Caledonian Igneous Rocks of Great Britain

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Chapter 2

Early Ordovician volcanic rocks and associated ophiolitic assemblages of Scotland

INTRODUCTION

P. Stone and D. Flinn

At the beginning of the Palaeozoic Era the continental masses of Laurentia, Avalonia and Baltica were separated by large oceans. The Iapetus Ocean separated Laurentia from Eastern Avalonia and Baltica while the two latter continents were divided by the Tornquist Sea. Volcanic activity occurred within and at the margins of these oceans. The rocks described in this chapter all originated as varieties of Iapetus oceanic crust and upper mantle which were caught up in the orogenic fold belt produced as the ocean closed. This is the Caledonian–Appalachian Orogen which, prior to the Mesozoic and later opening of the Atlantic Ocean, formed a sinuous deformation belt extending from what is now northern Scandinavia, through Britain and Ireland, and on into maritime Canada, the eastern seaboard of the USA and possibly beyond. A pre-Atlantic continental reconstruction is shown in Figure 2.1.

The Iapetus Ocean probably reached its maximum width during the Late Cambrian or early Ordovician. Closure of the ocean then followed during the Ordovician and Silurian with continental suturing complete by the Early Devonian. Along the length of the orogen are preserved vestiges of the original ocean in the form of ophiolite complexes, assemblages variously comprising spilitic lava, sheeted mafic dykes, gabbros and ultramafic plutonic rocks; the ophiolite concept is discussed in more detail below. The distribution of the principal ophiolite complexes within the orogen is shown in Figure 2.1. Those now seen in Britain and Canada were obducted onto the continental margin of Laurentia, the Norwegian examples were obducted onto the margin of Baltica on the other side of the closing ocean. Despite their wide geographical range there is a surprising uniformity of age with most apparently being generated in the Late Cambrian or early Ordovician and then obducted shortly after. Of the Scottish ophiolites, the early Ordovician, Ballantrae Complex has much in common with those now exposed in Newfoundland, whereas

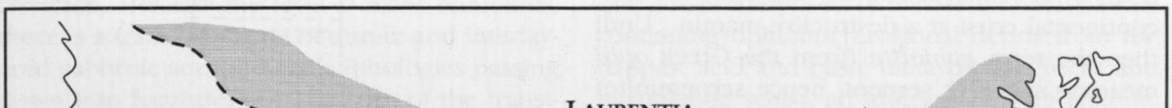

Figure 2.1 A pre-Atlantic reconstruction of the Caledonian Orogen showing the positions of the principal ophiolite complexes. A location map for the Scottish examples is shown inset.

the Shetland ophiolite is probably a little older with similar generation and obduction ages to some of the Norwegian examples. The broad coincidence of ages probably reflects the initiation of ocean closure when intra-oceanic subduction zones were formed adjacent to both continental margins. The volcanic products of these zones would have had greater obduction potential than the oceanic crust that was subsequently consumed by subduction at the continental margin. At that time the margins of the Iapetus Ocean were probably similar in character to the modern west Pacific Ocean with a complicated pattern of inter-related volcanic island arcs and back-arc basins.

The ophiolite model

The common association of serpentinite, gabbro, spilitic pillow lava and pelagic sedimentary rock has long been recognized by geologists and the idea of a kindred genetic relationship was introduced by Steinmann (1927). However, it was the later development of the plate tectonic theory that allowed a coherent interpretation as oceanic crust, generated at a constructive plate margin but subsequently accreted or thrust onto continental crust at a destructive margin. Until then the term *ophiolite* (from the Greek *ophi* meaning snake or serpent, hence serpentinite) had been used in a variety of ways but in 1972 the Geological Society of America convened a conference to formalize both its use and the underlying concept; the resulting proposals have been generally adopted. *Ophiolite*, as now defined (Anon, 1972), refers to a distinctive assemblage of mafic to ultramafic rocks which, in a complete development (Figure 2.2), consists of rock types in the following sequence, starting from the top and working down:

- *Mafic volcanic complex*, commonly of pillowed lava.
- *Mafic sheeted dyke complex*.
- *Gabbroic complex*, commonly with cumulus textures.
- *Ultramafic complex*, consisting of variable proportions of harzburgite, lherzolite and dunite (ultramafic rock names are defined in Figure 2.3), commonly serpentinized and with a tectonic fabric.

Associated rock types include: overlying pelagic sedimentary rocks such as chert, podi-

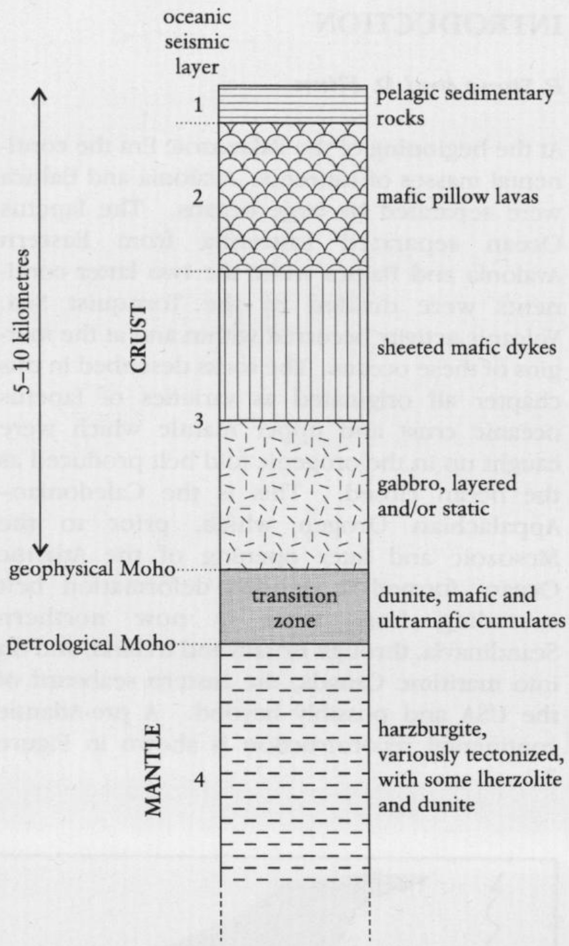

Figure 2.2 An idealized ophiolite succession compared to the seismic structure of oceanic crust and mantle (after Coleman, 1977). As an indication of scale, beneath the oceans the depth to the geophysical moho is between 5 and 10 km.

form chromite bodies in dunite, sodium-rich felsic intrusive rocks. The definition stresses that although *ophiolite* is generally interpreted to be oceanic crust and upper mantle, the use of the term should be independent of the supposed origin. As this terminology has been applied it has become common practice to refer to the whole assemblage as an *ophiolite complex*.

The comparison with oceanic crust relates the ultramafic rocks to the upper mantle and the gabbroic rocks and sheeted dykes to seismic layer 3 of the oceanic crust; the pillow lavas form

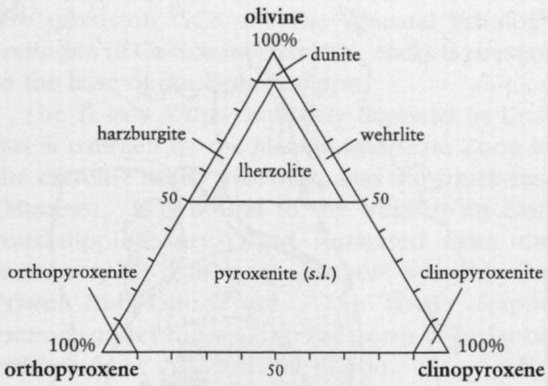

Figure 2.3 Descriptive nomenclature for ultramafic rocks in terms of their relative content of olivine and pyroxene.

layer 2 and the overlying pelagic strata form layer 1 (Figure 2.2). On this basis the geophysically defined Moho (occurring in the 5–10 km depth range beneath oceanic crust) coincides with the base of the gabbroic unit in an ophiolite complex. Beneath this level in many ophiolites, there is a transition zone of dunite and interlayered gabbroic and ultramafic lithologies passing down into harzburgite. The base of the transition zone has become known as the petrological Moho, which may lie up to several kilometres beneath the geophysical Moho. This zone is preserved within the Punds to Wick of Hagdale and Skeo Taing GCR sites on Shetland.

As work on ophiolites has progressed it has been established that many are in fact atypical of oceanic crust and many have been generated above subduction zones, i.e. at a destructive plate margin, rather than at a constructive mid-ocean ridge. Interpretation difficulties are also caused by the incomplete and fragmentary nature of most preserved ophiolites. The Scottish Caledonian examples of Shetland, the Highland Border and Ballantrae illustrate many of these problems and are of international importance in the overall assessment of the relationship of ophiolites to oceanic crust.

The Shetland Ophiolite

The Shetland Ophiolite forms a large part of the islands of Unst and Fetlar in the NE of Shetland (Gass *et al.*, 1982; Flinn, 1996). It is one of the series of ophiolite fragments distributed along the Caledonian orogenic belt from North America to Norway (Figure 2.1), most of which, including Shetland, were formed and obducted around 500 Ma or soon thereafter. Radiometric age data from Shetland suggests intrusive activity at 492 ± 3 Ma (U-Pb on zircon in a 'plagiogranite', *sensu* albite-granite, vein; Spray and Dunning, 1991) and the initiation of obduction at 498 ± 2 Ma (Ar-Ar on hornblende in amphibolite; Flinn *et al.*, 1991). The nearest to Shetland of the other Caledonian ophiolites is Karmoy in Norway, 400 km to the ESE across the orogen, while the nearest in the UK (excepting the small fragmentary vestiges of the Highland Border Complex) is Ballantrae, 600 km to the SSW along the orogen. The upper part (layers 1 and 2) of the characteristic ophiolite sequence (Figure 2.2) is not preserved in Shetland but the lower part is exceptionally well developed and exposed. Overall, the Shetland Ophiolite is among the finest and most accessible examples of ophiolitic rocks in Europe.

The ophiolite complex is composed of two tectonic units, the Upper and Lower nappes (Flinn, 1958), separated and underlain by imbricate zones composed of metasedimentary strata containing ophiolitic erosional debris from the nappes, acid and basic metavolcanic rocks and hornblende schist, all interleaved with tectonic slices of ophiolitic rocks. Figure 2.4a shows the distribution of these units together with that of the gneissose metasedimentary basement to the complex which crops out farther west; Figure 2.4b summarizes the lithological outcrop.

The *Upper Nappe* occurs as a series of widespread klippen composed largely of metaharzburgite of mantle origin; it rests either directly on the Lower Nappe or on the Middle Imbricate Zone separating the two nappes. The Upper Nappe is made of highly magnetic rocks, which enable it to be traced under the sea by means of positive aeromagnetic anomalies. The main Vord Hill Klippe occupies the centre of Fetlar and extends north under the sea to the island of Haaf Gruney. The nappe extends northwards to Hill of Clibberswick and westwards to the island of Sound Gruney via a series of small klippen, several of which also occur along the western edge of the complex in Unst. The Virva GCR site reveals that the Upper Nappe is underlain by hornblende schist derived from basic magma intruded into the thrust plane beneath the nappe and later cut into tectonic

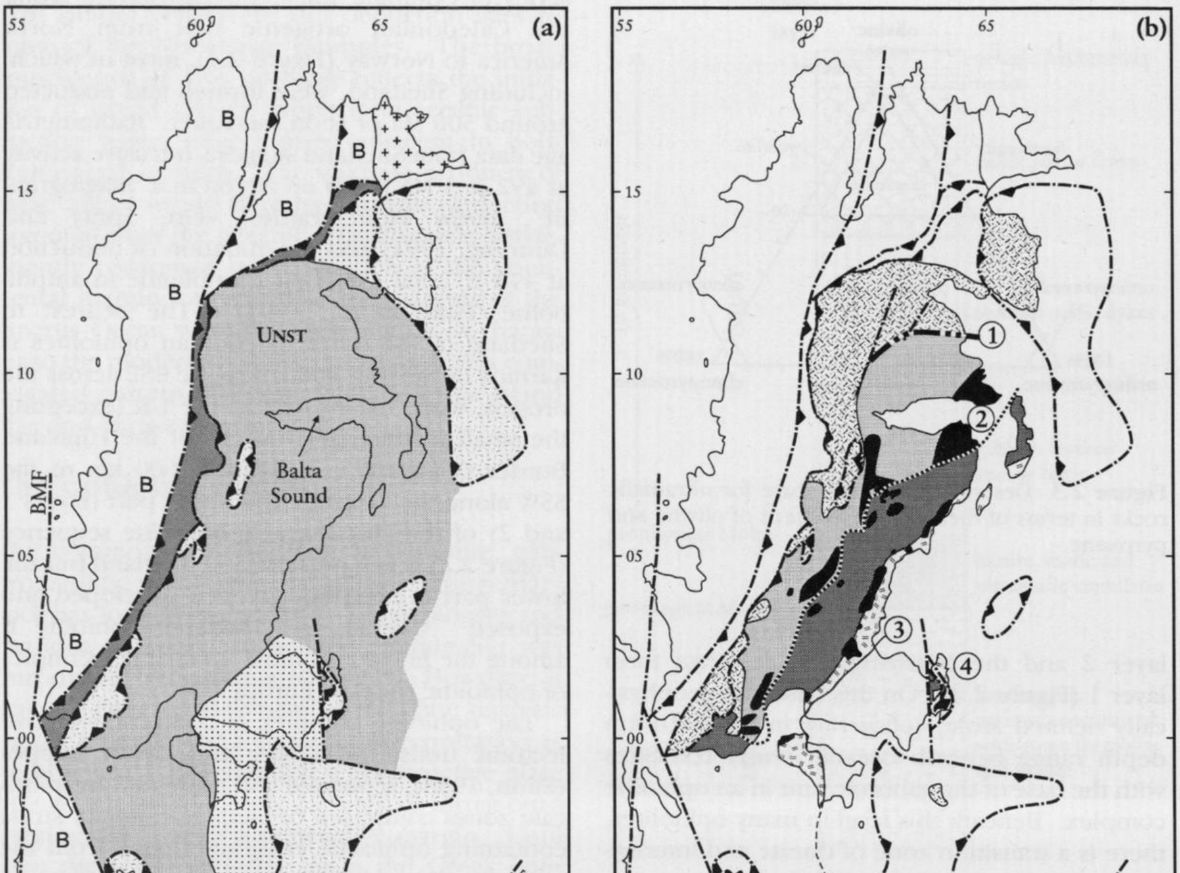

Figure 2.4 Maps of the Shetland Ophiolite (after Flinn, 1996): (a) principal tectonic units, (b) lithological outcrops. GCR sites: 1, The Punds to Wick of Hagdale; 2, Skeo Taing to Clugan; 3, Qui Ness to Pund Stacks; 4, Ham Ness; 5, Tressa Ness to Colbinstoft; 6, Virva.

slices by the thrusting, while in the Tressa Ness to Colbinstoft GCR site the unusual lithology rodingite (a Ca-rich metasomatic rock) is present in the base of the Upper Nappe.

The *Lower Nappe* is widely exposed in Unst but is overlain by the Middle Imbricate Zone in the extreme north (Norwick) and the south-east (Muness). It is bound to the west by an eastward-dipping thrust and separated from the metamorphic basement farther west by the Lower Imbricate Zone. The Lower Nappe extends under the sea into the north of Fetlar on either side of the Vord Hill Klippe.

Unlike the Upper Nappe, the Lower Nappe is composed of a conformable series of vertically orientated layers each several kilometres thick. The layers represent much of the conventional ophiolitic pseudo-stratigraphical succession (Figure 2.2). A thick layer of metaharzburgite, very similar to that forming the Upper Nappe, forms the northern part of the Lower Nappe (Figure 2.4b). It is followed to the south (up the ophiolitic succession) by a thick metadunite layer. The Punds to Wick of Hagdale GCR site presents evidence that the metaharzburgite is serpentinized infertile mantle whereas the metadunite is a serpentinized intrusive mass overlying the mantle and separated from it by the petrological Moho (Figure 2.2). It corresponds to the ultramafic layered lower crustal unit (transition zone *sensu lato*) of the conventional ophiolite succession. The intrusive nature of the metadunite layer and the adjacent metagabbro layer to the south can be demonstrated in the Skeo Taing to Clugan GCR site by the presence in both layers of xenoliths and screens of banded wehrlite–clinopyroxenite. In the Qui Ness to Punds Stack GCR site the eastern boundary of the metagabbro layer is marked by a discontinuous layer of wehrlite–clinopyroxenite xenoliths and screens, beyond which is an upper metagabbro layer. The latter is characterized by fine-grained metagabbros intruded by a large number of basic, dyke-like sheets giving this layer a quasi-sheeted appearance. The Ham Ness GCR site presents the upper metagabbro layer overthrust by the Upper Nappe.

