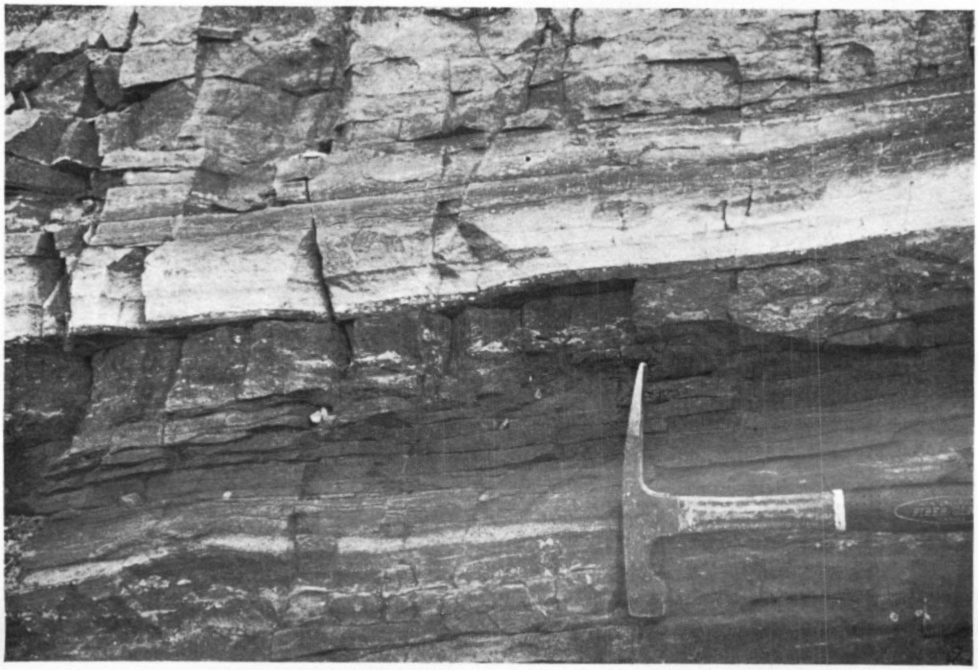


Fig. 3. A flysch-like sequence of mudstones, siltstones and sandstones comprising the mudstone member on the foreshore 1.25 km north-north-west of Ocoa Point. Normal grading can be seen in the pale grey sandstone layers. The hammer head is 18 cm in length.

sequence passes up into coarse sandstones and granule- to pebble-conglomerates. They are grey-green and rust-coloured, friable and moderately to poorly sorted. Calcite cement is ubiquitous, although a few conglomerates have minor amounts of fine-grained matrix as well. Beds of several metres thickness, with scattered blocks up to 30 cm in diameter, crop out in the eastern part of the area but they are more commonly 20–30 cm thick. Sub-rounded to angular pebbles up to 5 cm in diameter are scattered through the beds and there are thin pebble horizons which are frequently the only guide to bedding. Buff-coloured siltstones and mudstones occur rarely, forming beds seldom thicker than 10 cm; lithologically similar clasts with parallel lamination occur in some conglomerates. In granule-conglomerates and sandstones south of Chester Cone, interbedded buff-green tuffs contain accretionary lapilli. Sedimentary structures other than normal grading are absent.

Within the sandstone–conglomerate sequence there are areas of fine-grained rocks, the thickness and outcrop extent of which are poorly known. Approximately 1 km north of Sealer Hill, at station P.238 (Fig. 2), unfossiliferous buff-green to brown papery shales are thickly developed. They are interbedded with minor sandstones and thin, pale grey, sparsely fossiliferous micritic limestones. Thicker bedded black shales with a sparse fossil fauna occur a little lower in the sequence. In other places (e.g. P.248), well-laminated sandstones interbedded with brown shale pass into black shales with large ovoid to tabular calcareous nodules up to 2 m across; smaller septarian nodules are sometimes present. At the southern end of station P.241 (Fig. 2), pyrite nodules were observed in coarse-grained flaggy sandstones below a shale sequence without calcareous nodules.

Lavas and interbedded pyroclastic rocks form isolated outcrops within the marine sequence. In the best example, at Point Smellie, coarse agglomerate grades eastward into conglomerate and is overlain by a totally autobrecciated lava, followed by non-brecciated lavas interbedded with marine shales and siltstones. The headland is capped by lavas, interbedded with tuffs and conglomerate lenses. Small isolated outcrops of non-stratified dark brown lapilli-tuff, which have no analogous lithologies nearby, probably represent minor short-lived vents (Fig. 2).

Petrography. The sandstones, which dominate the mixed marine member, contain sub-angular (less often angular or sub-rounded) lithic and crystal clasts, almost entirely of volcanic derivation. Lithic clasts are commonest and are mainly dense vitrophyric lava types, sometimes porphyritic (plagioclase and an altered ferromagnesian mineral) and/or vesicular. More coarsely crystalline pilotaxitic lava fragments and flattened pumice are less common and there are very rare clasts of acidic plutonic rocks and polycrystalline quartz (P.239.4 and 246.4). The crystal clasts, which locally dominate over lithic clasts (e.g. P.454.4), are mainly of plagioclase but minor quartz, rarely showing marked strain extinction (P.246.4), is invariably present and subhedral apatite occurs in specimens P.246.4 and 454.4. Some plagioclase is subhedral and quartz occasionally has a rounded and/or lobate shape. Carbonaceous matter and fragments of bivalves may also be present (e.g. P.239.47 and 240.4). The sandstones are usually cemented by blocky calcite but a (?) clay fringing cement occurs in some specimens (P.240.4 and 242.18). Vesicular fragments are more common in the conglomerates and the clasts show a greater degree of rounding; they are poorly sorted but are otherwise similar to the sandstones. Clay matrix may be present (P.247.2) but drusy (P.448.6) or blocky calcite cement is more typical of these rocks.

The limestones are micritic with minor amounts of silt-sized angular quartz and feldspar, opaque ore and traces of carbonaceous matter; fragments of (?) bivalves are also present. The "doggers" are also largely formed of micrite with matrix-supported pebbles of lavas (including rare epidotized types) and angular pumice.

The lavas, which crop out in the marine member, are all basaltic andesites with abundant phenocrysts of twinned and zoned andesine, labradorite and olivine, accompanied by less common microphenocrysts of augite and opaque ore. Many of the plagioclase phenocrysts show "finger-print" textures consisting of blebs of dark brown glass. The crystalline groundmass is formed of andesine and labradorite laths with abundant intergranular augite and opaque ore; pigeonite is sometimes present (e.g. P.446.7a and 450.7).

Alteration in these rocks consists mainly of widespread clay growth (including (?) smectite) in plagioclase and replacing vitric material; partial replacement of plagioclase by zeolite has also occurred and is common in the clastic rocks. Olivine is invariably pseudomorphed by serpentine, bowlingite, iddingsite or clay minerals, or more rarely calcite and silica.

Palaeontology. The mixed marine member represents the most fossiliferous part of the Byers Formation and its fossils have already been the subject of study by Chilean workers (González-Ferrán and others, 1970; Tavera, 1970; Covacevich, 1976). Fossils occur in all the sedimentary rock types encountered, although not at every exposure; interestingly enough, the best ammonites are found in the coarser clastic rocks, rather than in the shales. In addition to a mainly molluscan fauna, there are fragmentary representatives of several other groups, including brachiopods, echinoids, ophiuroids, crustaceans, serpulids and fish. The molluscan faunas are dominated by ammonites but bivalves are locally abundant, and rare gastropods and belemnites were also found.

The areal and stratigraphical distribution of the faunas is difficult to assess because of poor exposure and peculiarities of the rocks themselves. In particular, the shales appear to be poorly fossiliferous but this impression may be partly due to their highly friable nature, which causes them to weather rapidly and makes them difficult to collect from. Where the shales have been secondarily indurated in close proximity to intrusions, fossils are more readily obtained.