The Highland Border Complex

Within the Highland Boundary Fault Zone (Figure 2.1), a series of structurally isolated slivers contain various combinations of serpentinite, pillow lava, black shale, chert, limestone and sandstone, apparently ranging in age from Early Cambrian to late Ordovician. The igneous components of this lithological association are distinct from the Dalradian Supergroup rocks to the north and the whole assemblage contrasts with the Upper Palaeozoic strata of the Midland Valley Terrane to the south. All of the disparate elements are grouped together as the Highland Border Complex which remains largely enigmatic. The current consensus view of the regional development of the complex (see Robertson and Henderson, 1984; Curry *et al.*, 1984) envisages its formation in a small oceanic basin marginal to the Laurentian continent. However, there are aspects of the relationship between the Highland Border Complex and the Dalradian that remain poorly understood. In particular, some of the Early Cambrian sedimentary strata seem to lie in stratigraphical continuity with parts of the Dalradian succession and to have experienced a similar structural history (Tanner, 1995 and references therein). The problems are accentuated by the extreme tectonic disruption whereby the complex now exists as about ten fault-bound lenses, spread along the length of the Highland Boundary, from Stonehaven on the east coast of Scotland to Arran in the west (Figure 2.1).

Despite the tectonic disruption an overall succession has been established. The serpentinites are regarded as the oldest igneous component, of Late Cambrian or early Ordovician (Tremadoc?) age, and are overlain unconformably by conglomerates with serpentinite clasts; an associated limestone unit contains a mid-Arenig fauna. This part of the succession is well displayed at the Balmaha and Arrochymore Point GCR site and serpentinite is also present in the Garron Point to Slug Head GCR site. The relationship of the serpentinite to a slightly younger sequence dominated by basic pillow lava and black shale is uncertain but the lava-shale unit appears to range in age from the Arenig, through the Llanvirn and possibly into the Llandeilo. Associated with the lava and shale are tuff, chert and quartz-greywacke. A probable Arenig part of this unit, including an impressive array of pillow lava, is well exposed in the North Glen Sannox GCR site; the pillow lavas seen in the Garron Point to Slug Head GCR site may possibly be younger (Llanvirn–Llandeilo) in age although the evidence is meagre. The youngest part of the Highland Border Complex appears to be a Caradoc to possibly Ashgill sedimentary sequence of sandstone with ophiolitic, acid volcanic and quartz clasts, limestone and shale. Elements of this sedimentary association are seen in the Balmaha and Arrochymore Point and the North Glen Sannox GCR sites.

The closure of the Highland Border Complex basin (or basins) seems likely to have been a polyphase and protracted event involving oblique terrane amalgamation. A complicated and possibly protracted metamorphic history is suggested by stable isotope and petrographical evidence presented by Ikin and Harmon (1984). Indications of Cambrian tectonism come from the radiometric ages of around 540 Ma reported by Dempster and Bluck (1991) for an amphibolite associated with serpentinite on the island of Bute. This was taken as evidence for Cambrian initiation of ophiolite obduction but, elsewhere, Ordovician sedimentation clearly occurred at least during the Arenig–Llanvirn and Llandeilo–Caradoc intervals (Curry *et al.*, 1984). Structural evidence is similarly ambiguous. Henderson and Robertson (1982) thought that tectonic features suggest a similar deformation history for the Highland Border Complex and the Dalradian succession to the NW; they envisaged thrust emplacement of the complex towards the SE. Conversely, Curry *et al.* (1984) proposed that remnants of a marginal basin floor were obducted towards the NW onto the already deformed Dalradian. As a further complication, Harte *et al.* (1984) concluded that the later transcurrent faulting was largely responsible for the present lenticular disposition of the complex along the Highland Boundary Fault.

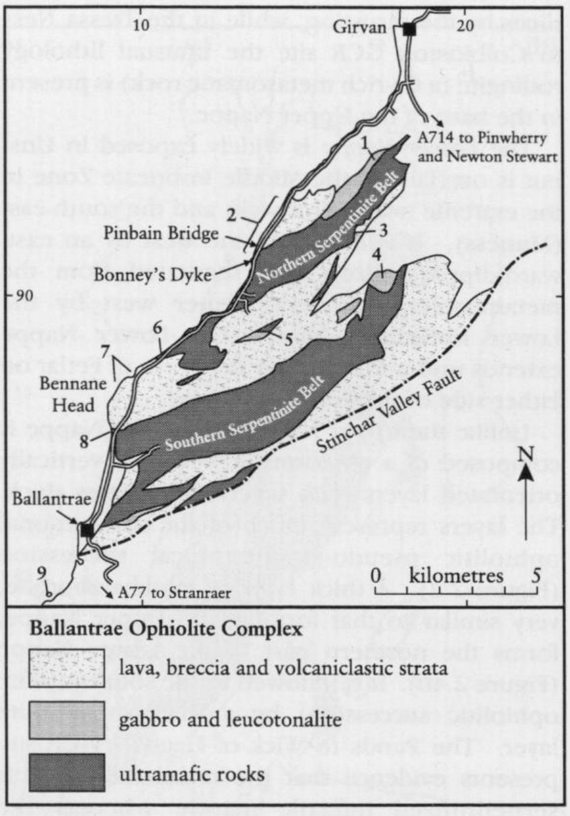

Figure 2.5 Outline geology of the Ballantrae Complex (after Stone and Smellie, 1988) showing the location of the GCR sites. 1, Byne Hill; 2, Slockenray coast; 3, Knocklaugh; 4, Millenderdale; 5, Knockormal; 6, Games Loup; 7, Balcreuchan Port to Port Vad; 8, Bennane Lea; 9, Sgavoch Rock.

The Ballantrae Complex

Between Girvan and Ballantrae, on the west coast of southern Scotland, a structurally complicated assemblage consisting mainly of serpentinized ultramafic rocks and tholeiitic lavas crops out over an area of about 75 km². This is the Ballantrae Complex, which has been widely interpreted as an obducted ophiolite (Church and Gayer, 1973; Bluck, 1978a) although there is little evidence of sheeted dykes and only fragmentary occurrences of layered gabbro. The outline geology and location of the GCR sites is shown in Figure 2.5. The internal structural relationships are complicated and all of the major lithological boundaries are faulted. The volcanic rocks form three fault-defined blocks trending NE–SW and separated by two blocks of ultramafic rock, the Northern and Southern serpentinite belts.

The lavas, together with abundant volcaniclastic breccia and sandstone, have been lithostratigraphically defined as the Balcreuchan Group (Stone and Smellie, 1988) aspects of which may be seen within the Slockenray Coast, Games Loup, Balcreuchan Port to Port Vad and Bennane Lea GCR sites. The lavas were erupted in a submarine environment with pillow structure widely developed. At a few localities, graptolites of Arenig (early Ordovician) age have been recovered from sedimentary interbeds (Stone and Rushton, 1983; Rushton *et al.*, 1986) while the tholeiitic basalt lavas have themselves given Sm-Nd radiometric ages of 501 ± 12 Ma and 476 ± 14 Ma (Thirlwall and Bluck, 1984). Eruption would therefore seem to have commenced during the Tremadoc and probably reached its peak during the Arenig. However, a number of studies of the lava geochemistry have concluded that eruption took place in disparate geotectonic environments with the Balcreuchan Group therefore containing a polygenetic structural assemblage (Wilkinson and Cann, 1974; Thirlwall and Bluck, 1984). The basis for this conclusion lies in the different trace element ratios and abundances characteristic of basalts erupted in either mid-ocean ridge, or within-plate, or island arc settings. A clear consensus has emerged that island-arc and within-plate lavas are both present but that there is very little evidence for a mid-ocean ridge volcanic component (Wilkinson and Cann, 1974; Lewis and Bloxam, 1977; Thirlwall and Bluck, 1984; Stone and Smellie, 1990). Further support for an island-arc involvement has come from the discovery of boninite lavas and breccias at the Balcreuchan Port to Port Vad GCR site (Smellie and Stone, 1992; Smellie *et al.,* 1995); these are unusual lithologies with modern examples known only from primitive, oceanic island arcs.

The ultramafic components of the Ballantrae Complex are now largely serpentinized; the protoliths were principally dunite and harzburgite with a small proportion of wehrlite and various pyroxenites. Ultramafic rocks are a particular feature of the Knockormal, Knocklaugh and Games Loup GCR sites. An origin within 'normal' (mid-ocean ridge) oceanic mantle was initially assumed in the ophiolite interpretation, and a geochemical study of a limited area of the Northern Serpentinite Belt by Jelínek *et al.* (1980) has supported this assumption. However, Stone and Smellie, (1990) considered that a range of features are atypical of ultramafic rock from mid-ocean ridge ophiolites but instead match those considered characteristic of supra-subduction zone (i.e. volcanic arc) environments by Pearce *et al.* (1984). Important aspects encouraging this conclusion are the overall lithological dominance of titanium-depleted harzburgite, the dominance of wehrlite over troctolite in the layered sequences, and the relative chromium enrichment of the accessory spinels over those found in a mid-oceanic setting. Thus the ultramafic rocks of the Ballantrae Complex are also indicative of an origin, for at least part of the complex, at an island arc above an oceanic subduction zone. The involvement of older oceanic mantle, perhaps forming the foundation to the arc development, is suggested by a Sm-Nd radiometric age of 576 ± 32 Ma reported by Hamilton *et al.* (1984) from a garnet clinopyroxenite body within the Knockormal GCR site, that was interpreted as a segregation in mantle harzburgite by Smellie and Stone (1984). A minimum age is provided by a U-Pb on zircon radiometric date of 483 ± 3 Ma reported by Bluck *et al.* (1980) from a leucotonalite, intruded into and chilled against the Northern Serpentinite Belt, within the Byne Hill GCR site.

The Byne Hill leucotonalite is gradational from gabbro but the gabbroic and sheeted dyke components of the ideal ophiolite assemblage are sparse and fragmentary within the Ballantrae Complex. Foliated gabbros cut by numerous dykes within the Millenderdale GCR site have been likened to a sheeted dyke association but with reservations (Bluck, 1978a). An exceptionally coarse pegmatitic gabbro segregation within harzburgite occurs within the Slockenray Coast GCR site.

The most likely setting for generation of the Ballantrae Complex would seem to be within a Tremadoc to Arenig oceanic island arc and associated marginal, back-arc basin. The youngest graptolite fauna recovered from the complex is of late Arenig age (about 480 Ma) (Stone and Rushton, 1983) while a number of radiometric dates overlap around an upper range of 475–485 Ma. Obduction of the ophiolite, perhaps caused by collision of the volcanic arc with the continental margin of Laurentia, could have followed soon after. Its initiation may well be dated by the K-Ar age of 478 ± 8 Ma reported by Bluck *et al.* (1980) from an amphibolite within the Knocklaugh GCR site. This amphibolite is part of a metamorphic assemblage thought to have formed in a dynamothermal aureole adjacent to the base of the Northern Serpentinite Belt when, as a slab of hot mantle material, it was thrust up through the crustal carapace of volcanic-arc lavas. A minimum age for obduction of the Ballantrae Complex is provided by the unconformably overlying, Llanvirn shallow-marine strata.

The Southern Uplands

The Southern Uplands Terrane formed as an accretionary thrust complex at the Laurentian continental margin during Ordovician and Silurian subduction of the Iapetus Ocean. The terrane has been widely interpreted as a fore-arc prism, formed above the active subduction zone

(Leggett *et al.*, 1979; Leggett, 1987). Alternatively, it may have developed from an Ordovician back-arc setting into a mid-Silurian foreland basin, the latter generated as the thrust front migrated onto the Avalonian continent following closure of the Iapetus Ocean (Stone *et al.*, 1987). In either model the resulting structural configuration is the same: a sequence of southward propagating imbricate thrusts separating lithostratigraphical tracts of steeply inclined strata that strike NE–SW. Internally the tracts have an overall sense of younging towards the north, whereas the minimum age of each tract decreases southwards. This phenomenon has been recently illustrated and discussed by Rushton *et al.* (1996 and references therein) while a full stratigraphy is given by Floyd (1996). The dominant lithology within each tract is turbidite-facies greywacke but this overlies a thin basal assemblage of black mudstone, chert and sporadic basic lava and hyaloclastite. The lava and hyaloclastite are found in the northern part of the terrane and appear to range in age from Arenig to Caradoc; recent reviews are provided by Phillips *et al.* (1995b) and Armstrong *et al.* (1996). The mudstone and chert range in age from the Caradoc well up into the mid-Silurian and throughout the succession are interbedded with subalkaline and mildly peralkaline metabentonite layers (Merriman and Roberts, 1990).

Phillips *et al.* (1995b, 1999) show that the oldest (Arenig) lavas include both alkaline within-plate basalts and tholeiitic varieties with probable mid-ocean ridge basalt characteristics. Other lavas of uncertain age within the Arenig–Caradoc range have compositions more suggestive of an island-arc source, while definite Caradoc lavas are of within-plate, ocean island affinity. Some of the latter are interbedded with the greywacke sequence rather than forming a base to the mudstone. Apart from these lava occurrences, three other examples of Ordovician volcanic rocks are worthy of note.

1. Peralkaline rhyolite and tuff (the Wrae or Tweeddale lavas), interbedded with greywacke of late Caradoc or early Ashgill age, may have been emplaced as a debris flow from a sea-mount volcano (Thirlwall, 1981b).
2. The Bail Hill Complex is a 1.8 km thickness of pyroclastic rock and lava, the latter comprising mugearite and hawaiite, thought to represent a sea-mount volcano enclosed by transgressive early Caradoc greywackes (Hepworth *et al.*, 1982; Phillips *et al.*, 1999).
3. The Downan Point Lava Formation (Stone and Smellie, 1988 and references therein) occupies a faulted wedge sandwiched between the northern margin of the Southern Uplands Terrane and the Ballantrae Complex. Tholeiitic basalts predominate and have within-plate, ocean island geochemical affinities. The formation has been associated historically with the Ballantrae Complex of Arenig age but a current consensus favours a younger, early Caradoc age and considers it to form the northernmost tract of the Southern Uplands. The coastal exposures present an extraordinary array of pillow structures, as seen at the Sgavoch Rock GCR site.

The full assemblage of Ordovician volcanic rocks in the northern part of the Southern Uplands Terrane is difficult to readily reconcile with either of the geotectonic models proposed for its development; fore-arc accretionary prism or back-arc (to Silurian foreland basin) thrust belt.

THE SHETLAND OPHIOLITE

THE PUNDS TO WICK OF HAGDALE (HP 647 103–644 113)

Derek Flinn

Introduction

The site lies on the east side of Unst within the Lower Nappe of the Shetland Ophiolite. On the east side is a continuously exposed low-lying cliff section which is of great importance in that it contains the uppermost kilometre of the mantle, the petrological Moho, and a kilometre of the overlying metadunite layer. Inland to the west abundant exposures, ideally weathered, ice-smoothed and lichen-free, reveal in unusual detail the petrography and mineralogy of these rocks. Hitherto, these two units of the ophiolite pseudo-stratigraphy in Unst have received the conventional labels of 'tectonized harzburgite' and 'cumulate dunite' and the latter has been interpreted as the result of the cumulate crystallization of olivine in the base of a magma chamber resting on tectonized mantle at an oceanic constructive plate margin (Gass *et al.*, 1982;

Prichard, 1985). However, Flinn (1996) has reached the contrary conclusion that the metadunite is an independent intrusive unit which did not form as a cumulate layer in a magma chamber and that the harzburgite is not tectonized. The site offers a magnificently exposed and easily accessible view of the mantle-lower crust junction.

Description

The rocks to the north of the petrological Moho in Figure 2.6, representing the uppermost mantle, are dominantly *metaharzburgite* with lesser amounts of metadunite. Both rocks, being uniformly and extensively altered to hydrous minerals, contain about 13 wt% of water. Lizardite-serpentine, which replaces olivine, weathers to a strong ochrous colour due to a relatively high iron content, while the hydrous minerals replacing pyroxenes weather white and stand slightly proud of the serpentine. Thus the two rocks are easily distinguished, even where intimately mixed. However, adjacent to some of the major shear zones the lizardite-serpentine has been recrystallized to antigorite-serpentine with magnetite and weathers white making it difficult there to distinguish between metaharzburgite and metadunite.

It is apparent that prior to serpentinization the metaharzburgite was composed dominantly of olivine together with 20–30% of orthopyroxene, and about 0.5% of accessory chromite. Thin sections commonly contain about 1% of clinopyroxene. The metadunite was originally almost entirely composed of olivine, with about 1% of accessory chromite. Clinopyroxene is present only very locally (see below). The grain size of the olivine and pyroxene is about 0.5 cm.

The metaharzburgite is rhythmically banded in places. The bands are sharply defined, rectilinear and dip steeply north. They are formed by alternations in olivine/pyroxene proportions, but are ungraded. Olivine-rich and pyroxene-rich pairs are about 30 cm thick, and are parallel to the petrological Moho towards which their persistence and frequency of occurrence increases. Generally, the bands can rarely be followed for more than a few metres, only rarely show signs of disruption or folding and are untectonized. The banding is very well displayed on the coast at The Viels (6444 1108) (Figure 2.7) and on the hillside to the west of the road.

The metaharzburgite is cut by many thin pyroxenite veins filling planar joint-like fractures. The veins are formed of pyroxenite with grain sizes up to one centimetre or more, and much less commonly of olivine-pyroxenite. They are always completely pseudomorphed by fine-grained hydrous minerals. They have been referred to as dykes (Prichard, 1985), but to avoid confusion with the basic dykes forming the quasi-sheeted dyke layer they are referred to here as veins. Many strike parallel to the rhythmic banding, where present, but rarely dip parallel to it; hence they cut the banding. In unbanded areas they have no preferred orientation. In a small number of localities they are seen to have been deformed by cross-cutting early shear zones. Pyroxenite veins are cut by metadunite bodies in the metaharzburgite. Since most orthopyroxene in the metaharzburgite is altered to the same secondary minerals as the pyroxenes in the veins, whereas clinopyroxene generally remains fresh, these were probably orthopyroxenite veins. An irregular sheet of fresh, olivine-free orthopyroxenite about 10 cm thick occurs at The Viels (6446 1108).

Metadunite within the metaharzburgite occurs as rare and sporadically distributed bodies that increase in frequency towards the petrological Moho within the northern part of Figure 2.6. Most are sub-equidimensional bodies or clots a metre or so in diameter with no detectable preferred orientation of their shapes; larger bodies up to a hundred metres or more across also occur and these are commonly of very irregular shape. Several of the larger metadunite bodies contain podiform chromite masses. Parallel-sided sheets of metadunite up to 20 cm thick and several metres in length occur with increasing frequency southward and, for two or three hundred metres north of the petrological Moho, the metadunite clots and sheets are particularly densely distributed, equalling in places the metaharzburgite in volume.

The metadunite sheets within the metaharzburgite adjacent to the petrological Moho are sub-parallel to the rhythmic banding and to the Moho; a high proportion are obliquely cross cutting by 10° or less, while a much smaller proportion cut at much greater angles or are exactly parallel. The clots of metadunite, especially those close to the petrological Moho often have invasive boundaries with thin wedge-like and sheet-like protuberances cutting the enclosing

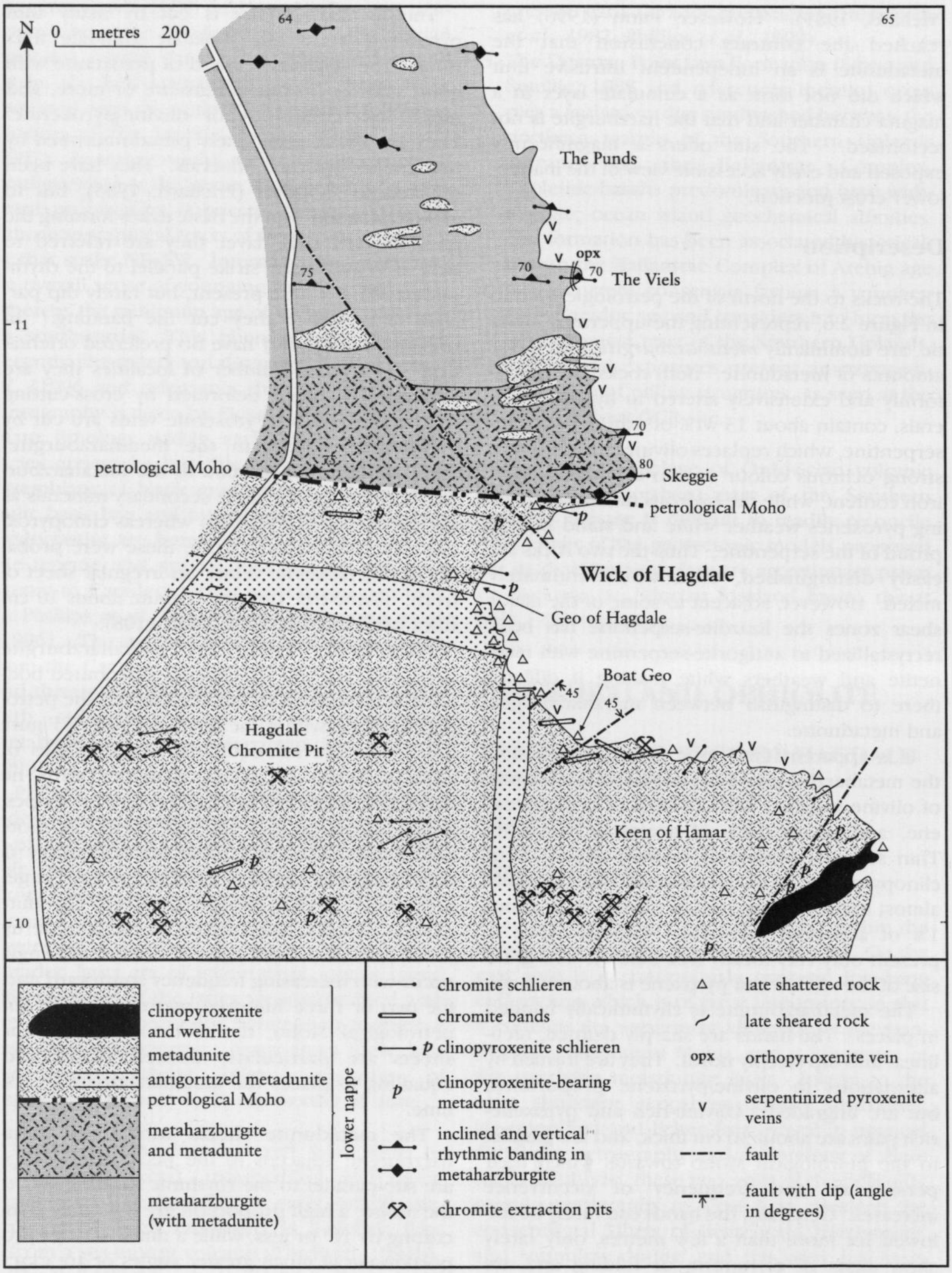

Figure 2.6 Map of the Wick of Hagdale area, Unst.

Figure 2.7 Rhythmic banding in the metaharzburgite at The Viels (HP 6444 1108) close to the petrological Moho. The hammer is 37 cm long. (Photo: D. Flinn.)

metaharzburgite and giving them an intrusive appearance.

The *petrological Moho* is taken as the southern edge of the southernmost band of metaharzburgite. It has an E–W strike parallel to the rhythmic banding, and has a 70–80° dip to the north. It is exposed on the coast (Figure 2.6) and in several inland exposures.

The *metadunite* layer is dominantly composed of rock indistinguishable from the metadunite forming bodies and sheets within the metaharzburgite layer. Thin sections reveal equidimensional olivine grains about half a centimetre in size forming a triple-junction grain-boundary network, but which have been extensively replaced by iron-rich mesh-type serpentine so that the rock weathers to a bright ochrous colour, lacking the white-weathering altered pyroxenes characteristic of the metaharzburgite. The metadunite is lithologically uniform and texturally isotropic, except for the local presence of chromite and clinopyroxene.

Chromite, in both the metadunite layer and in the metadunite sheets and bodies within the metaharzburgite, occurs as disseminated accessory grains no more than 1 mm in size and as coarser grains several millimetres in diameter forming band- or vein-like concentrations that locally expand into podiform bodies. Some accessory chromite forms schlieren, typically several centimetres thick and ten times as long, in which the individual chromite grains are several millimetres apart. Schlieren are well exposed on the coast just south of the Moho in Figure 2.6 but occur more commonly in the metadunite to the west of the road. The bands formed of more concentrated coarsely crystalline chromite cannot be studied as extensively. They were used by prospectors as guides to the presence of podiform masses and have almost all been excavated, leaving pits, trenches and spoil heaps to mark their previous existence. The best remaining locality for the study of in-situ banded chromite occurs in the cliffs 100 m east of Boat Geo (6456 1030). The spoil heaps of some chromite pits are a good source of banded chromite, but the waste heaps of the famous Hagdale Quarry have been despoiled by collectors and carried away for infill. Specimens remaining in waste heaps elsewhere in Unst reveal bands of chromite cut by metadunite. The presence of angular fragments of massive chromite floating in metadunite and the folding of the chromite bands indicate that fracturing and deformation of the deposits took place while the dunite was still mobile. Disseminated accessory chromite is also present in the metaharzburgite, though much less abundantly than in the metadunite.

Clinopyroxene occurs locally and rarely as sub-millimetre, isolated, interstitial fresh grains. It may be either uniformly distributed or form schlieren in the metadunite layer but it never occurs in metadunite within the metaharzburgite. Both types of occurrence of clinopyroxene are to be found closely associated with similarly occurring chromite on the coast of the Wick of Hagdale in the 200–300 m to the south of the petrological Moho (Figure 2.6) and inland elsewhere. Interstitial clinopyroxene occurs most commonly in the top of the metadunite layer close to wehrlite–clinopyroxenite masses. An example is indicated in the SE corner of Figure 2.6 but this relationship is better displayed in the Skeo Taing to Clugan GCR site.

Interpretation

The metaharzburgite has been recognized as infertile mantle (Gass *et al.*, 1982; Prichard, 1985), on the basis of its composition and by analogy with other ophiolite complexes (Gass, 1980). This makes its boundary against the metadunite layer to the south the upper limit of the mantle and therefore the petrological Moho (Figure 2.2).

The metaharzburgite resembles that of many other ophiolites in the presence of the enclosed metadunite bodies, the pyroxenite veins, the rhythmic banding and the accessory chromite. Harzburgite in ophiolites is generally held to have been tectonized by convectional flow at a constructional plate margin before obduction and the tectonite nature of the harzburgite has been cited as evidence of the ophiolitic nature of the Shetland complex by the above mentioned authors. Bartholomew (1993) reported the occurrence throughout the metaharzburgite of two vertical foliations intersecting at high angles. These foliations were recognized in the field by the flattening of the accessory chromite grains and were confirmed in thin-section by a preferred orientation of grain shapes, interpreted by Bartholomew as arising from simple shear during mantle flow. However, Flinn (1996) found no trace of the two foliation directions and reported only that, in a number of places, a slight flattening of the chromite grains parallel to the plane of schlieren is present, but never in two widely different directions in the same area. He also noted that pyroxene grain-shapes are clearly etched out by weathering and show no grain-shape preferred orientation, even in rhythmically banded areas. The tectonite status of the metaharzburgite therefore remains controversial.

The conventionally accepted origin (Prichard, 1985) for the metadunite bodies within the metaharzburgite and the metadunite above the petrological Moho is that they crystallized out of basic magma, produced by adiabatic melting deep in the mantle, as it rose to form a magma chamber above the mantle. However, field evidence from this and the Skeo Taing to Clugan GCR site has been interpreted as showing that the metadunite layer above the mantle is an intrusive unit and not a layer accumulated on the floor of a magma chamber (Flinn, 1996).

In support of his interpretation Flinn cited Kelemen *et al.* (1995) who have suggested that MORB (mid-oceanic ridge basalt) magma produced by adiabatic melting deep in the mantle would be pyroxene undersaturated and would rise through the mantle by intergranular flow. Such magma would dissolve the pyroxene grains in the mantle and so become focused into conduits made more porous than the surroundings, the conduits thus becoming columns of dunite. It is possible that the enhanced porosity of the dunite in the MORB conduits (a MORB : dunite ratio of 1:10, according to Kelemen *et al.*, 1995) could so lower the cohesion of the olivine grains that the rising MORB would fluidize the dunite. The fluidized dunite would then flow up its conduits through the mantle giving rise to the observed invasive relationships of the dunite relative to the enclosing harzburgite, and including the sheet-like injections close to the Moho. The upward flowing dunite accumulated above the mantle as an intrusive layer several kilometres thick and when flow ceased the conduits within the mantle broke up into independent bodies.

Neither the conventional account of the origin of the metadunite bodies within the mantle (Prichard, 1985), nor that of Kelemen *et al.* (1995) provide an adequate explanation for the presence of podiform masses of chromite in some of the metadunite bodies within the mantle in Unst, whereas other, nearby and larger bodies contain only sparsely distributed accessory chromite grains. Nor do they account for the apparent random distribution of the chromite podiform masses in the metadunite layer above the mantle, or the fact that they are indistinguishable from those nearby within the mantle. Possibly the podiform masses were formed at depth in the mantle and carried up to their present positions in Unst in the mobilized dunite.

The origins of the clinopyroxene in the metadunite layer and of the wehrlite–clinopyroxenite masses in the upper part of the metadunite layer south of Figure 2.6 are more conveniently discussed in the context of the Skeo Taing to Clugan GCR site.

Conclusions

The Punds to Wick of Hagdale GCR site features an uninterrupted passage through the top kilometre of the mantle and the bottom kilometre of the overlying lower crustal succession. The profusion of glacially smoothed exposures makes the area ideal for the study of the environment of the petrological Moho, thus providing a rare opportunity of international importance in ophiolite geology. In particular, the site allows close investigation of the controversial intrusive relationship of the dunite to the mantle.

SKEO TAING TO CLUGAN (HP 646 067–639 084)

Derek Flinn

Introduction

The site lies on the east side of Unst within the Lower Nappe of the Shetland Ophiolite. It is bound to the east by a nearly continuously exposed low-lying cliff section which is of great importance in that it illustrates the uppermost kilometre of the metadunite layer, the geophysical Moho (not exposed), and a kilometre of the overlying lower metagabbro layer. Inland the site is not well exposed but presents evidence complementary to that in the cliffs. The metadunite layer has been interpreted conventionally as an ultramafic layer which, like overlying basic layers, accumulated in a magma chamber resting on the mantle below a constructive plate margin (Gass *et al.,* 1982; Prichard, 1985). However, much field evidence supports an alternative origin wherein the metadunite layer and the basic layers are separate intrusive bodies and not cumulate layers. Xenoliths of interbanded clinopyroxenite and wehrlite are prominent (Flinn, 1996). The site offers an easily accessible view of the passage from ultramafic to mafic rocks in the ophiolite succession and the very different form that this takes from the conventional ophiolite model.