At least two main faunas can be recognized. The older of the two is that first reported by González-Ferrán and others (1970) and described by Tavera (1970). It occurs principally in the area of President Beaches between Laager Point and Devils Point (Fig. 2), and is characterized by the ammonite genus *Spiticeras*. It also includes *Himalayites hyphasis* (Blanford), *Himalayites* sp. nov. (*?*) *Blanfordiceras* sp. juv. (*?*) and *Bochianites* sp.; no belemnites have been recovered from this part of the sequence. The authors' collections suggest that two species of *Spiticeras* are present, one of which has relatively inflated whorls, whereas the other has a more compressed shell. Although Tavera (1970) compared his two fragments to *S.* (*Spiticeras*) *spitiensis* (Blanford) and *S.* (*Kilianiceras*) *damesi* (Steuer), they are poor specimens (especially the latter) and the identifications are doubtful. Identification to the sub-generic level is not normally possible in the spiticeratids without the inner whorls. Nevertheless, with the exception of a few early forms in the Tithonian, *Spiticeras s.l.* is a typically Berriasian genus. The presence of *Himalayites*, normally regarded as Tithonian, suggests that the beds could range from the earliest part of the Berriasian, although the genus is believed to range into the Berriasian in Alexander Island (Thomson, 1974).

The suggestion that the sequence ranges as high as late Barremian (Tavera, 1970) cannot be upheld as his deduction was based on an extremely poor fragment, identified as *Favrella* cf. *steinmanni* (Tavera, 1970, pl. 3, fig. 9), but which is barely recognizable as ammonoid.

The second main fauna occurs in an angular granule-conglomerate overlying finely bedded shales with interbedded micrite limestones, approximately 1 km north of Sealer Hill (P.238). It has been described in detail by Covacevich (1976), who identified the following ammonite species: *Bochianites* aff. *glaber* Kitchin, *B.* aff. *gerardi* (Stoliczka), *Bochianites* sp. A, *Uhligites* sp. nov. (*?*), *Neocomites neocomiensis* aff. *premolica* Sayn and *Neocomites* sp. ind., in addition to a number of small bivalves and a scaphopod. Largely on the basis of *N. neocomiensis* aff. *premolica*, Covacevich assigned a Valanginian age to the fauna. Similar epiclastic rocks occur as dogger-like concretions in the top of the shales immediately beneath the ammonite bed. These contain robust cylindrical belemnite guards with broad, deep median ventral grooves, identified by Dr L. E. Willey as *Belemnopsis alexandri*, first described from the Berriasian of Alexander Island (Willey, 1973, p. 37, figs 3a–e).

A fauna which may represent an intermediate stratigraphical position between the two outlined above was recovered from baked shales underlying a coarse-grained diorite intrusion at station P.250 (Fig. 2). More than a dozen ammonite fragments represent a fauna of neocomitids and spiticeratids. Few of these are readily identifiable but the best is a neocomitid with fine radiate bifurcating ribs bunched at umbilical tubercles. Although similar to *Neocomites*, the presence of deep constrictions suggests closer affinities to *Thurmanniceras* Cossmann.

Around the shores of Midge Lake and west of station P.238, silicified logs have weathered out of coarse-grained tuffaceous rocks more typical of the volcanic member (below), whereas lithologically similar rocks, 0.6 km west of Chester Cone and in the escarpment behind South Beaches (P.247), contain rare ammonites.

Volcanic member

This unit crops out principally east of the mixed marine member (Fig. 2) and it extends eastward to Elephant Point, and possibly north-eastward as far as Cape Shirreff (Fig. 1). Since the majority of the rocks are soft, easily eroded epiclastic and volcaniclastic sediments, the relief is generally low with a subdued topography providing only rare poor exposures of the sediments. However, better-exposed outcrops of lavas and intrusions produce topographic highs of up to 100 m. The commonest rock type is lapilli-tuff, followed by mudstones and shales, lavas, ignimbrites and rare agglomerates. Emplaced into this sequence are a number of hypabyssal intrusions.

The volcaniclastic rocks range in size throughout the lapilli size limits and beyond into mudstones and agglomerates. At the coarser end of the scale, the clasts attain a maximum diameter of 1 m in rare agglomerates but they are more usually about 5 cm in diameter. The agglomerates range up to a few metres thick and cannot be traced laterally for more than tens of metres. They crop out in the vicinity of Negro Hill, Villard Point and in the centre of South Beaches. Their composition is similar to that of the lapilli-tuffs, i.e. fragments of lava (usually showing flow structure), pumice, glass, quartz, plagioclase (An_{40-60}) and occasional re-worked volcaniclastic material, typically set in a fine-grained groundmass of glass fragments. Generally the tuffs are coarsest in the area to the north-east of Chester Cone, gradually becoming finer eastward; colours are varied in shades of red, green and black. Exposures along the escarpment west of Negro Hill include mudstones, siltstones and shales with occasional sandy lenses. In at least one locality (Fig. 2) these are plant-bearing and probably represent lacustrine deposits. The only sedimentary structures noted were ripple marks, indicating an east–west current direction. Fine grain-size and a general lack of sedimentary structures suggest a low-energy depositional environment.

Fifteen to twenty lava flows, each 5–10 m thick, crop out in the volcanic member. Many of the lavas are grouped in twos and threes within a 3 km radius of Chester Cone; the remainder occur as isolated flows in the vicinity of Negro Hill (Fig. 2). A feature of several flows is their brecciation into blocks 10–30 cm in diameter with a khaki-coloured calcareous or a cryptocrystalline silica joint filling, possibly as a result of the lava entering shallow water and brecciating. The topography at the time of extrusion was probably low, although small-scale (2 m) variations in the orientation of columnar jointing at the base of at least one lava flow is suggestive of a linear depression or stream channel. It is probable that some of these lavas are roughly synchronous and merely represent different lobes of one lava extrusion.

Petrography. The lavas are fairly fresh basaltic andesites, containing small plagioclase phenocrysts (2–3 mm), usually in a flow-structured matrix. The phenocrystic plagioclase is labradorite (An_{50-70}) with the larger crystals tending to be more calcic than the smaller ones. Groundmass plagioclase compositions, where they can be determined, are usually basic andesine (An_{40-50}). Clinopyroxene (augite and/or pigeonite) is the dominant mafic mineral and it occurs almost entirely in the groundmass. Olivine and/or hypersthene pseudomorphs (about 1 mm long) are rarely present. There is alteration of the phenocrystic plagioclase rims to chlorite but extensive alteration is only developed in the groundmass where it produces zeolites, clay minerals, opaque minerals and alkali-feldspar.

Only one rhyolite flow has been recorded and this crops out 1 km south of Chester Cone. It contains phenocrysts of resorbed quartz, altered plagioclase and alkali-feldspar (2–3 mm in diameter), with subordinate (1.5 mm) partly altered biotite lying in a devitrified groundmass of silica and feldspar. Welded ash-flow tuffs are also rare and are only found near Chester Cone and Negro Hill. They display well-developed fiamme and contain many small (1–2 mm), euhedral and subhedral, roughly flow-aligned andesine (An_{40-50}) crystals. Small subhedral and anhedral pyroxenes are occasionally present; these are generally clinopyroxene but there are also a few altered orthopyroxene crystals. Well-preserved tri-cuspate glass shards occur in a poorly exposed non-welded ash-flow tuff, 1.5 km south-east of Chester Cone. It also contains resorbed quartz phenocrysts, altered plagioclase crystals and lithic clasts of pumice and lava in the partly devitrified glass/clay matrix.

Palaeontology. The volcanic member appears to be almost entirely non-marine and fossil remains are limited to petrified wood and leaf impressions. Although large fragments of trunks and branches occur at many places within the pyroclastic rocks, the most spectacular example seen was about 1 km south of Chester Cone, where a petrified stump (20 cm in diameter) is preserved in growth position. Possible lacustrine sandstones at a single locality, 500 m west of Negro Hill, have yielded a leaf flora dominated by ferns. An initial estimate of its age by Fuenzalida (Araya and Hervé, 1966, p. 66) was "Wealden", whereas in a study of more material by Hernández and Azcárate (1971) it was compared to the Baqueró Formation (Barremian) of Argentina.