Description

The geology of the Skeo Taing to Clugan site is summarized in Figure 2.8; note that the metadunite outcrop shown is a continuation (across strike) of that part of the same layer shown in Figure 2.6 and previously described at Wick of Hagdale (see The Punds to Wick of Hagdale GCR site). At Skeo Taing the top of the metadunite layer is exposed and typically is rich in wehrlite–clinopyroxenite xenoliths and screens, much more so than in the equivalent layer at Wick of Hagdale. Clinopyroxenite xenoliths too small to show on the map also occur in the Skeo Taing area; all are closely associated with interstitial clinopyroxene in the host metadunite.

In this account, metadunite containing interstitial clinopyroxene will be called clinopyroxene-bearing metadunite even where it contains sufficient clinopyroxene to be petrographically classed as wehrlite (Figure 2.3). This is done in order to distinguish it from the wehrlite associated with clinopyroxenite from which it is texturally distinct. In the metadunite small, fresh clinopyroxene grains occur interstitially to the larger, several-millimetre-sized serpentinized olivine grains whereas, in the wehrlite, the serpentinized small olivine grains are interstitial to the much larger, fresh clinopyroxene grains which in some samples are as much as a centimetre across.

The larger wehrlite–clinopyroxenite masses, those measured in tens of metres, form mound-like exposures rising a metre or two above the enclosing metadunite. The presence in them of abundant, fresh, large clinopyroxene grains makes the wehrlite–clinopyroxenite more resistant to erosion than the enclosing serpentinized dunite and clinopyroxene-bearing metadunite. As a result the wehrlite–clinopyroxenite masses

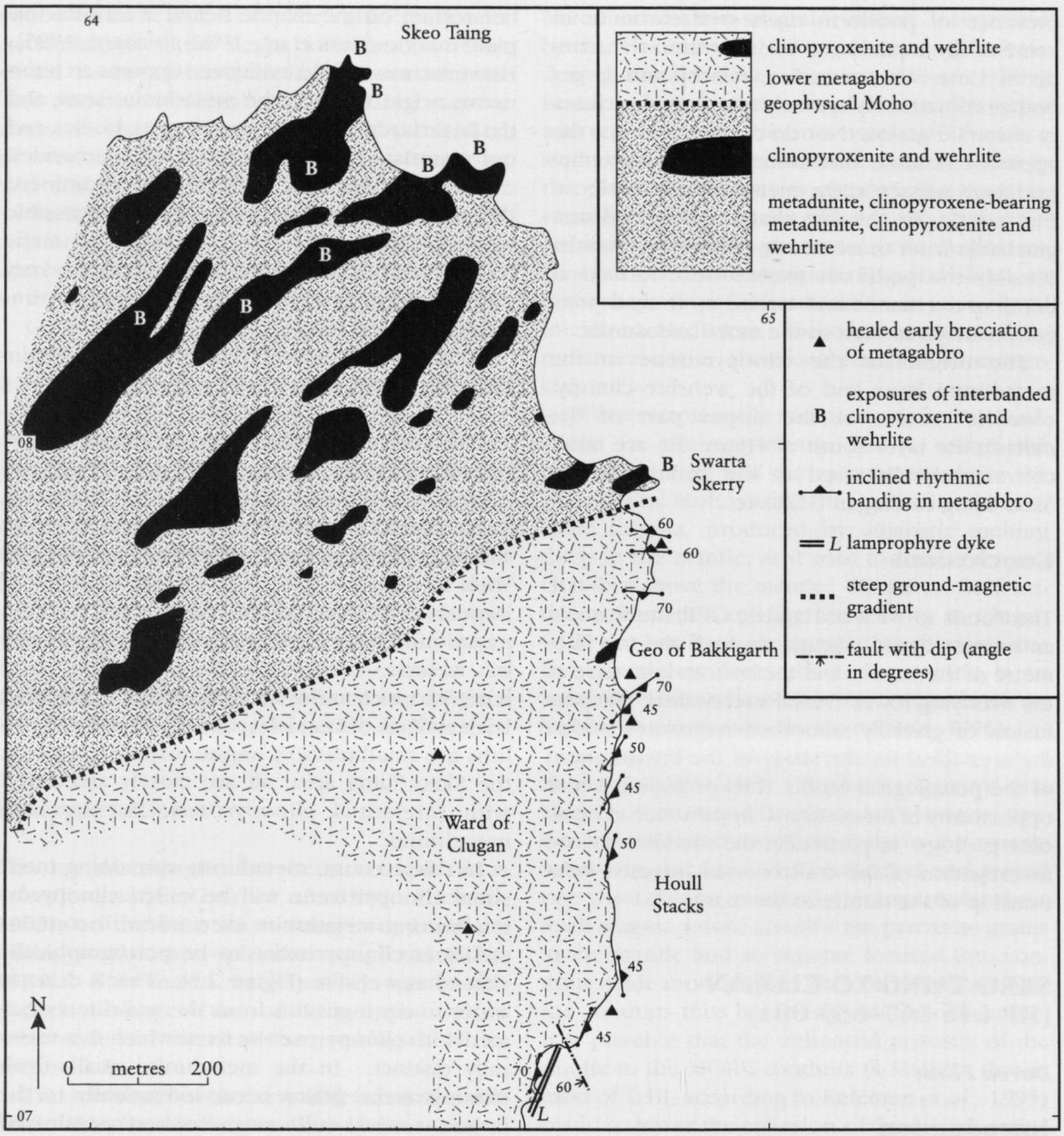

Figure 2.8 Map of the Skeo Taing area, Unst.

form the characteristic protuberant features that commonly carry an enhanced vegetation cover, probably due to the fertilizing effect of the Ca in the clinopyroxene. The low ground between the wehrlite–clinopyroxenite masses is occupied by their metadunite and clinopyroxene-bearing metadunite host rocks which, in the coastal exposures, can be seen to contain smaller wehrlite–clinopyroxenite xenoliths. The wehrlite–clinopyroxenite masses are only well exposed on the coast where it is apparent that they are formed of wehrlite, olivine-clinopyroxenite and clinopyroxenite complexly interbanded and intergraded on a scale ranging from several centimetres to several decimetres. The bands are much less rectilinear than those in the metaharzburgite and the metagabbro and can rarely be followed for more than a metre; some occurrences exhibit disruption and contortion of the banding.

The contact between the metadunite layer and the lower metagabbro layer to the south is seen in only one small exposure in Unst and that is at 610 054, outside the area of Figure 2.8. However, the contact is closely confined by exposures on either side for considerable distances in several places and everywhere it can be followed by magnetometer since it gives rise to a very steep ground-magnetic gradient. The contact thus sharply defined lacks transitional lithologies and is curvilinear and unsheared. In the upward succession it marks the incoming of feldspar and the exit of olivine (fresh or serpentinized) except in xenoliths. This creates the most profound lithological change in the succession and is therefore considered to represent the geophysical Moho.

The lower metagabbro is as uniformly hydrated as the metaharzburgite and the metadunite (see The Punds to Wick of Hagdale GCR site). The clinopyroxenes have been replaced by single-crystal plates of colourless to pale-green amphibole and the feldspars have been altered to very low birefringence, very fine-grained pseudomorphs. As a result the rock weathers to a speckled pale-green and white pattern. Relict clinopyroxene is rare. Twinned, fresh albitic plagioclase occurs sporadically, associated with the altered plagioclase. The metagabbro contains no evidence of having ever contained olivine. Fresh interstitial grains of brown hornblende occur as a sparsely distributed original accessory. The early uniform hydration, as in the metaharzburgite, was followed by a later low-grade metamorphism closely related to shearing.

The metagabbro is characterized locally by regular alternations of pyroxene-rich and feldspar-rich ungraded bands, each several centimetres thick (Figure 2.9). South of the geophysical Moho in the coastal section (south of (648 076); Figure 2.8), this banding is seen to be repeated continuously for several hundred metres but most occurrences are small patches or even single bands. The attitude of the bands varies considerably but generally the dip direction is between east and south.

Large areas of the metagabbro have been intensively brecciated but subsequently healed. The metagabbro fragments, each no more than several centimetres across, have all been slightly rotated relative to each other but the rock is as resistant as the unbrecciated variety and the frac-

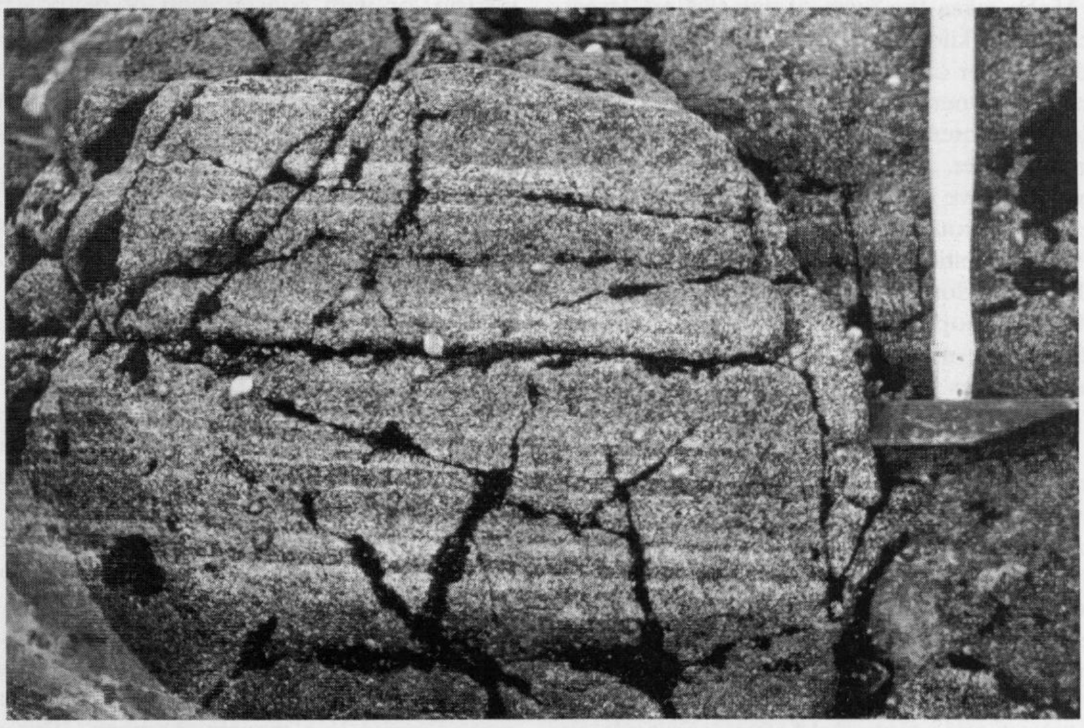

Figure 2.9 Rhythmic banding in metagabbro (HP 647 077). (Photo: D. Flinn.)

tures are not apparent in thin section. Such brecciation occurred earlier than the more common and widespread late shattering and shearing. This healed brecciation makes rhythmic banding difficult to detect.

At Geo of Bakkigarth (6475 0765), 200 m south of the geophysical Moho in Figure 2.8, there is a small xenolithic mass of wehrlite-dominated wehrlite–clinopyroxenite associated with some late shearing. This is one of a series of xenoliths, each up to several tens of metres across, composed of wehrlite–clinopyroxenite and clinopyroxene-bearing metadunite. These occur within the metagabbro close to its western boundary with the metadunite, but otherwise outside the area covered by Figure 2.8. Some are associated with swarms of pyroxenite fragments up to 10 cm across. Rhythmic banding in the metagabbro close to these blocks varies more widely than in the general mass of the metagabbro. All these wehrlite–clinopyroxenite masses are indistinguishable from those in the metadunite layer and like those are considered to be xenoliths.

Interpretation

In Shetland the succession above the petrological Moho takes the form of two distinct layers, each several kilometres thick, one of metadunite and the other of metagabbro. Both these layers contain fragments of a disrupted, coarsely crystalline and intensively banded wehrlite–clinopyroxenite layer. In the metadunite layer these xenoliths have been partly assimilated so that they are surrounded by metadunite containing isolated interstitial clinopyroxene grains.

The metadunite layer with its xenoliths of wehrlite–clinopyroxenite has been presented hitherto as ultramafic cumulates in which 'dunite grades into wehrlite and then pyroxenite'. This description is generally accompanied by idealized sections showing separate successive layers of dunite, wehrlite and clinopyroxenite (Gass *et al.,* 1982; Prichard, 1985; Lord *et al.,* 1994). While this representation is not wholly inappropriate, it hides the full sequence of relationships: wehrlite–clinopyroxenite xenoliths are enclosed in contamination aureoles of clinopyroxene-bearing metadunite suspended in an intrusive metadunite layer. Recently, the obvious absence in the field of continuous layers of wehrlite and clinopyroxenite has been explained as the result of tectonic disruption by NE-trending sinistral faults (Lord *et al.,* 1994). However, the absence of any evidence on the ground or on aerial photographs of such faults, together with the unfaulted continuity of the nearby metagabbro–metadunite boundary (as proved by magnetometery – see above), casts doubt on this interpretation.

Prior to the emplacement of the dunite layer it is possible that the wehrlite–clinopyroxenite masses formed a continuous unit that rested on the mantle and formed *in situ*, as cumulates from an early magma chamber, on the Gass (1980) model. This continuous unit was later disrupted and lifted piecemeal to a higher level, first by the arrival of the intrusive dunite layer and later by the gabbro.

The nature of the geophysical Moho, so clearly revealed in this area, is particularly relevant to the long-running volume problem in serpentinization. Serpentinization of an olivine-rich rock results in a decrease in density of about 20% and, according to petrographic calculations, either an increase in volume of about 45% (*c.* 13% linearly) or a loss of magnesia and silica of about 35% by weight (O'Hanley, 1992). In the past, serpentinization has been accepted as occurring with conservation of volume and loss of material to the surrounding rocks by metasomatism, or with conservation of material and volume increase taken up in bounding faults, with or without internal fracturing. Unst is unique in providing evidence that no expansion has taken place across the serpentinite–metagabbro contact and that no metasomatic transfer has occurred, despite the serpentinization of an olivine-rich rock. The problem remains an enigma.

Conclusions

The Skeo Taing to Clugan GCR site is of international importance in that it provides an accessible and almost continuously exposed coastal section through the geophysical Moho; a traverse across the boundary between the underlying ultramafic rocks and the overlying basic rocks in the ophiolite succession. However, the relationships revealed show that it is intrusive contacts that are involved and not compositional divisions within a continuous series of related rocks formed in a magma chamber. This controversial aspect adds to the overall value of the site in terms of ophiolite research.

QUI NESS TO PUND STACKS (HP 622 032–622 035)

Derek Flinn

Introduction

The site presents a readily accessible view of the upper metagabbro, the highest exposed level of the pseudo-stratigraphy of the Shetland Ophiolite, within the Lower Nappe on the east side of Unst. It is bound to the east by a continuously exposed low-lying cliff section lying within the upper metagabbro. The site extends to the west for nearly a kilometre to include xenolithic screens of wehrlite–clinopyroxenite marking the boundary between the lower metagabbro layer and the upper metagabbro layer; it also includes the western limit of a swarm of basic 'sheeted dyke'-like intrusions (Flinn, 1996). The coastal section provides very good exposures of the latter and shows their relationship to rhythmic banding, minor fine-grained gabbroic intrusions, hornblende pegmatites, quartz-albite ('plagiogranite') veins and lamprophyres, all characteristic of the upper metagabbro layer. The 'sheeted dyke'-like intrusions, although normal to the present erosion surface, are emplaced obliquely to both the petrological and geophysical Mohos and to the metadunite and lower

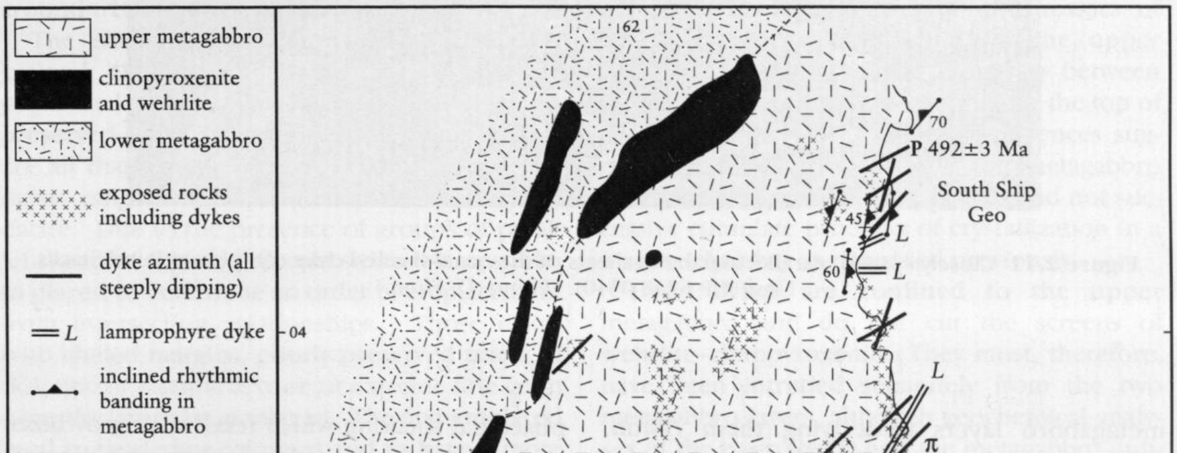

Figure 2.10 Map of the Qui Ness area, Unst.

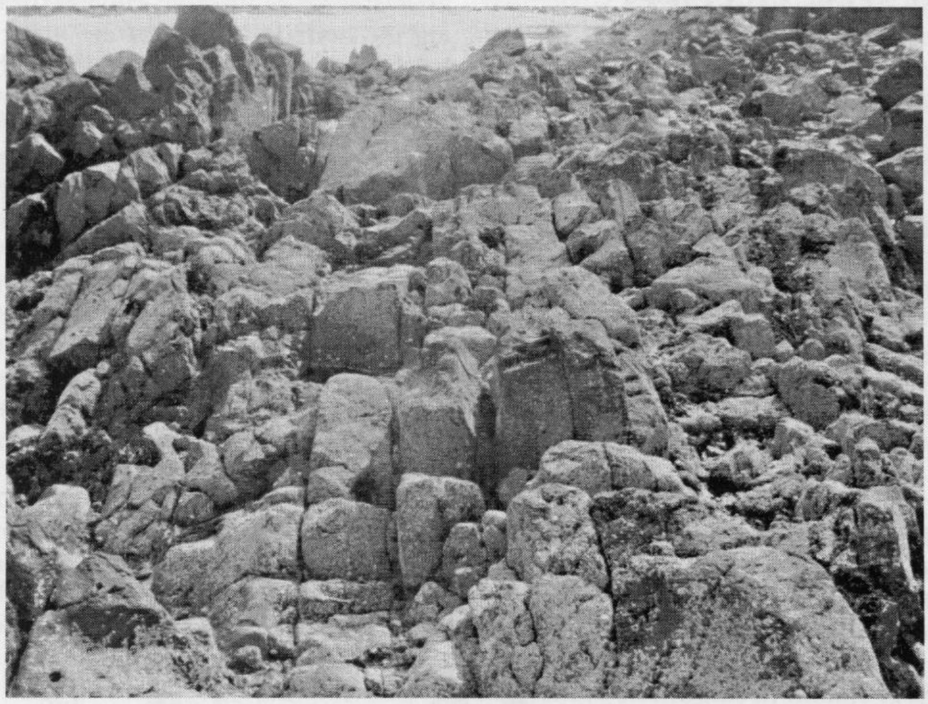

Figure 2.11 Closely spaced, parallel metabasic sheets of the quasi-sheeted-dyke complex at Pund Stacks, Unst (HP 6218 0340). (Photo: D. Flinn.)

metagabbro layers underlying them (Flinn, 1996). In the Gass (1980) model, hitherto used to represent the Shetland Ophiolite (Gass *et al.*, 1982; Prichard, 1985), the dykes are shown as normal to the underlying layers.

Description

The boundary between the lower metagabbro layer and the upper metagabbro layer is not as clearly or sharply defined as the boundary between the lower metagabbro layer and the underlying metadunite layer (see Skeo Taing to Clugan GCR site). Along the east side of Unst it is marked by a discontinuous string of wehrlite–clinopyroxenite screens, lithologically indistinguishable from those above and below the metadunite–lower metagabbro contact, the geophysical Moho. The geological relationships are summarized in Figure 2.10.

The upper metagabbro layer is composed of a poorly contrasted assemblage of metagabbro, fine-grained metagabbro and 'sheeted-dyke'-like metabasic sheets, all subject to phyllitization and phacoidal shearing which tend to destroy lithological boundaries and to reduce the contrast between lithologies. The rocks of the upper metagabbro differ from the lower metagabbro in several respects: amphibole, where it occurs, is generally green rather than colourless and has feathery instead of granular grain boundaries; the plagioclase is commonly sodic and fresh or partly fresh instead of saussuritized; opaque grains with leucoxene/titanite rims are common.

Rhythmic banding occurs locally in the upper metagabbro, where it is texturally similar to the lower metagabbro, but it is of much more variable nature and band thickness than that seen in the lower metagabbro. Thin clinopyroxenitic bands occur locally but the greatest difference is seen at Qui Ness where there are several clearly defined bands of leucogabbro about 1 m thick.

The fine-grained metagabbros occur inextricably intermixed with coarser lower-metagabbro-like rock but, due to similar lithology and phyllitization, they are particularly difficult to distinguish from the metabasic sheets. The latter are best displayed at Pund Stacks (Figure 2.11), although they can be seen intermittently along the coast for 3 km to the north of that locality. Inland, the sheets are much more difficult to detect than on the coast due to lichen cover and generally poorer exposure. At most, the sheets occupy no more than 50% by volume of the outcrop. They are nearly vertical, have north to north-easterly strikes and are less than half a metre wide, discontinuous, and separated by narrow screens of host rock. They are oblique to any rhythmic banding developed in the host rock. None of the sheets cut the wehrlite–clinopyroxenite screens marking the western boundary of the upper metagabbro layer, even where these are well exposed, but west of the upper metagabbro boundary sheets do occur in the gaps between the screens, for example around 610 030.

The 'sheeted' dykes are invariably very fine-grained, highly altered by hydration, and often phyllitized so that chilled margins and rarely preserved traces of doleritic or basaltic texture are all that remain of their original state. The dykes range in composition from boninite to dacite. Due to the presence of groups of parallel sheets of slightly different strike it is possible, in places, to determine an order of emplacement from intersection relationships. Those dykes with chilled margins, poorly preserved traces of doleritic or basaltic texture or extreme fine grain were the last to be emplaced. The first emplaced tend to be darker coloured and to have a more northerly strike than general. Many of these early dykes also have a vertical penetrative schistosity, about 20° clockwise relative to their strike, which results in lenticular dismemberment. This makes them particularly difficult to distinguish from the similarly schistose and lenticularized fine-grained metagabbro components of the upper metagabbro.

Minor components of the upper metagabbro are hornblende pegmatite, lamprophyre and small quartz-albite ('plagiogranite') veins of hydrothermal or pegmatitic appearance. Elsewhere 'plagiogranite' occurs in the form of albite-granite (locally with granophyric texture) or quartz-albite porphyry. Pegmatites containing brown hornblende and altered plagioclase occur as very minor, irregular bodies in the area of Figure 2.10, although larger examples are found elsewhere. The lamprophyres can be seen in places along the coast forming more prominent, continuous sheets than those forming the metabasic, quasi-sheeted-dyke complex.

Interpretation

The upper metagabbro layer has been previously interpreted as the roof region of a high-level magma chamber at a constructive plate margin (Gass *et al.*, 1982; Prichard, 1985). Underplated gabbro was held to have been intruded by fine-grained gabbro and late felsic differentiates, such as hornblende pegmatites and 'plagiogranite'. The presence of swarms of basic sheets was considered to indicate a passage into an overlying sheeted-dyke layer. This formed the roof to the magma chamber, underlying the sea floor, and was either removed tectonically during obduction or was subsequently eroded. However, in terms of this model, it is difficult to explain the presence of xenolithic masses of wehrlite–clinopyroxenite in both the upper metagabbro and across the boundary between the base of the lower metagabbro and the top of the metadunite layer. These occurrences suggest that the metadunite and the metagabbro layers are separate intrusive bodies and not successive cumulate products of crystallization in a magma chamber as proposed previously.

Basic dykes are confined to the upper metagabbro and do not cut the screens of wehrlite–clinopyroxenite. They must, therefore, have been intruded separately from the two metagabbro layers, although geochemical analyses of the basic sheets and the metagabbro indicate that they are similar in composition (Gass *et al.*, 1982; Spray, 1988). Both the petrological Moho and the geophysical Moho in the Lower Nappe are steep to vertical and parallel to one another but the sheeted-dyke-like metabasic sheets in the upper metagabbro, instead of being normal to the Mohos as in the Cyprus model (Gass, 1980; Gass *et al.*, 1982) are more nearly parallel to them. There is no evidence to indicate whether the sheets were intruded before or after the rotation of the mantle, metadunite and lower metagabbro layers into the vertical position that they now occupy in the Lower Nappe in Unst. However, from the similarity of their metamorphic state (greenschist facies) to that of the ultramafic and basic lower layers, basic-sheet intrusion probably took place prior to the metamorphism of the ultramafic and basic layers. It is possible that the slice of crust and upper mantle forming the Lower Nappe slipped and rotated into a recumbent position on a listric fault at the constructive margin prior to the intrusion of the dykes.

Spray and Dunning (1991) obtained a U-Pb age of 492 ± 3 Ma from zircons extracted from a quartz-albite 'plagiogranite' vein at South Ship Geo (Figure 2.10). They considered that the vein was formed at high temperature, before the greenschist facies hydration of the metagabbro, whereas Flinn considers that it is a dacitic sheeted dyke. Either way, the zircon age obtained must be close to the age of formation of the ophiolite. This age is very close to that obtained from the Virva GCR site for the obduction of the Upper Nappe.

Conclusions

This site provides an easily accessible continuous section through the upper metagabbro and quasi-sheeted-dyke complex forming the uppermost exposed level in the Shetland Ophiolite. It reveals the lithologically and structurally complex nature of this layer and its separation from the underlying lower metagabbro by xenolithic screens of wehrlite–clinopyroxenite. This example of ophiolitic-sheeted dykes is the clearest and most extensive available in Britain.

HAM NESS (HP 639 010–634 017)

Derek Flinn

Introduction

The site lies in the SE of Unst where continuous coastal sections and good inland exposure reveal a klippe of serpentinized ultramafic rock (the Mu Ness Klippe) of the Upper Nappe and its basal thrust resting on the Lower Nappe. The Lower Nappe is formed of two tectonically juxtaposed fragments. Rhythmically banded lower metagabbro lies to the south and a fine-grained facies of the upper metagabbro with quasi-sheeted-dyke-like metabasic sheets, quartz-albite ('plagiogranite') veins and lamprophyre dykes, lies to the north (Gass *et al.,* 1982; Prichard, 1985; Flinn, 1996). The site presents an easily accessible view of all these features characteristic of the highest level of the Shetland Ophiolite pseudo-stratigraphy and also illustrates the overthrust nature of the Upper Nappe.

Description

The Mu Ness Klippe is composed of highly sheared and serpentinized ultramafic rock forming the highest parts of the Mu Ness peninsula (Figure 2.12). The serpentinization is of antigorite type and has completely transformed the olivine in the rock so that the protolith of the serpentinite is not immediately obvious in the field. Thin-section examination shows that the klippe is composed of a mixture of metadunite, clinopyroxene-bearing dunite and wehrlite–clinopyroxenite. At the south end of the klippe are chromite prospecting pits and trenches, dug early in the 20th century, the waste from which shows the presence of banded coarse chromite in the dunite (see also The Punds to Wick of Hagdale GCR site report).

The klippe is separated from the underlying metagabbro by a thrust plane, which in many places is clearly exposed and easily accessible (Figure 2.13). At 638 009 it is flat lying and the overlying serpentinite is coarsely sheared parallel to the thrust, imparting a slate-like lamination enclosing sporadic fat, unsheared lenses. At 6388 0068 the shear zone includes a thin lens of blackish phyllite, which could be mistaken for a metasedimentary rock belonging to the Middle Imbricate Zone (Figure 2.4a). In the SE corner of the site, at 637 005, the thrust plane forms an upright, isoclinal synform in the cliffs with the closure only just above sea level. Here the shearing parallel to the thrust is very well displayed in both the serpentinite and the metagabbro.

The metagabbro mass on which the serpentinitic klippe rests is divided into two parts by a steeply dipping, structurally complex zone of gneissose or flaser gabbro. To the south of the flaser zone the metagabbro is of typical lower metagabbro type, both in its possession of the medium-grained, white and green speckled appearance and of the characteristic lower metagabbro rhythmic banding and the healed early brecciation (compare with the Skeo Taing to Clugan GCR site). North of the flaser zone, the rocks have many of the characteristics of the upper metagabbro (compare with the Qui Ness to Pund Stacks GCR site) with variable, fine- to medium-grain size and abundant metabasic sheets of quasi-sheeted-dyke type. The sheets range up to half a metre thick and alternate with screens of host metagabbro in equal volume. The intrusive sheets are particularly well displayed on the beach, at 6758 0160 and around 638 014, where they have been etched out by differential marine erosion, though large boulders interrupt the continuity. A reconstruction

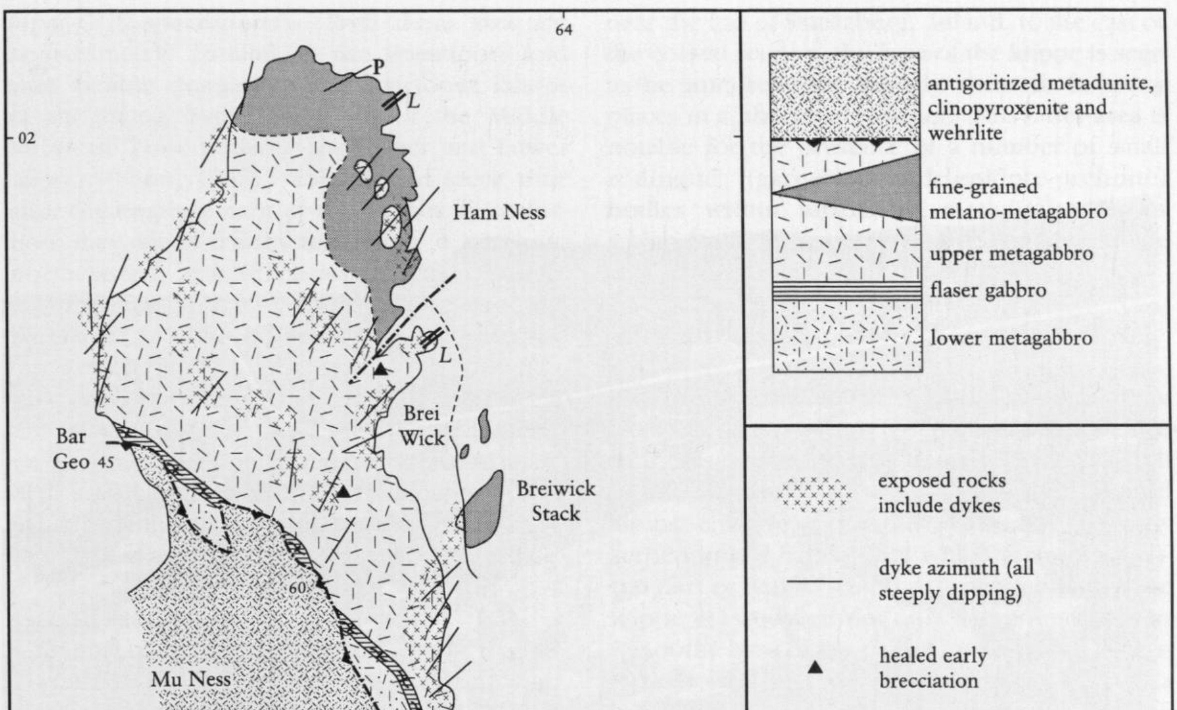

Figure 2.12 Map of Ham Ness and Mu Ness, Unst.

of the occurrence (at 638 014) has been published by Prichard (1985). Inland exposures also show metagabbro intruded by many metabasic sheets but these are more difficult to detect without the enhancing of visual contrast between the intruded rock and the host rock provided by marine erosion.