Agglomerate member

North-western Ray Promontory (Fig. 2), within a radius of 3.5 km of Essex Point, is an elevated area of irregular relief formed by a thick succession mainly of coarse vent breccias and agglomerates. Rugged Island, 4 km to the south-west, is topographically similar and it may be composed of comparable rocks. The agglomerates are grey-green and brown, and they form unsorted chaotic beds 2–35 m thick. Clasts are angular to sub-angular and rarely sub-rounded. They range up to 3 m in diameter but they are generally 5–20 cm and 10–15 cm in the grey-green (heterolithological) and brown (usually monolithological) agglomerates, respectively. Thinner lapillistone beds are not uncommon, the finer varieties with a clast size of 2–4 mm being typically well sorted. Lavas are infrequent and are often difficult to distinguish from the breccias. They are invariably autobrecciated and locally form thick piles of dark brown monolithological clinker with thin non-brecciated lenses. The poor stratification of the lava piles, gradational bedding surfaces between most of the agglomerates and the lack of eroded surfaces suggest that deposition was rapid. Contemporaneous (?) fumarolic activity is indicated by rare funnel-shaped red-coloured areas which terminate abruptly upwards against overlying agglomerates. Unconformities showing a major change in angle of deposition of beds are present and they may be related to changes in location of the eruptive centres (Fig. 4).

A sharp contact between agglomerates and underlying marine shales and siltstones of the mixed marine member is exposed at the base of east-facing cliffs, 1.5 km south-east of Richards Cove. The contact was interpreted as an unconformity by Valenzuela and Hervé (1971) but no age was assigned to the agglomerates. Despite the remarkable contrasts in grain-size and local channelling at the base of the agglomerates, it is suggested that the unconformity is not of major time-stratigraphical significance, for the following reasons:

Fig. 4. Rocks of the agglomerate member on the northern flank of Start Hill and in exposures to the north. The conspicuous differences in bedding orientation can probably be attributed to a change in the location of the volcanic centre from which the rocks were erupted. Approximately 150 m of agglomerates are shown in the photograph.

1. The contact trends up the cliff face due to the cumulative displacements of many small faults; this gives the *impression* of a discordant contact when the exposure is viewed from a distance.
2. Differences in bedding orientation above and below the contact are small when considering the problems involved in taking measurements within a coarsely bedded volcanic sequence.
3. There is no erosional plane, nor discoloration of the marine shale sequence, to suggest a period of subaerial exposure.
4. No clasts of the underlying shales occur in the basal agglomerate of the overlying succession, suggesting a depositional, rather than erosional, upward progression.
5. Thin ash- and pumice-flow deposits are interbedded with marine shale a few metres below the basal agglomerate, indicating active volcanism immediately prior to deposition of the agglomerates.
6. The occurrence of marine fossils (below) at one locality 0.5 km south-south-west of Richards Cove indicates that marine conditions also existed during the deposition of at least part of the agglomerate succession.

It is likely that the agglomerate member also conformably overlies the mudstone member on the south-west side of Ray Promontory. Bedding orientations are similar on either side of the contact (which is nowhere exposed) and active volcanism during the period of deposition of the mudstone member is indicated by the presence of rare tuffs within the mudstone sequence. The fact that the agglomerates can rest directly both on the mixed marine member and the mudstone member without a marked unconformity is rationalized in the facies interpretation set out in Table I.

Petrography. Clasts in the agglomerates and lapillistones are mostly fragments of the lavas and thus share a common petrology with them. Other fragments include vesicular volcanic glasses, and discrete igneous minerals. In a typical lapillistone, the glass is frequently almost opaque or has a marked red coloration. Zoned labradorite (An_{50-70}) and pyroxene (mostly augite) are the only primary igneous minerals, although less common bowlingite/iddingsite pseudomorphs probably represent original olivine. The cementing matrix may be glass, calcite or various platy and fibrous zeolites; calcite and zeolite are also found as amygdale minerals. Although discrete plagioclase crystals are generally fresh, plagioclase in the lava fragments is often badly altered to zeolite.

Palaeontology. Although shell fragments have been observed in thin section (P.452.2), macrofossils were observed *in situ* in the agglomerates at only one locality (P.453), 0.5 km south-south-west of Richards Cove (Fig. 2). Most of these are fragmentary oyster shells but they also include *Mytilus* (*?*), a *Lithophaga*-like form boring an oyster shell, and a large pleurotomariid gastropod. Unfortunately these fossils are more diagnostic of facies than age and they would appear to represent the remains of an oyster bed, destroyed by a lava which advanced into a shallow bay or estuary. Oysters also occur in the marine sequence 1.5 km east-north-east of Point Smellie but a stratigraphical correlation on such tentative evidence is not warranted at this stage.

INTRUSIONS

Intrusions form practically all the major headlands on the north and west coasts and most of the prominent topographical features inland (Figs 5 and 6). They include sills, discordant sheets, dykes and plugs, with compositions varying from basalt to andesite. The thicker andesite and basaltic andesite intrusions often show variation to diorite or microgabbro (e.g. at Devils Point). Sheets of diorite and microdiorite, up to 20 m thick, form extensive outcrops west and south-west of Chester Cone (Fig. 2). They are intensely altered and are often soft and friable in the hand specimen. This deuteric alteration has also affected the overlying and underlying sediments which form conspicuous piles of pale cream-coloured debris (Fig. 7) flanking the intrusion outcrops. Many of the plugs are columnar jointed, the most spectacular examples being Negro Hill, Penca Hill and Sealer Hill (Fig. 6).

Alternating pairs of dark and pale brown "layers" are a striking feature of several intrusions, mainly sills. No cumulate textures are evident in these rocks, and the use of the term "layer" is for descriptive purposes only. The "layers" are generally 1 m thick and laterally continuous parallel to the planar cooling joints of the intrusions; some are wedge-shaped. The pale layers differ mineralogically from the dark layers only in the presence of much zeolite.

Petrography. The intrusions are petrographically similar to the lavas (p. 62). However, with the exception of the coarsely porphyritic basaltic andesites, which intrude the agglomerate member, and a small number of isolated occurrences elsewhere (e.g. P.460.6 and 7), they are less strikingly porphyritic; they also lack opaque ore phenocrysts, pigeonite and interstitial glass. The basalts contain phenocrysts of olivine, labradorite-bytownite and minor augite in a crystalline groundmass formed of the same minerals and opaque ore (probably magnetite). The andesites are sparsely porphyritic with plagioclase (composition undetermined) and hypersthene phenocrysts and microphenocrysts of augite set in a groundmass of andesine, augite, hypersthene, opaque ore (? magnetite) and rare accessory apatite. In specimen P.446.9b, pseudomorphs composed of granular opaque ore, augite and a clay mineral may represent