The north tip and the east side of the peninsula have been intruded by a fine-grained, epidote-hornblende gabbro. A lithologically similar rock forms The Vere, a half-tide rock in the sea 1.5 km to the NNE. This intrusion cuts the

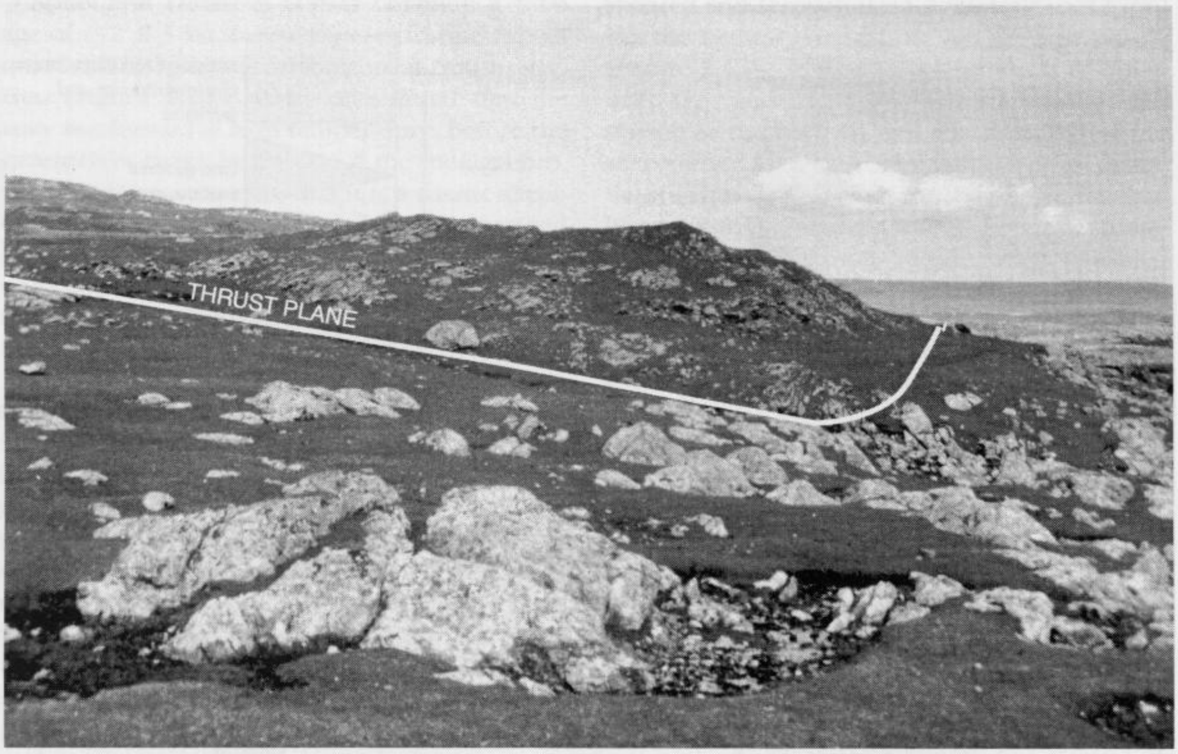

Figure 2.13 The Mu Ness serpentinite klippe (dark coloured) resting on lower metagabbro of the lower nappe (light coloured) as seen from (HP 637 000). (Photo: D. Flinn.)

metabasic sheets, but is itself cut by several similar sheets of similar orientation. It is also cut by a quartz-albite vein ('plagiogranite') and by several lamprophyre dykes.

Interpretation

Within the square kilometre of Mu Ness and Ham Ness a thrust mass of ultramafic rock rests on metagabbro and on fine-grained metagabbro intruded by quasi-sheeted dykes. Three of the four main components to be expected in an ophiolite complex are thus clearly displayed in visible tectonic contact, in an easily accessible environment. The same components are also displayed elsewhere in the Lower Nappe in their natural (non-tectonic) relationship to each other (see Skeo Taing to Clugan and Qui Ness to Pund Stacks GCR sites).

Comparison of the two gabbroic ophiolite components represented in Mu Ness with the larger, in-situ bodies in the Lower Nappe reveals so little difference that the same interpretation may be applied. The serpentinite forming the Mu Ness Klippe is formed of metadunite, clinopyroxene-bearing metadunite and wehrlite–clinopyroxenite, similar to that described in the Skeo Taing to Clugan GCR site but here occurring at the base of the Upper Nappe. It is probably a tectonic slice cut from a level in the ophiolite above the mantle and tectonically emplaced immediately below the Upper Nappe, rather than forming a part of it. Elsewhere in the Shetland Ophiolite the island of Sound Gruney is lithologically and tectonically similar (Figure 2.4).

An upright isoclinal fold of the thrust plane below the Mu Ness Klippe, with a northerly axial trend and an amplitude of about 10 m, is clearly exposed in the cliff in the SE corner of Mu Ness. It is of interest in determining the history of obduction. No obduction can have taken place to either the east or the west in the presence of this synformal downfold as such movement would have had the effect of detaching it from the base of the klippe. The axis of the fold is approximately parallel to, and in line with, the axis of the major downfold of the thrust below the Vord Hill Upper Nappe klippe to the south in Fetlar (see the Tressa Ness to Colbinstoft and

Virva GCR site reports). Both these axes are approximately parallel to the lineations, fold axes, pebble elongations and L-tectonite fabrics of the metasedimentary rocks of the Middle Imbricate Zone between the Upper and Lower nappes (Flinn, 1958). Evidently at some time after the emplacement of the nappes by obduction they were constricted about a north to north-easterly axis, which resulted in folding on this axis and a pebble extension and lineation parallel to it (Flinn, 1958, 1992). The alternative interpretation, that thrusting during obduction took place in a northward direction parallel to the lineation and the fold axes (Cannat, 1989), was based largely on the conventional concept that lineations are parallel to transport directions, but this seems less plausible in the context of the Shetland Ophiolite.

Conclusions

The Ham Ness GCR site provides an easily accessible view of a detached thrust slice (klippe) of ultramafic rock from the Upper Nappe, resting on quasi-sheeted-dyke complex and lower metagabbro. There are abundant inland exposures and a continuous coastal section. This site is of particular importance for the Shetland Ophiolite in that it provides examples of the component ophiolitic units in tectonic juxtaposition and allows their structural relationships to be deduced. It is thus of major importance for any large-scale interpretations of the Caledonian Orogen.

TRESSA NESS TO COLBINSTOFT (HU 612 940–620 949)

Derek Flinn

Introduction

The site lies along the base of the Vord Hill Klippe of the Upper Nappe of the Shetland Ophiolite in Fetlar. The site is bound to the NW by a continuous low cliff exposing the base of the nappe, but not the thrust beneath it. Farther south the thrust is exposed in the coast and passes up the side of Stackaberg where it overlies thin slices of lower metagabbro which separate the nappe from the underlying metasedimentary phyllites of the Middle Imbricate Zone. The latter are also exposed in a window through both the klippe and the underlying metagabbro slices near the top of Stackaberg. Inland, to the east of the coastal section, the base of the klippe is seen to be intensely but crudely sheared, in many places in a phacoidal manner. This latter area is notable for the presence of a number of small rodingite (garnet-diopside-epidote-prehnite) bodies within unsheared metadunite lenses which form enclaves at the sheared base of the klippe (Phemister, 1964; Flinn, 1996).

Description

The site is bound to the west by accessible low cliffs of coarsely sheared serpentinite very close to the east-dipping basal thrust of the Vord Hill Klippe (Figure 2.14). The klippe is composed almost entirely of ochrous-weathering, strongly serpentinized harzburgite which forms the central part of Fetlar. The thrust plane beneath the klippe is associated not only with intense shearing of the base of the nappe but also with recrystallization of the serpentinized harzburgite to a white-weathering antigorite serpentinite (see also The Punds to Wick of Hagdale GCR site). Between Tressa Ness and Colbinstoft the zone of shearing and antigoritization extends about a quarter of a kilometre to the east of the coastal section and is very well exposed (Figure 2.15).

In the coastal section the shear fabric is continuous; it strikes parallel to the coastline and dips variably to the east. The shear fabric varies in nature along the coast from a crudely parallel-spaced cleavage to a coarsely anastomosing foliation enclosing close-packed lenticles of serpentinite. Inland the rocks present much the same appearance, except that larger (up to 100 m long) lenses of non-schistose and unsheared serpentinite are there enclosed in sheared serpentinite indistinguishable from that along the coast.

On the coast south of Easter Gors Geo the thrust marking the base of the klippe runs inland from under the sea. It is steeply dipping and separates the overlying metaharzburgite of the klippe from metagabbro of lower metagabbro type. The metagabbro, in turn, overlies phyllitic metasedimentary rocks of the Middle Imbricate Zone exposed along the coast to the west. The base of the klippe and the underlying metagabbro slices can be traced up the hillside to the summit area of Stackaberg where the thrust becomes horizontal. On the summit of Stackaberg the metaharzburgite forms a group of irregular, closely spaced erosional remnants of the klippe resting on a tectonic sheet of

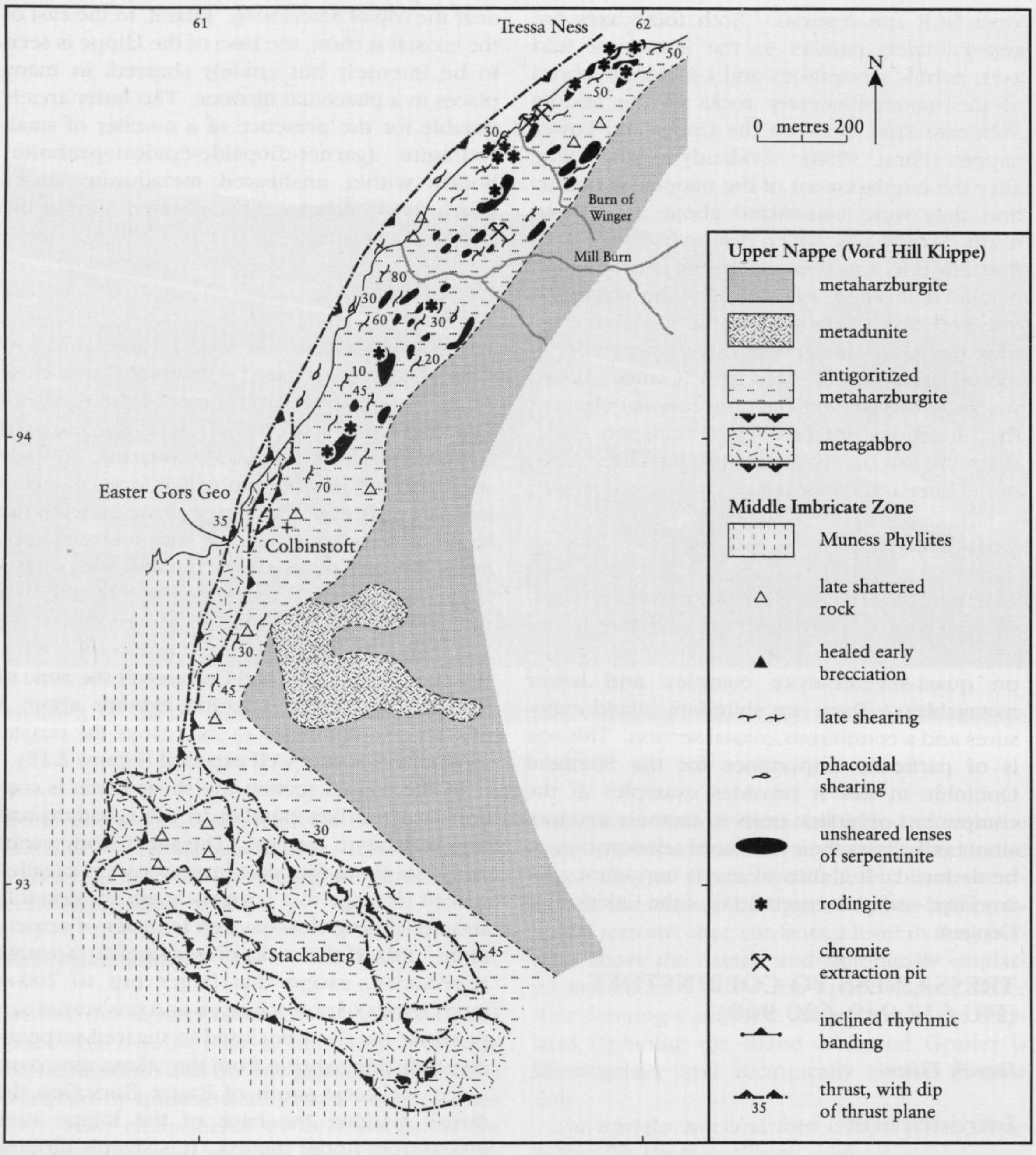

Figure 2.14 Map of the Stackaberg to Tressa Ness area, Fetlar.

metagabbro no more than a few tens of metres thick. In one place a window through the klippe and the underlying metagabbro reveals metasedimentary rocks of the Middle Imbricate Zone.

The most unusual lithological feature of the site is the presence of a number of rodingite bodies in the Colbinstoft to Tressa Ness area (Phemister, 1964). These are closely associated with large unsheared lenses of serpentinite enclosed within the schistosity or cleavage that dominates the base of the klippe. The rodingite occurrences are very irregular and ill-defined, vein-like bodies rarely more than a metre across in any direction. They grade continuously into the adjacent serpentinite, which has a similar greenish colour and surface texture. However, some, and especially the larger ones, have pink or even red cores due to the dominance of

Figure 2.15 Tressa Ness from the south showing the sheared antigoritized harzburgite (dark colour) and the unsheared lenses of antigoritized dunite (light colour). (Photo: D. Flinn.)

macroscopic garnet, for example at 6192 9487. The mineral content of the fine-grained, outer part of the veins can only be determined in thin section; diopside, epidote-type minerals, idocrase and chlorite are the principal constituents. The rodingite veins are not cut by the shearing associated with the base of the nappe. However, it is not clear if this is because the veins post-date the shearing or because they are protected by the unsheared lenses.

Interpretation

The unsheared lenses of serpentinite contained in the sheared base of the metaharzburgite klippe are formed of antigorite serpentinite derived from dunite (Phemister, 1964). Metadunite bodies of equidimensional, irregular and sheet-like form occur in the unsheared ochrous-weathering metaharzburgite nearby, but nowhere else in the Shetland Ophiolite are they more resistant to shearing than the enclosing metaharzburgite. The lenses weather white as a result of the complete antigoritization of the serpentine. In thin section they show no trace of serpentinized pyroxene, which is commonly visible in antigoritized harzburgite, and characteristically they contain more and slightly larger accessory chromite grains than would be expected of harzburgite.

According to Phemister (1964) the rodingite is the result of the metasomatic transformation of veins of gabbro (the protolith of the metagabbro) present within the ultramafic rock, by solutions from the cooling gabbro. He based this conclusion on the close proximity of the metagabbro with the serpentinite and the fact that in the whole of the Shetland Ophiolite rodingite only occurs where these two rocks are in contact. However, there are no 'veins of gabbro' present and the nearby metagabbro is in tectonic contact with the serpentinite, not in igneous contact as concluded by Phemister.

Rodingite occurs in Shetland only in the base of the Upper Nappe in Fetlar, Unst and the island of Sound Gruney to the west of Fetlar. The rodingite veins have the appearance in the field and in thin section of being hydrothermal and metasomatic. They contain diopsidic clinopyroxene (very little Fe or Al), garnet (grossular 50% and andradite 43%), epidote minerals including zoisite and clinozoisite, prehnite, idocrase, chlorite and titanite (Phemister, 1964). According to Phemister (1964) 'the transformation which resulted in the rodingites of Fetlar was induced by residual solutions from the gabbroic rock which were effective in carrying off Si, Al, and Na from the existing rock [gabbro] and possibly added Fe, Ca, and Mg'. Phemister's conclusions are similar to those in vogue today in other parts of the world for rodingite deposits associated with serpentinite (O'Hanley, 1996). However, it must be stressed that in Fetlar there is no evidence for the presence of gabbro veins in the base of the nappe and the gabbro was already consolidated and metamorphosed before it reached its present position below the serpentinite nappe, hence there is no clear gabbroic source of the metasomatizing solutions. Phemister (1964) also concluded that the rodingite veins were formed after the antigoritization (i.e. after the local, secondary serpentinization) and before the shearing of the base of the nappe. No definitive answer seems currently possible and the source of the Ca-rich metasomatizing solutions, the localization of the veins, and the timing of their formation are all problems waiting to be solved.