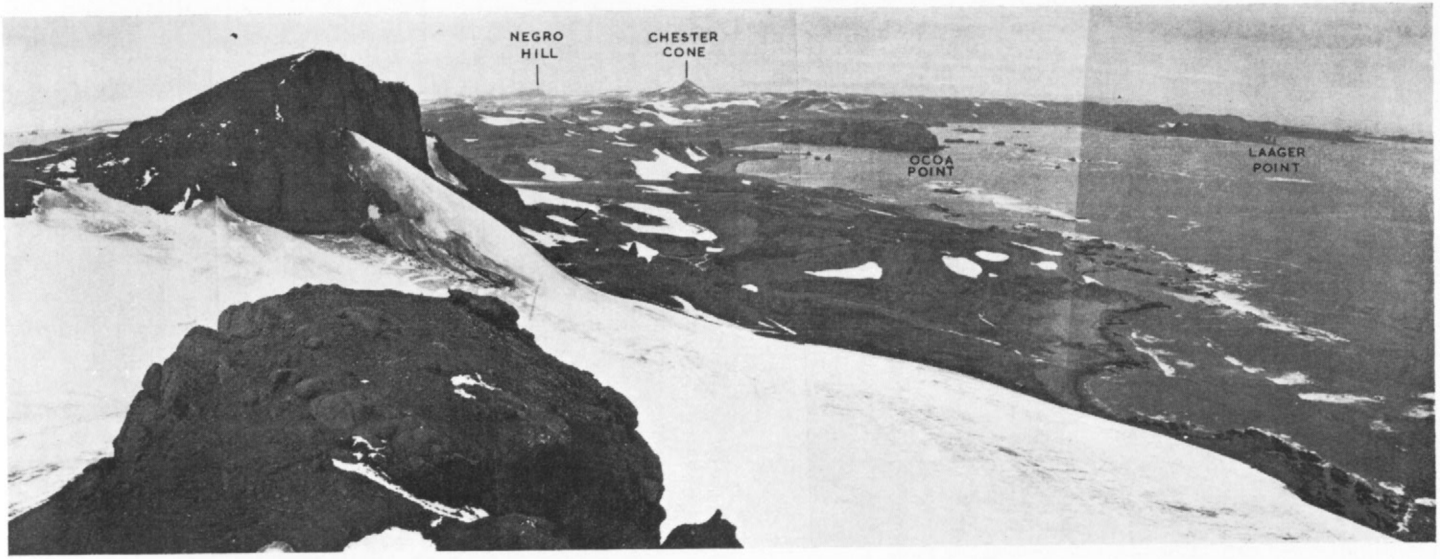


Fig. 5. Panorama of Byers Peninsula from Start Hill (altitude 268 m). The high foreground of Ray Promontory is formed in rocks of the agglomerate member, whereas the evenly dipping beds of the mudstone member are well-exposed along the near foreshore. Across the bay (along and behind President Beaches) are low poorly exposed outcrops of the mixed marine member, with more resistant sills and lavas forming the higher ground and the headlands. The plugs of Chester Cone and Negro Hill project above the irregular surface of the volcanic member. The northern coast of the peninsula is just visible on the left of the photograph. Chester Cone is situated south-east of Start Hill.

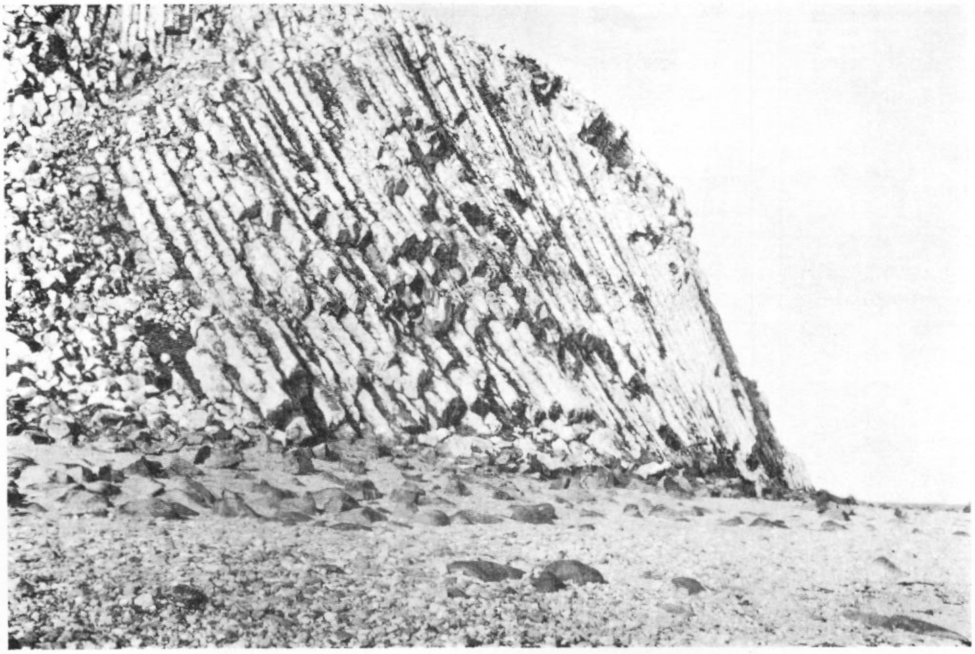


Fig. 6. Coarse columnar jointing in a basaltic intrusion at Sealer Hill. The larger columns are about 2 m wide.

Fig. 7. A dark grey andesite sill capping a small hill 2 km east-north-east of Laager Point. The conspicuous pale-coloured talus slopes are formed by sedimentary rocks zeolitized by the sill.

original hornblende phenocrysts; rare quartz xenocrysts with reaction rims of prismatic augite are also present in this rock.

Chester Cone is a plug formed of hornblende-andesite. It contains phenocrysts of andesine-labradorite, olive-green hornblende, rare augite, (?) hypersthene and opaque ore in a patchy groundmass of plagioclase (composition unknown), quartz, minor (?) hypersthene, opaque ore and accessory apatite. Plagioclase and hornblende megacrysts, up to 2.0 and 10.0 cm in diameter, respectively, and rare hornblende–plagioclase enclaves are scattered through the intrusion but they are particularly common at Usnea Plug. The megacrysts invariably have resorbed outlines. Reaction rims composed of granular (?) augite, opaque ore and plagioclase surround the hornblende and augite phenocrysts.

The microgabbro and microdiorite–diorite intrusions are mineralogically similar to the finer-grained intrusions but they show textural differences, including subophitic augite with oscillatory zoning (P.457.5) and skeletal opaque ore (? ilmenite). Altered olivine occurs in the microgabbros, whereas hypersthene is prominent in the microdiorites and diorites. Phenocrysts are rare.

The alteration of these rocks is closely similar to that shown by the lavas but it is typically more intensive. In particular, the coarser intrusions show considerable development of zeolites (mainly replacing plagioclase), prehnite and chlorite; thin veins of these minerals are not uncommon. Sphene alteration of ilmenite and partial replacement of plagioclase by albite have rarely occurred. Sericite is locally common and some of the thinner dykes and sills are largely replaced by calcite and clay minerals.

Intrusion breccia

One group of intrusions requires a more detailed description than the rest. In the area between Chester Cone and Sealer Hill, and west of Sealer Hill, there are isolated discontinuous outcrops of brecciated, almost clinker-like basaltic rock (Fig. 2). Although they were first mapped by Valenzuela and Hervé (1971) as breccia pipes, Pankhurst and others (1980) later referred to them as basalt/shale breccia, resulting from intrusion into wet sediments. Typically, the breccia consists of angular basaltic fragments, fringed by sparry calcite, and loosely packed (often matrix-supported) in a calcareous micrite matrix (Fig. 8a). The fragments are unsorted and vary in size from sand grade to about 10 cm (occasionally 30 cm) in diameter. In thin section, at least the smaller fragments are highly vesicular and glassy, with fresh plagioclase and augite phenocrysts up to 1.7 and 0.2 mm, respectively. Some contain pyrite clots or are invaded by calcite. Locally, the matrix occurs in irregular pockets up to 20 or 30 cm in extent. Incorporated sand-grade material frequently shows these to be bedded and at times graded. Thin (2–5 cm) dyke-like injections of calcareous micrite, cutting typical matrix-supported breccia, are common. Sometimes they can be traced to bedded pockets from which lateral, downward and upward intrusion can be proved. Although these rocks were studied in some detail in the field, neither fossil trees (Valenzuela and Hervé, 1971) nor shale (Pankhurst and others, 1980) were observed in the breccia. However, discrete clasts of limestone are present; these are generally no more than 3–4 cm in size but blocks, with internal bedding, up to 20 cm across occur ocasionally.

The breccia is everywhere too poorly exposed to map its geological relationships confidently. It is frequently associated with unbrecciated "basalt" at the same level and, where contacts are seen, the basalt is often shatter-cracked for several metres adjacent to the breccia. Although the wider cracks in the shattered rock may contain a micrite infilling, sharp contacts are usually defined between the true breccia and the cracked basalt. Both vertical and horizontal contacts between the two were observed over short distances.

The characteristics of these rocks are consistent with their formation by intrusion into wet sediment, probably at a high level in the sedimentary pile. The intense vitrification and vesicula-

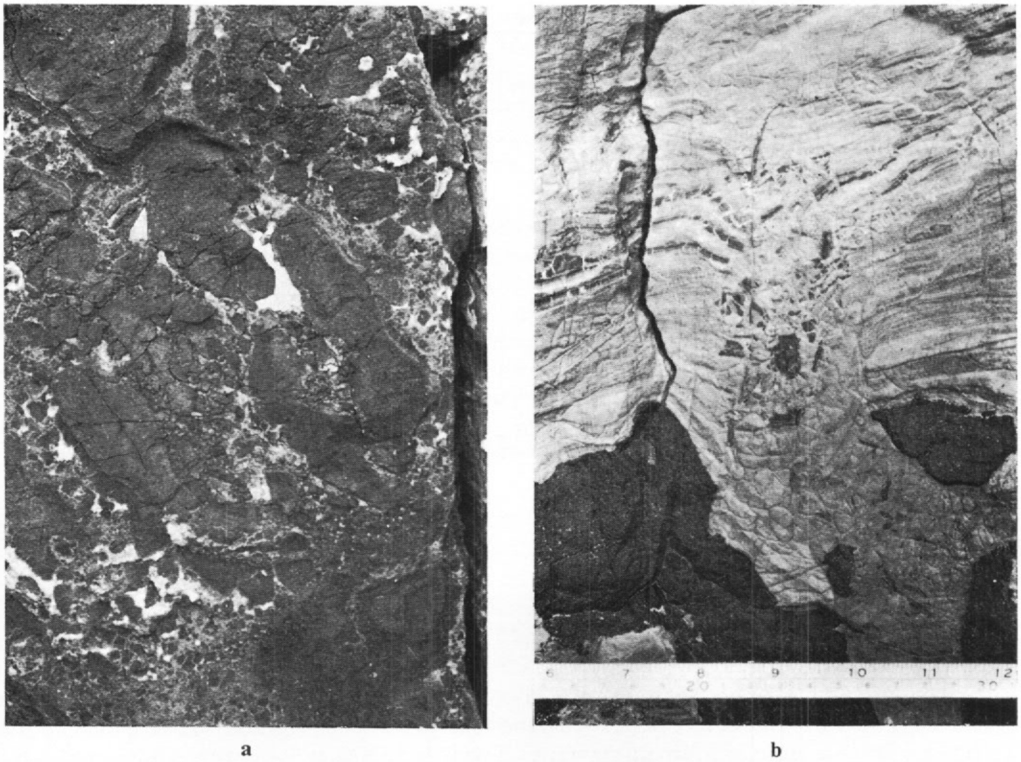


Fig. 8. a. A typical exposure of the breccia intrusion showing its unsorted and highly fragmented appearance. Paler grey areas are highly calcareous interstitial matrix and the white patches are "amygdales" of calcite. Note the cracks in the larger blocks and their incipient fragmentation. The area shown is about 1 m high.

b. A large pocket of calcareous matrix in the breccia intrusion, showing internal bedding disrupted by subsequent re-mobilization in a pipe-like or fissure structure. The scale is graduated in inches (above) and centimetres (below).

tion indicate rapid cooling. Brecciation of the rapidly chilled magma was probably explosive, allowing re-mobilized sediment to filter between dispersed clasts. Re-mobilization was sufficient in some cases to allow re-deposition and the development of bedding in the larger interstitial areas. The presence of micrite dykes cutting the breccia suggest that the process was either prolonged or repetitive. In this respect, an outcrop 1 km north of Sealer Hill (western end of station P.238) is particularly significant (Fig. 8b). There, bedded micrite with included vesicular basalt fragments has been secondarily re-mobilized in a small pipe or fissure, in which chaotically disposed fragments, representing original lithified bands within the micrite, have sunk down into structureless micrite.

ALTERATION

The lavas and clastic rocks show features characteristic of diagenesis and/or very low-grade metamorphism:

i. Unaltered volcanic glass is sometimes present, both in the lavas (e.g. P.569.1) and volcaniclastic rocks (e.g. P.10.1); crystals of plagioclase, augite and biotite are typically unaltered; primary igneous textures are essentially unchanged.

ii. Serpentine, bowlingite, iddingsite and clay minerals (including (?) smectite) are common and may be the only evidence for alteration (e.g. P.589.1); they have formed by alteration of olivine, orthopyroxene, plagioclase and glass; augite is seldom affected. Smectite may form under conditions ranging from diagenesis to the lower greenschist facies (Utada, 1965; *in* Miyashiro, 1973, p. 143); serpentine, bowlingite and iddingsite are common products of deuteric alteration, low-grade metamorphism and weathering.

iii. Zeolites are present in minor amounts, generally replacing interstitial glass (in the lavas), vitric fragments (in the volcaniclastic rocks) or forming along fractures in plagioclase phenocrysts.

iv. The plagioclase phenocrysts and groundmass laths are typically unaltered or contain (?) smectite without the development of albite. Epidote, a principal product of albitization of plagioclase, is also absent. Since albitization is predominant in rocks altered at temperatures above 250–280°C (Eskola and others, 1937; Coombs and others, 1959), a somewhat lower temperature may be postulated for the rocks considered here.

v. With the exception of the New Plymouth mudstones, chlorite is absent. Chlorite begins to form in metamorphic rocks of prehnite–pumpellyite-facies grade; in the zeolite facies, its place is taken by smectite, illite and mixed-layer clay minerals (Miyashiro, 1973, p. 163).

The distinction between diagenesis and metamorphism appears to be rather arbitrary. Miyashiro (1973) included in the zeolite facies all zeolite-bearing rocks altered by burial alone, whereas Zwart and others (1967) defined the commencement of metamorphism as characterized by the occurrence of laumontite with quartz (thus assigning some zeolites to diagenesis). Blatt and others (1972, p. 567) regarded the "laumontite facies" as "the highest sedimentary condition, and the pumpellyite facies [*sic*] as the lowest truly metamorphic facies". Following the scheme used by Miyashiro (1973), the lavas and clastic rocks on Byers Peninsula contain secondary minerals characteristic of diagenesis and very low-grade (zeolite-facies) metamorphism. The presence of (?) metamorphic chlorite in the New Plymouth mudstones suggests that these rocks have attained a metamorphic grade equivalent at least to the prehnite–pumpellyite facies. This is supported by their occurrence in the field as indurated, sometimes flinty rocks in contrast to the friable nature more typical of rocks on Byers Peninsula; several thin sill-like bodies intrude these rocks and may have supplied the heat necessary for their metamorphism.

Alteration in the hypabyssal intrusions is mineralogically similar to that shown by the extrusive rocks but it is more extensive and includes greater amounts of zeolite and clay minerals. Also, prehnite, chlorite, calcite, sericite, albite, sphene and pyrite are locally abundant, which contrasts with the general metamorphic grade of the extrusive rocks. It is thought that the secondary minerals in the hypabyssal intrusions formed principally by deuteric rather than metamorphic processes.