Conclusions

The Tressa Ness to Colbinstoft GCR site presents a remarkably well-exposed section through the base of the Upper Nappe. It illustrates the tectonics of emplacement of the nappe and the metamorphic and metasomatic processes that have taken place there to form rodingite deposits. The unusual characteristics of the latter provide one of the more enigmatic aspects of the Shetland Ophiolite. This adds to the value of the site since late-stage Ca-metasomatism is a widespread but poorly understood feature of many ophiolite complexes elsewhere in the world.

VIRVA (HU 645 921)

Derek Flinn

Introduction

The site is bound to the east by a continuously exposed sea-cliff section revealing the base of the Vord Hill Klippe, a part of the Upper Nappe, the underlying thrust and the rocks below the thrust. Except at the south end of the site the cliffs are almost entirely composed of sheared and shattered serpentinite forming the cataclastic base of the Upper Nappe. In three places hornblendic rocks belonging to the underlying Middle Imbricate Zone are exposed. Both the hornblendic rocks and the overthrust ultramafic nappe are very well exposed and readily accessible. Hornblende schists are a characteristic feature of ophiolite complexes and are commonly found immediately beneath ophiolite nappes. They are conventionally interpreted as dynamothermal aureoles caused by the obduction of hot mantle nappes over basic volcanic rocks of the sea floor (Williams and Smyth, 1973). This interpretation has been applied to Fetlar by Gass *et al.* (1982), Prichard (1985) and Spray (1988), but has been rejected by Flinn on the basis of field evidence obtained from this site (Flinn *et al.*, 1991; Flinn, 1993).

Description

The sea cliffs northwards from Virva are formed of intensively shattered serpentinite lying immediately above the basal thrust of the Vord Hill Klippe of the Upper Nappe (Figure 2.16). The main body of the klippe to the west, underlying central Fetlar, is almost entirely formed of strongly ochrous-weathering metaharzburgite. However, one-quarter to one-half a kilometre west of the shoreline in Figure 2.16 the weathering colour changes eastward to white and the rocks become increasingly shattered and sheared. The change in weathering colour is due to the recrystallization of ochrous-weathering, relatively iron-rich lizardite serpentine to white-weathering, iron-poor antigorite serpentine and magnetite, a recrystallization that occurs along major shears in the metaharzburgite throughout the ophiolite. On the cliff face the serpentinite is seen to be intensely shattered and is coarsely sheared locally. Continuous fractures, many of which dip to the west, cut through this chaotically arranged assemblage. The condition of the serpentinite varies from a mass of fine powder to coherent blocks a metre or more across.

North of Virva the thrust beneath the serpentinite is submerged beneath the sea at the foot of the cliff and is exposed only at Demptster's Geo. At Virva (Figure 2.17) it is exposed as it passes inland to the south, where its course is marked by a very minor escarpment. The thrust plane is a thin zone of finely crushed and sheared rock a few centimetres thick. In the Virva area the base of the klippe adjacent to the thrust has been sheared into a crushed conglomeratic mass

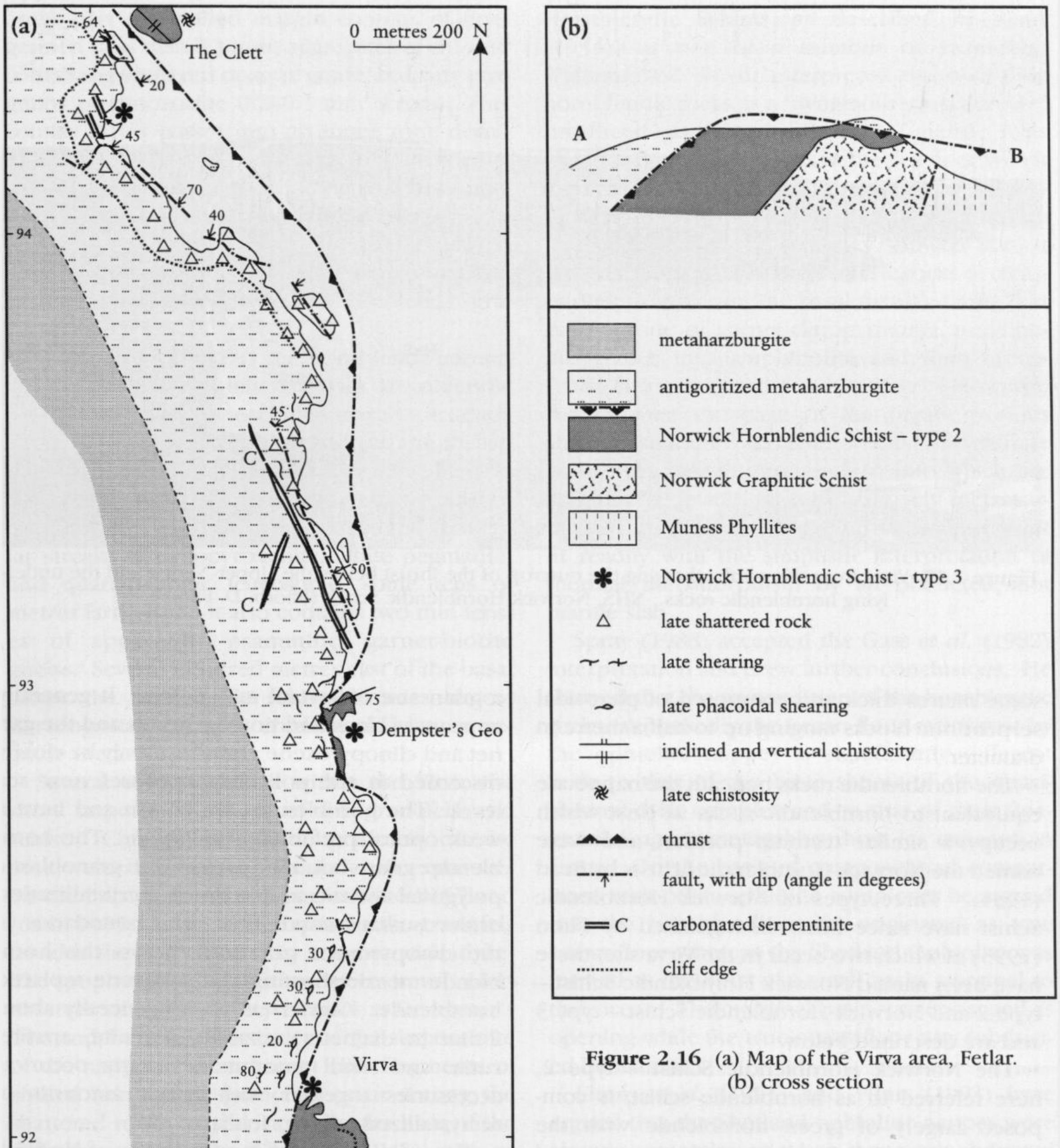

Figure 2.16 (a) Map of the Virva area, Fetlar. (b) cross section

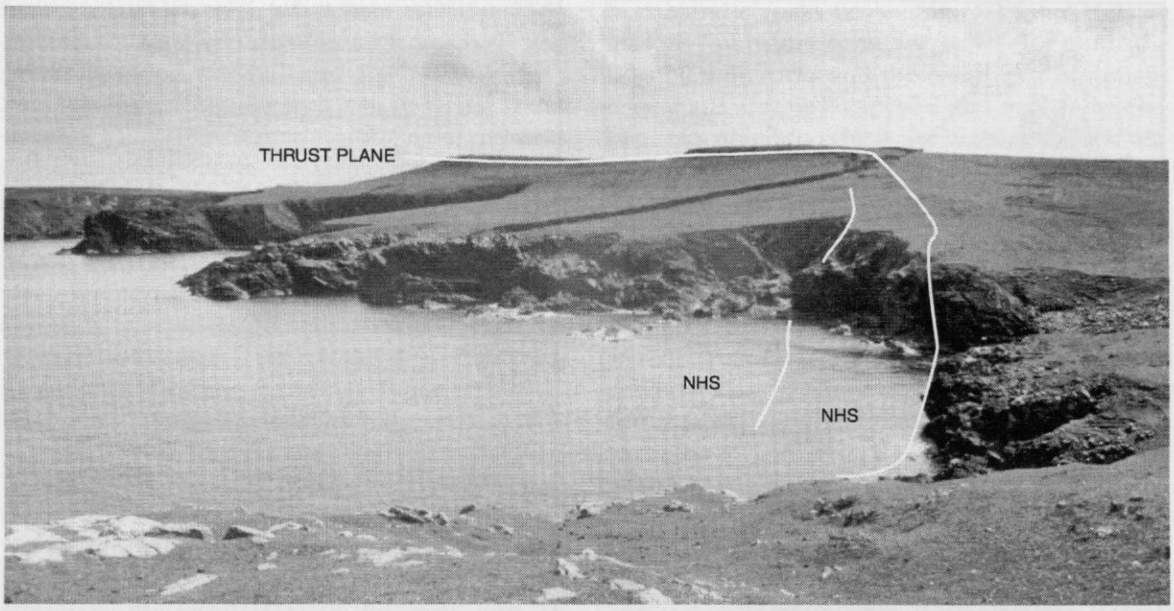

Figure 2.17 Virva from the north, showing the outcrop of the thrust below the Upper Nappe and the underlying hornblendic rocks. NHS, Norwick Hornblendic Schist (Photo: D. Flinn.)

some metres thick and composed of phacoidal serpentinite blocks ranging up to half a metre in diameter.

The hornblendic rocks beneath the nappe are equivalent to hornblendic rocks in Unst which occupy a similar tectonic position and were named the Norwick Hornblendic Schists by Read (1934). Three types of Norwick Hornblendic Schist have since been distinguished by Flinn (1993) of which two occur in the Virva site; these have been named Norwick Hornblendic Schist – type 2 and Norwick Hornblendic Schist – type 3 and are described below.

The Norwick Hornblendic Schist – type 2, here referred to as hornblende schist, is composed largely of green hornblende with the grains elongated parallel to the c-axes and arranged in the plane of the schistosity making them S-dominant tectonites with little sign of a lineation. The maximum grain size is rarely more than one millimetre. The schistosity is often emphasized by closely spaced, feldspar-rich, fine lenticular laminations. The schists contain epidote and chlorite in variable amounts, often associated with low-grade alteration. Titanite is a very common and often abundant accessory.

The Norwick Hornblendic Schist – type 3, here referred to as hornblende granofels, is a hornblende-clinopyroxene-garnet granofels with a grain size of several millimetres. It generally appears as black and homogeneous and the garnet and clinopyroxene grains can only be clearly discerned in the polished exposures near sea level. The hornblendes are brown and have a weak optical preferred orientation. The hornblende grain boundaries form a granoblastic polygonal mosaic indicative of recrystallization under isostatic conditions. The boundaries of the clinopyroxene grains cut across this hornblende mosaic showing that pyroxene replaced hornblende. Garnet (as grains generally about 2 mm in diameter), titanite, ilmenite, apatite, rutile and small amounts of quartz occur as accessories together with minor amounts of recrystallized (albitic) feldspar.

The relationship of the two types of Norwick Hornblendic Schist to the serpentinite nappe are best displayed at Virva. There, Gass *et al.* (1982) and Spray (1988) described a contact between the serpentinite and a clinopyroxene-garnet-hornblende rock (the granofels) which grades east (away from the serpentinite) through hornblende schist into greenschist. As described above, the mineral constituents of the granofels can only be identified in the field at the cliff-foot, where it has been scoured by the sea. Thin sections reveal that in fact the granofels grades to both east and west into chilled margins, a metre or more wide, against the adjacent rocks. In its outer part the chilled margin consists of fine-grained (less than 0.1 mm) aggregates of chlorite containing scattered opaque grains and very rare grains of clinozoisite 0.2–0.3 mm across. This contact facies passes into an inner zone dominated by closely packed anthophyllite needles in a matrix of chlorite but including rounded spots of pure chlorite about 0.3 mm across. This in turn passes through a random mixture of amphiboles, chlorite, epidotes, albite, ilmenite, and titanite, into the high-grade hornblende granofels.

At Virva (6445 9208) the hornblende-garnet-clinopyroxene granofels (Norwick Hornblendic Schist – type 3), with its chilled margins, extends along the coast for some 40 m from the thrust. The hornblende schist (Norwick Hornblendic Schist – type 2) cropping out immediately east of the granofels contains widely scattered lenticular streaks of clinopyroxenite, albite pegmatite, and quartzo-albitic leucosome; some tens of metres farther east it also contains two thin lenses of apparently psammitic, garnet-biotite gneiss. Several hundred metres east of the basal thrust the hornblende schist becomes almost unrecognizable due to crushing and late retrograde metamorphism.

Several other localities within the Virva site show important aspects of the granofels and schist. Chilled hornblende granofels can be seen welded to serpentinite in small exposures on the hillside south of Virva, at 6446 9157 and (6442 9185). At the latter exposure the granofels invades the serpentinite in a metre-wide dyke-like apophysis with a south-westerly trend. Hornblende schist forms the sea stack at Dempster's Geo and, between the stack and the serpentinite of the nappe in the adjacent cliff, fragments of chilled hornblende granofels occur among the shattered rock. Farther north, at the base of the cliff opposite to The Clett (640 943), two or three south-dipping intrusive sheets, less than a metre thick, consist of chilled hornblende granofels. At the southern margin of Figure 2.16, to the south of the coast at Virva, the basal thrust of the klippe can be seen to cut across the Norwick Hornblendic Schist into the underlying graphitic phyllite layer of the Middle Imbricate Zone.

Interpretation

Williams and Smyth (1973) showed that hornblende schists are a characteristic component of ophiolite complexes and used the Norwick Hornblendic Schists, as described by Read (1934), as one of a number of examples. Williams and Smyth interpreted the ophiolitic hornblendic rocks as a 'dynamothermal aureole' produced as hot harzburgitic mantle was obducted over basic rocks of the sea floor which were thus heated and thermally metamorphosed. Gass *et al.* (1982) applied this model to the Shetland Ophiolite. They showed that at Virva the grade of the hornblendic rocks decreases downwards from the basal thrust, stating that 'A thin zone of garnet-clinopyroxene amphibolite grades into amphibolite and then greenschist facies away from the contact'. However, farther west the base of the nappe overlies sheared, shattered and antigoritized serpentinite and to the east the greenschist, into which the amphibolite grades, is retrogressively metamorphosed amphibolite. These relationships do not fit readily with the simplistic interpretation of thermal metamorphism by an obducted, hot mantle slab.

Spray (1988) accepted the Gass *et al.* (1982) interpretation and drew further conclusions. He showed that the composition of the hornblendic rocks (supposedly the ocean floor overthrust by the ophiolite nappe) is significantly different both to that of the basic sheets of the quasi-sheeted-dyke complex and to that of their fine-grained upper metagabbro host (as exposed in Unst). On the basis of compositional comparisons with other ophiolite complexes he argued that the hornblendic rocks originated as true ocean floor whereas the Shetland Ophiolite was formed as the floor of a small basin marginal to that ocean. The marginal basin was envisaged as opening while the true ocean floor was subducted beneath it.

Flinn *et al.* (1991) and Flinn (1993) have argued that the Shetland ophiolitic nappes were already serpentinized when they were obducted and were thus too cool to have metamorphosed basic volcanic rocks of the ocean floor to form hornblende schist. It was also suggested that the protolith of the hornblende granofels was not basaltic oceanic floor but basic magma adiabatically melted from the mantle and segregated in the initial fracture arising from decoupling at the onset of obduction. The hornblende schist may have originated in a similar manner but somewhat earlier so that it served to lubricate the thrust during movement and thus became schistose as it cooled and crystallized.