STRUCTURE

Structural data for the Byers Formation are presented in Figs 9a and b, and 10. The broad scatter of points shown in Fig. 9a and b reflect the open, "dome and basin" style of folding observed in the field. However, there is a bias towards east- to east-north-east-dipping orientations, indicating that regional tilting has also occurred; this is particularly well shown by the data for the volcanic member (Fig. 9b). The number of intrusions and the amount of deformation apparently increase from east to west (Fig. 9a and b). However, there is no obvious causal relationship between any of the intrusions and the large-scale structures, and cross-cutting relationships can sometimes be demonstrated (e.g. at Laager Point). Also, the orientation of bedding on Ray Promontory (excluding the agglomerate member; see below) appears to describe a large-scale antiformal structure centred on a locality 1.5 km north-north-west of Penca Hill; no intrusions on a scale similar to this have yet been recognized, although they may exist at depth. Minor kink-like folds, sometimes associated with shearing, occur in bedding

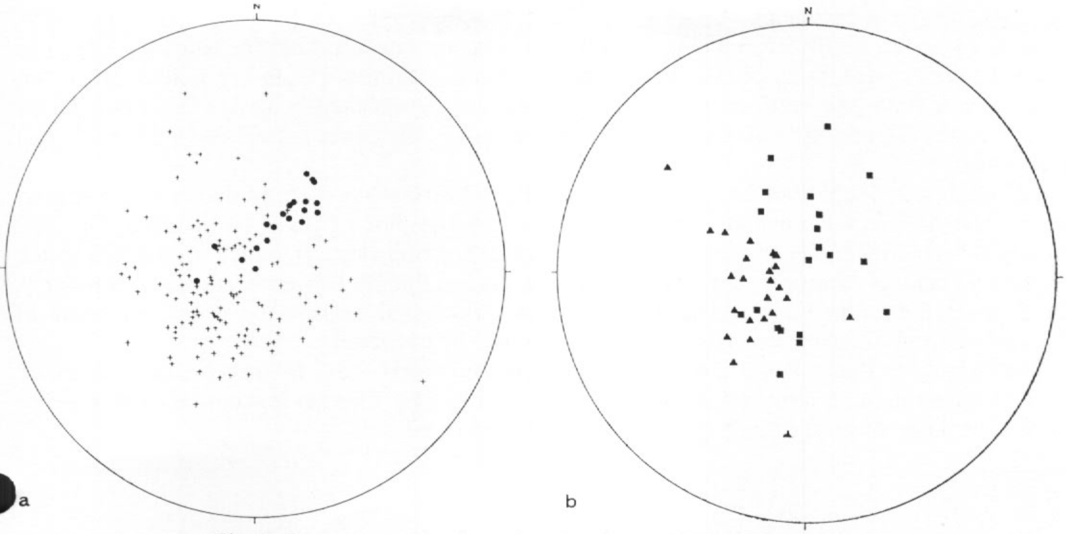


Fig. 9. Stereograms of poles to bedding for the Byers Formation.
a. Mudstone (circles) and mixed marine (crosses) members.
b. Volcanic (triangles) and agglomerate (squares) members.

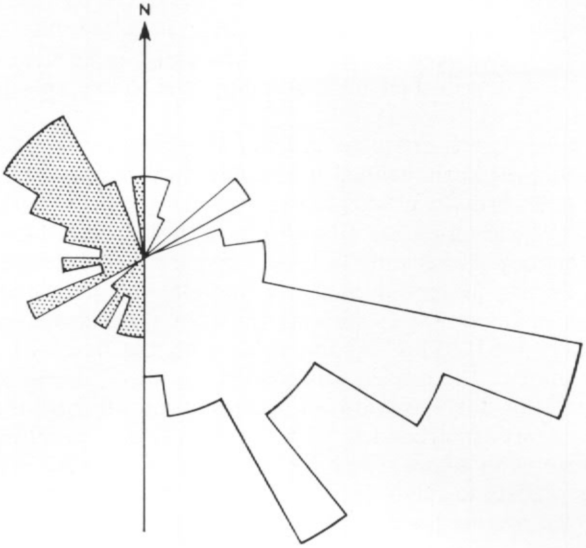


Fig. 10. Rose diagram showing the trends of faults (stippled) and dykes on Byers Peninsula.

adjacent to some intrusions and they probably formed due to local compressive stresses set up during their emplacement.

In gully and stream exposures approximately 1–1.5 km east-north-east of Point Smellie the usual tilted attitude of the bedding is disrupted by a number of tight folds with north-easterly trending axes and small-displacement faults. The folds vary from monoclinal to tight anticlinal, even isoclinal; axial planes are steeply inclined. The fold amplitudes are typically no more than a few metres and bedding on the flanks of folds in shaly lithologies is sometimes crumpled. Poor exposure prevents the structures being traced for any distance but bedding in the rocks in the escarpment inland of President Beaches, north and south of the fold gully, appears to be "normally" tilted (Fig. 2). Lack of cleavage suggests that the folds were formed when the rocks were incompletely lithified and a slump or similar origin is possible. However, they do not have the reclined axes characteristic of the excellently preserved much larger slumps in the Upper Jurassic and Lower Cretaceous of Alexander Island (Bell, 1975; Taylor and others, 1979).

Although it is likely that the agglomerates of Ray Promontory suffered deformation similar to the rest of Byers Peninsula, the scatter shown by the plot of poles to bedding (Fig. 9b) mainly reflects differences in the original angles of deposition (Fig. 4), which are related to the (?) several centres of eruption active in this area during the (?) Lower Cretaceous. Similarly, the data for the mudstone member (Fig. 9a) show a restricted distribution which is a result of the very limited outcrop area from which data could be collected.

The dykes on Byers Peninsula have a predominant north-west to west-north-west trend, which is also the main fault trend for the peninsula (Fig. 10). There is less convincing evidence for a subsidiary dyke and fault trend to the east-north-east.

Discussion

The Byers Formation is interpreted here as a complex intra-arc sequence of intertonguing marine and terrestrial facies, with minor lacustrine developments. All the marine rocks are characterized by an extremely high clastic volcanic content, whereas the terrestrial facies are represented almost entirely by lavas and pyroclastic rocks. On the basis of field mapping, four rock members (mudstone, mixed marine, volcanic and agglomerate; p. 56) of imprecisely defined geographical extent have been recognized. These range in age from (?) early Tithonian to Barremian. As an initial step in their understanding, the following schematic sedimentary model is proposed (Fig. 11a–c).

Mudstone-member times were characterized by the deposition of marine terrigenous sediments, comprising thin regularly bedded mudstones and siltstones with a sporadic nekto-benthonic shelly fauna and normally graded immature sandstones. Bioturbation was high with numerous occurrences of bedding-planar *Chondrites*. Although high bioturbation may occur under diverse sedimentary conditions and *Chondrites* by itself is of little use as a depth indicator (Crimes, 1973, p. 128, fig. 14), it appears that bedding-planar burrows are best developed in deposits of deeper, less agitated waters such as those of the marine shelf (McAlester and Rhoads, 1967; Seilacher, 1967). This is compatible with the flysch-like appearance of the sequence. In this environment, the graded sandstones may have formed as storm sand layers (Reineck and Singh, 1975) or during periods of increased run-off from the source area. Pyroclastic tuffs with vitric clasts interbedded with the mudstones point to contemporaneous volcanic activity. However, this appears to have been located at some distance from the area of deposition and the mudstone member probably accumulated under more open marine conditions than the mixed marine member (Fig. 11a and b).

Despite differences in induration noticeable in the field, it is possible that at least part of the *mixed marine member* in the Penca Hill area (shales with interbedded thin sandstones and granule-conglomerates) represents a near-shore equivalent of the mudstone member. However, exposure in this area is so poor and stratigraphical control so weak that this suggestion can only be tentative at present.

The close of mudstone-member times probably saw the inception of a major volcanic vent in the Start Point area and perhaps smaller vents over the Byers Peninsula area as a whole (Fig. 11b). It is believed that these created a somewhat enclosed marine sedimentary basin into which volcaniclastic material was introduced by re-working and by direct aerial fall-out. Most of the clasts in the mixed marine member are of a similar composition to the rocks of the agglomerate member and are similarly altered; fragments derived from other sources include

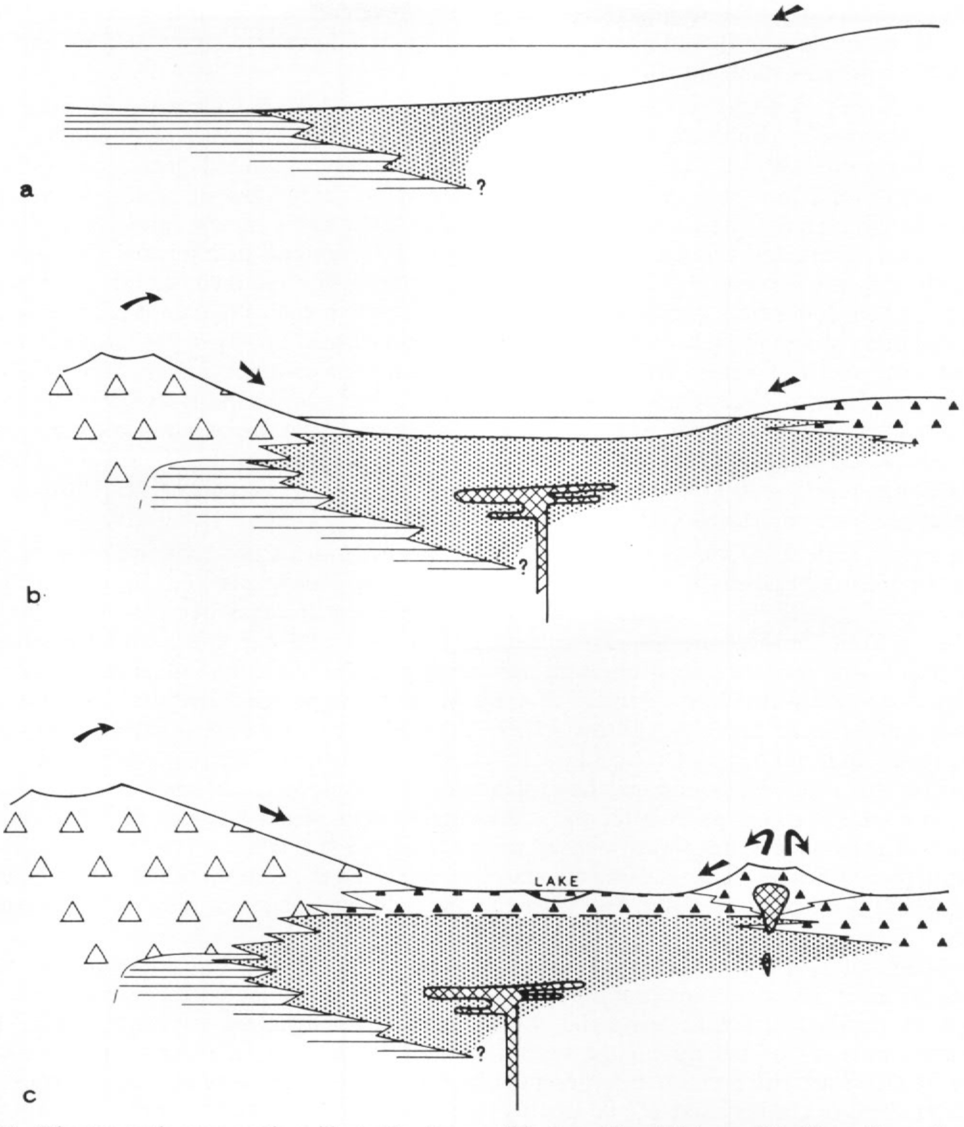


Fig. 11. Diagrammatic cross-sections illustrating the possible depositional history of the Byers Formation and the field relationships between the individual members. The key used is the same as that in Fig. 2.

a. *Mudstone member times.* Deposition was in relatively deep water on a marine shelf with a detrital source area external to the basin. The existence of the mixed marine member at this time lacks field evidence but it is tentatively included to illustrate landward shoaling of the basin and the likely field relationships between the two members.

b. *Mixed marine member times.* Detritus was principally derived from within-basin sources (small vents, represented here by the agglomerate member), although there was still a relatively reduced external supply of sediment. It is likely that the basin was rapidly shoaling.

c. *Volcanic member times.* The basin may have been completely infilled by this time and deposition occurred largely by direct air-fall (pyroclastic) processes. Normal weathering of local vents also supplied detritus, some of which accumulated in shallow lakes. Much of the land may have been forested (not shown).

minor amounts of volcanic quartz and very rare occurrences of epidotized volcanic rocks, plutonic rocks, and polycrystalline and strained quartz. Localized agglomerates and lavas probably represent small short-lived island vents.

The sedimentary rocks of the mixed marine member exhibit a wide range of grain-sizes, from mudstones to pebble-conglomerates, and they represent a number of diverse facies, all of which are probably of shallow-water origin. Horizontally laminated clean sands with thin, often wispy mudstone intercalations resemble sediments laid down in periods of relatively high-velocity currents (Reineck and Singh, 1975, p. 107), and coarsely interlayered beds of thick sandy layers and thinner, finer-grained layers are common in intertidal flats although typically they are strongly bioturbated. Shale sequences within the member represent periods of quiet-water deposition, perhaps within protected embayments. Frequent lack of bioturbation and often sparse fossil content of these rocks could suggest poorly oxygenated conditions. In the area north of Sealer Hill, one such shale sequence contains 2–3 m of virtually unfossiliferous micritic limestones. Lime is a common constituent in the matrices of the clastic rocks and was presumably derived principally by leaching of the surrounding volcanic terrain. The only clastic material in the limestones consists of occasional tiny highly angular grains of (?) volcanic quartz and feldspar. This suggests that the area was receiving very little detrital input at the time of their formation.

On a gross scale, the rocks of the mixed marine member coarsen upwards and grade laterally eastward into the essentially terrestrial *volcanic member*. Valenzuela and Hervé (1971) first recognized a zone of mixed marine/terrestrial facies in a north–south belt extending just west of Chester Cone. In this zone, angular volcaniclastic rocks, similar in appearance to associated terrestrial rocks, contain rare ammonites indicating periods of marine influence. These rocks are typically well washed and contain discrete horizons of pebble-sized clasts suggestive of coastal wave-worked deposits (Clifton, 1973). Although their moderate to poor sorting could alternatively indicate deposition by alluvial processes, a marine influence is undoubted, and it is possible that rapid deposition may have prevented the development of good sorting. Silicified wood is common in the pyroclastic and sedimentary rocks of the volcanic member and this suggests that the land surface was forested; at one locality a tree trunk was observed in growth position (p. 63). Some of the volcanic member was deposited in shallow lakes, the best example being found south-west of Negro Hill where coarse tuffaceous sands contain poorly preserved leaves.

Although part of the volcanic member represents a time facies equivalent to the mixed marine member, its upper parts are probably younger. It is possible that the marine basin was completely infilled and that terrestrial rocks once extended across its surface (Fig. 11c). Probably most of the detritus in the volcanic member was derived from a major volcanic source in the Start Hill area, now represented by the *agglomerate member*. The incorporation of oyster shells in clinkery lava at one locality (p. 65) indicates that at least part of the volcanic pile was deposited under shallow marine, possibly brackish water conditions. However, smaller vents, such as those now represented by the plugs at Chester Cone and Negro Hill were also active and must have made a significant contribution. Interbedded lavas show preferred geographical distributions close to such centres, and there is geochemical evidence that some of the ash-flow tuffs are related to the plug forming Chester Cone (Pankhurst and others, 1980).

In addition to the extrusive expression of volcanic activity, it appears that many of the intrusions are related. Zeolitization (p. 68) and the brecciation of some intrusions (p. 68) in a manner analogous to peperites (cf. Williams, 1929) suggests emplacement at shallow depth into a pile of wet sediments. Some of these may even have locally burst through the sediment cover.

Lithologically comparable sedimentary–volcanic sequences of Upper Jurassic–Lower

Cretaceous age crop out on Isla Hoste (Patagonia) and Annenkov Island (Suárez and Pettigrew, 1976), and at Ducloz Head, South Georgia (Storey, in press). They consist of thin-bedded mudstones and interbedded tuffs deposited on a marine shelf (inland member, Ducloz Head Formation; lower tuff member, Annenkov Island Formation; parts of the Hardy Formation, Isla Hoste), which pass up into sandy shallow-water deposits (basal sandstones of the upper breccia member, Annenkov Island Formation; "Tekenika Beds", Hardy Formation) and coarse structureless volcanic breccias (upper breccia member, Annenkov Island Formation; parts of the Hardy Formation). A silicic unit, consisting of silicic volcaniclastic (predominantly pyroclastic) rocks, probably ignimbrites, tuffs and porphyritic rhyolites, is also included in the Hardy Formation and may be comparable with the volcanic member on Byers Peninsula.

Although grossly comparable, these rocks differ from those on Byers Peninsula in the following respects:

i. The marine shelf deposits of Patagonia and South Georgia contain much vitroclastic detritus which is absent from the mudstone member and the shales surrounding Penca Hill, to which they are otherwise similar. This difference is probably attributable to a more distant source area for the marine shelf deposits on Byers Peninsula.

ii. Tuffaceous rocks in the agglomerate member on Byers Peninsula contain abundant angular glassy fragments and they are here interpreted as primary pyroclastic ("ash-fall") deposits. As such, they probably accumulated in close proximity to an active vent. By contrast, the structureless breccias on Annenkov Island (which have a sandy matrix from which vitroclastic material is *absent*) were transported further by slumping and other mass-flow processes into the adjacent sedimentary basins (Suárez and Pettigrew, 1976).

These differences, however, probably reflect only local variations in depositional environment, and all the sequences discussed here may be regarded as those of a Mesozoic arc which was situated in eastern Patagonia and the Antarctic Peninsula (Scotia arc *s.l.*) (Suárez, 1976; Suárez and Pettigrew, 1976; Thomson, in press).

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REFERENCES

ARAYA, R. *and* F. HERVÉ. 1966. Estudio geomorfológico y geológico en las Islas Shetland del Sur, Antártica. *Publnes Inst. antárt. chil.*, No. 8, 1–76.

BARTON, C. M. 1965. The geology of the South Shetland Islands: III. The stratigraphy of King George Island. *British Antarctic Survey Scientific Reports*, No. 44, 33 pp.

BELL, C. M. 1975. Structural geology of parts of Alexander Island. *British Antarctic Survey Bulletin*, Nos. 41 and 42, 43–58.

BLATT, H., MIDDLETON, G. *and* R. MURRAY. 1972. *Origin of sedimentary rocks*. New Jersey, Prentice-Hall Incorporated.

BURCKHARDT, C. 1912. Faunes jurassiques et crétaciques de San Pedro de Gallo. *Boln Inst. Geol. Méx.*, No. 29, 264 pp.

CLIFTON, H. E. 1973. Pebble segregation and bed lenticularity in wave-worked versus alluvial gravel. *Sedimentology*, **20**, No. 2, 173–88.

COOMBS, D. S., ELLIS, A. J., FYFE, W. S. *and* A. M. TAYLOR. 1959. The zeolite facies, with comments on the interpretation of hydrothermal syntheses. *Geochim. cosmochim. Acta*, **17**, Nos. 1/2, 53–107.

COVACEVICH, C., V. 1976. Fauna Valanginiana de Península Byers, Isla Livingston, Antártica. *Revta geol. Chile*, No. 3, 25–56.

Crimes, T. P. 1973. From limestones to distal turbidites: a facies and trace fossil analysis in the Zumaya flysch (Paleocene–Eocene), north Spain. *Sedimentology*, **20,** No. 1, 105–31.

Eskola, P., Vuoristo, U. *and* K. Rankama. 1937. An experimental illustration of the spilite reaction. *Bull. Commn géol. Finl.*, No. 119, 61–68.

Feruglio, E. 1936. Palaeontographia patagonica. Pt. 1. *Memorie Ist. geol. miner. Univ. Padova*, **11,** 1–192.

González-Ferrán, O., Katsui, Y. *and* J. Tavera. 1970. Contribución al conocimiento geológico de la Península Byers de la Isla Livingston; Islas Shetland del Sur, Antártica. *Ser. cient. Inst. antárt. chileno*, **1,** No. 1, 41–54. [*Contrnes Inst. antárt. chileno*, No. 19.]

Hernández, P. J. *and* V. Azcárate. 1971. Estudio paleobotánico preliminar sobre restos de una tafoflora de la Península Byers (Cerro Negro), Isla Livingston, Islas Shetland del Sur, Antártica. *Ser. cient. Inst. antárt. chileno*, **2,** No. 1, 15–50. [*Contrnes Inst. antárt. chileno*, No. 25.]

Hobbs, G. J. 1968. The geology of the South Shetland Islands: IV. The geology of Livingston Island. *British Antarctic Survey Scientific Reports*, No. 47, 34 pp.

McAlester, A. L. *and* D. C. Rhoads. 1967. Bivalves as bathymetric indicators. *Mar. Geol.*, **5,** Nos. 5 and 6, 383–88.

Miers, J. 1820. Account of the discovery of New South Shetland, with geographical observations on its importance in a geographical, commercial and political point of view. *Edinb. phil. J.*, **9,** Pt. 4, No. 6, 367–80.

Miyashiro, A. 1973. *Metamorphism and metamorphic belts.* London, George Allen and Unwin Ltd.

Pankhurst, R. J., Weaver, S. D., Brook, M. *and* A. D. Saunders. 1980. K-Ar chronology of Byers Peninsula, Livingston Island, South Shetland Islands. *British Antarctic Survey Bulletin*, No. 49, 277–82.

Reineck, H. E. *and* I. B. Singh. 1975. *Depositional sedimentary environments.* Berlin, Heidelberg, New York, Springer-Verlag.

Seilacher, A. 1967. Bathymetry of trace fossils. *Mar. Geol.*, **5,** Nos. 5 and 6, 413–28.

Stevens, G. R. 1965*a*. The Jurassic and Cretaceous belemnites of New Zealand and a review of the Jurassic and Cretaceous belemnites of the Indo-Pacific region. *Palaeont. Bull., Wellington*, No. 36, 283 pp.

———. 1965*b*. A new belemnite from the Upper Jurassic of Indonesia. *Palaeontology*, **7,** Pt. 4, 621–29.

Storey, B. C. In press. The geology of the Ducloz Head area of South Georgia. *British Antarctic Survey Bulletin.*

Suárez, M. 1976. Plate-tectonic model for southern Antarctic Peninsula and its relation to southern Andes. *Geology*, **4,** No. 4, 211–14.

———. *and* T. H. Pettigrew. 1976. An Upper Mesozoic island-arc–back-arc system in the southern Andes and South Georgia. *Geol. Mag.*, **113,** No. 4, 305–28.

Tavera, J., J. 1970. Fauna Titoniana–Neocomiana de la Isla Livingston, Islas Shetland del Sur, Antártica. *Ser. cient. Inst. antárt. chileno*, **1,** No. 2, 175–86. [*Contrnes Inst. antárt. chileno*, No. 23.]

Taylor, B. J., Thomson, M. R. A. *and* L. E. Willey. 1979. The geology of the Ablation Point–Keystone Cliffs area, Alexander Island. *British Antarctic Survey Scientific Reports*, No. 82, 65 pp.

Thomson, M. R. A. 1974. Ammonite faunas of the Lower Cretaceous of south-eastern Alexander Island. *British Antarctic Survey Scientific Reports*, No. 80, 44 pp.

———. In press. Mesozoic palaeogeography of western Antarctica. (*In Third Symposium on Antarctic Geology and Geophysics. Madison, Wisconsin, U.S.A., 22–27 August 1977.*)

Valenzuela, E. *and* F. Hervé. 1971. Geology of Byers Peninsula, Livingston Island, South Shetland Islands. (*In* Adie, R. J., *ed. Antarctic geology and geophysics.* Oslo, Universitetsforlaget, 83–89.)

Willey, L. E. 1973. Belemnites from south-eastern Alexander Island: II. The occurrence of the family Belemnopseidae in the Upper Jurassic and Lower Cretaceous. *British Antarctic Survey Bulletin*, No. 36, 33–59.

Williams, H. 1929. The geology of the Marysville Buttes, California. *Univ. Calif. Publs geol. Sci.*, **18,** No. 5, 103–220.

Zwart, H. J., Corvalan, J., James, H. L., Miyashiro, A., Saggerson, E. P., Sobolev, V. S., Subramaniam, A. P. *and* T. G. Vallance. 1967. A scheme of metamorphic facies for the cartographic representation of regional metamorphic belts. *Circ. Lett. Int. Un. geol. Sci.*, No. 2, 57–72.