Spray (1988) reported six K-Ar ages for hornblendes separated from four hornblendic lenses beneath the nappes. The oldest age obtained, 479 ± 6 Ma, was from the Virva occurrence (6445 9208) and the remainder ranged up to 465 ± 6 Ma. Spray interpreted this range of different ages as resulting from diachronous accretion of the hornblende schist during obduction. However, the widespread retrogressive metamorphism in the hornblende schists, leading to partial loss of radiogenic argon, makes these dates unreliable. A more dependable date is the ^{40}Ar-^{39}Ar step-heating age of 498 ± 2 Ma for hornblende separated from a Virva sample of the hornblende granofels (Flinn *et al.*, 1991). In view of the likely low temperature of the nappe when obducted, the hornblende would not have taken long to cool to the closure temperature. This age is therefore regarded as a good estimate of the date of obduction of the Upper Nappe.

Conclusions

The coastal section north of Virva, Fetlar, provides an accessible and extensive view of the basal thrust of the Upper Nappe. This allows examination of the thrust's relationship to the underlying hornblendic granofels and schist. It is a crucial locality in the controversy surrounding the origin of these rocks and their relationship to the obduction of the Shetland Ophiolite and is of international significance. In particular, the site provides evidence that these hornblendic rocks cannot be a part of a thermal aureole formed as hot mantle was thrust over basic volcanic rocks of the ocean floor, following the conventional interpretation of hornblende schists found associated with ophiolites. Instead, a preferred interpretation is that the hornblendic rocks formed from basic magma which was intruded, from deep in the mantle, into the basal thrust of the ophiolite as it was obducted.

THE HIGHLAND BORDER COMPLEX

GARRON POINT TO SLUG HEAD (NO 893 877–886 873)

C. W. Thomas

Introduction

Mafic lavas, mafic and ultramafic intrusions and associated sedimentary rocks of the Highland Border Complex (HBC) are well exposed and readily accessible along the rugged, indented coastline between Slug Head and Garron Point, about 1 km NE of Stonehaven. At this site, the HBC crops out over about 1 km of strike, with an outcrop width of about 200 m, forming the most easterly and best exposed single occurrence of HBC rocks in Scotland. The complex is reverse-faulted against metagreywackes, grits and pelites of the Dalradian Southern Highland Group and is overlain unconformably by Old Red Sandstone facies, terrigenous clastic sediments of the Cowie Formation (Stonehaven Group), which are of Wenlock (Silurian) age.

The HBC rocks at Stonehaven were assigned a possible Arenig age on the first Geological Survey maps of the area (Sheet 67, Stonehaven), but Campbell (1911, 1913) provided the first detailed accounts of the rocks and described palaeontological evidence for their being Late Cambrian in age. The palaeontological evidence for the age of the rocks was based largely on brachiopod fragments found in the black shales intercalated with the lavas in Craigeven Bay. Although the faunal assemblages suggested a Late Cambrian age, Peach and others in the Geological Survey and Anderson (1947) considered the rocks to be Arenig or Caradocian in age, based on comparison with similar faunal assemblages in the Southern Uplands and the Stinchar Limestone of the Girvan district. Doubt was cast on the acceptability of the fossils from Craigeven Bay when Bulman and Rowell concluded that the identification of a comparable fauna from Aberfoyle was unreliable (Institute of Geological Sciences, 1963). However, more recent palaeontological work by Curry and colleagues (Curry *et al.*, 1982; Curry *et al.*, 1984; Ingham *et al.*, 1985) has restored an Ordovician age for at least part of the HBC elsewhere.

Campbell (1913), in collaboration with Peach and others of the Geological Survey, identified the contact between the HBC and the Dalradian rocks as the Highland Boundary Fault, which previously had been placed at the boundary between the HBC rocks and the overlying Old Red Sandstone facies Silurian rocks. They recognized that the Silurian sedimentary rocks are unconformable on the HBC, despite local modification of the unconformity by faulting.

The petrogenesis of the pillow lavas and associated igneous rocks has been studied by Bloxam (1982).

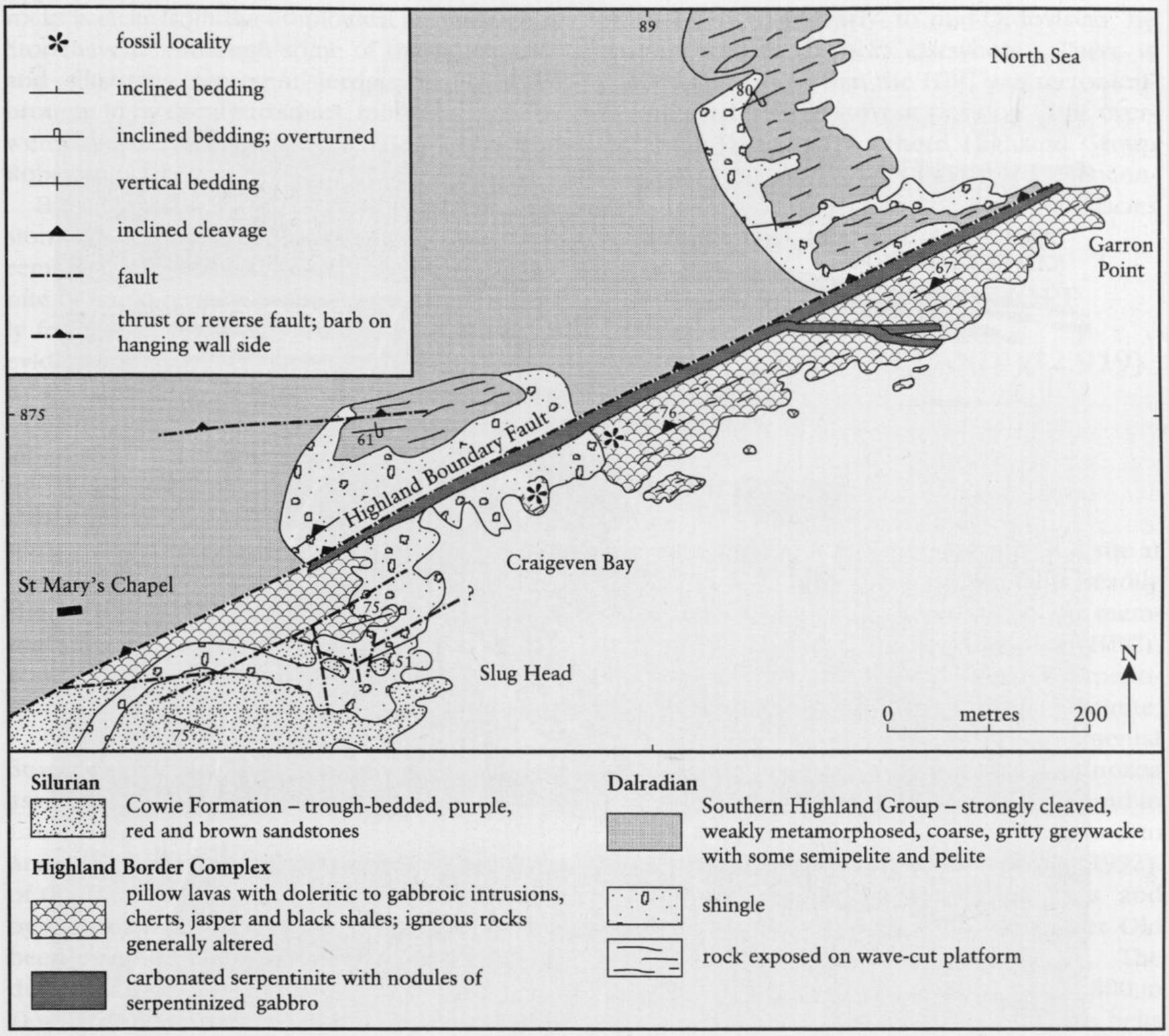

Figure 2.18 Map of the Garron Point to Slug Head area, Stonehaven. Based on BGS 1:10 000 Sheet NO88NE (1996).

Description

The Highland Border Complex is dominated largely by meta-igneous rocks at Stonehaven. Although metamorphosed to lower greenschist facies, extensively sheared and even phyllitic in places, spilitized mafic pillow lavas are generally well preserved (Campbell, 1913; Gillen and Trewin, 1987, plate 17) and clearly show the way-up, younging to the north. The rocks within the complex dip very steeply to the north and the pillow lavas become more intensely sheared and boudinaged northwards towards the tectonic contact with the Dalradian at the Highland Boundary Fault. The geological relationships are summarized in Figure 2.18 and illustrated in Figure 2.19.

The vesicular character of the pillow lavas is still recognizable and variolitic textures are commonly developed. Relict basaltic textures are preserved in places, although the development of epidote aggregates destroys most of the original igneous textures; fresh colourless calcic augite is occasionally preserved. Geochemical evidence from the pillow lavas suggests oceanic tholeiite affinities (Bloxam, 1982). Small, irregular doleritic and gabbroic intrusions in the pillow lavas are common on Garron Point.

A distinctive, orange-coloured, dolomitized and silicified serpentinite occurs along the Highland Boundary Fault and associated splays at Garron Point and is the most northerly unit of

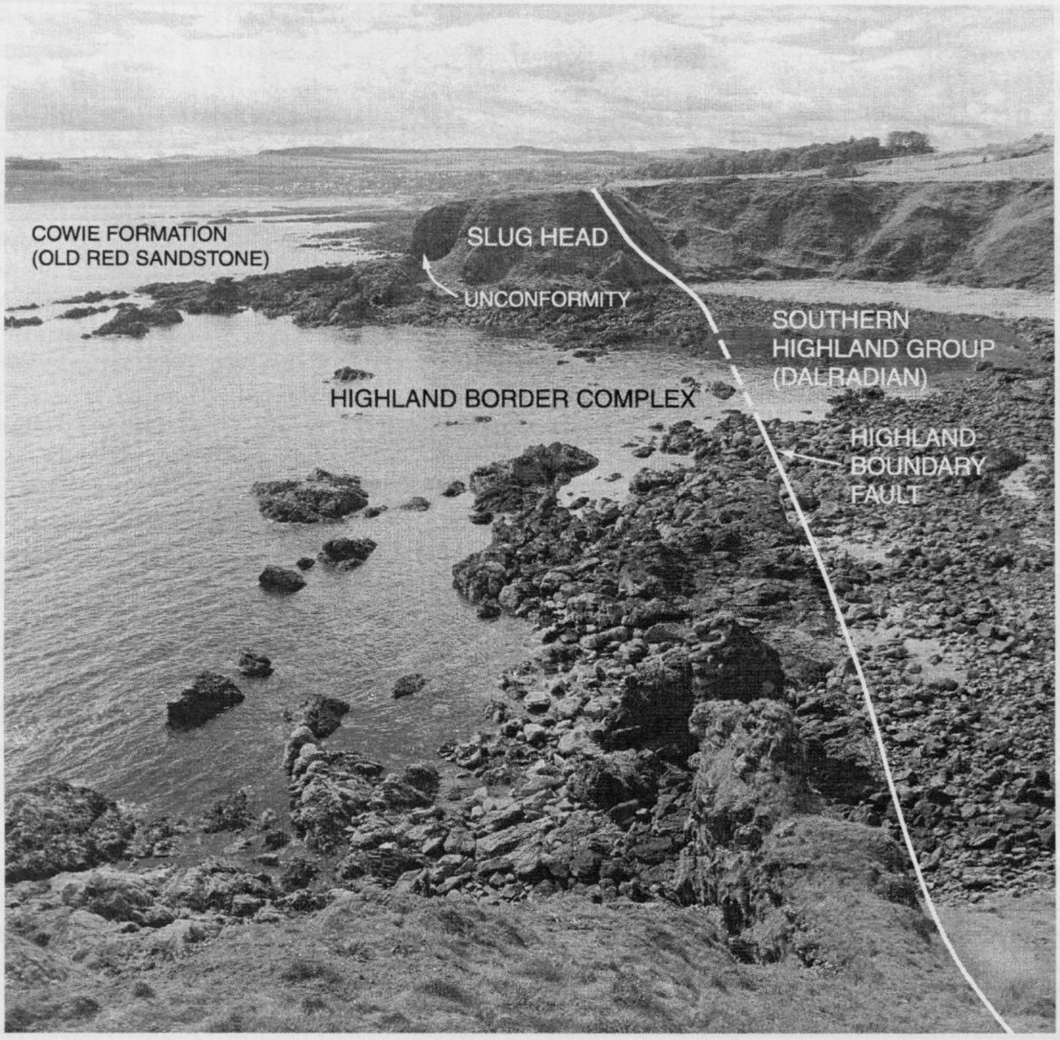

Figure 2.19 The Highland Border Complex at Slug Head and Garron Point looking SW towards Stonehaven. (Photo: C.W. Thomas.)

the HBC at Stonehaven. The serpentinite is well-seen in Craigeven Bay and over the headland on to Garron Point and is up to 15 m thick. The conspicuous colouration is imparted by iron in the dolomite. The serpentinite also contains nodular inclusions of serpentinized gabbro. Bloxam (1982) records that the inclusions are difficult to distinguish from the serpentinite itself, being much more highly altered than the mafic intrusions outside of the serpentinite. The feldspars are now albite and the mafic minerals are largely replaced by hornblende, chlorite and serpentine.

Associated with the lavas and mafic intrusions are red and black cherts, including parallel-laminated black cherts with abundant pyrite, and siliceous siltstones and mudstones (Campbell, 1913; Henderson and Robertson, 1982). These occur as intercalations between flows or as lenticular masses, possibly filling the space between adjacent pillows. It is from these rocks that the faunal assemblage, dominated by brachiopods, was recovered. Grading indicates younging to the north, consistent with the younging indicated by the pillow lavas.

Interpretation

The lithological assemblage comprising the Highland Border Complex between Slug Head and Garron Point represents fragments of oceanic crust formed during the early Ordovician and tectonically emplaced against the Dalradian. The oceanic volcanotectonic setting for the rocks is clear from the lithological assemblage at Stonehaven. Although some of the mudstones and siltstones represent terrigenous material brought in by distal turbidites, most of the cherts were considered to be pelagic (Henderson and Robertson, 1982).

Bloxam (1982) suggested that the mafic inclusions in the serpentinite could be disrupted remnants of sheets intruded into the serpentinite or could have been incorporated tectonically from adjacent rocks. On the petrographical evidence, it appears that both the ultramafic protolith to the serpentinite and the mafic intrusions were serpentinized after the intrusion and albitization of the metamorphic rocks. By analogy with Ballantrae and similar rocks in California, Bloxam (1982) considered that the ultramafic protolith was probably emplaced as 'a low temperature 'mush' and/or solid tectonic slices caught up in the Highland Border *mélange*'. Henderson and Robertson (1982) considered that the whole of the HBC was emplaced tectonically.

The radiometric age of the HBC at Stonehaven remains undetermined. The faunal assemblages elsewhere in the HBC have Laurentian affinities and range in age from Arenig to possibly Llandeilo. However, the age of the HBC at Stonehaven is poorly constrained by the palaeontology and an Ordovician age has been assigned largely by analogy with the age derived from the faunal assemblage in the Dounans Limestone of the Highland Border Complex at Aberfoyle (e.g. Curry *et al.*, 1982; Ingham *et al.*, 1985).

Conclusions

The Stonehaven section provides one of the most extensive outcrops of igneous rocks in the Highland Border Complex, an understanding of which is an essential part of any tectonic interpretation of the Scottish sector of the Caledonian Orogen. This enigmatic assemblage represents remnants of ocean crust, consisting chiefly of metamorphosed tholeiitic pillow lavas with intercalated cherts, siltstones and mudstones. The complex also contains subordinate gabbroic and doleritic intrusions. Associated spatially with these rocks is a carbonated and silicified serpentinite with nodules of serpentinized gabbroic rocks; serpentinization occurred after the intrusion of the gabbros. The age of the HBC at Stonehaven is not determined, but is likely to be early- to mid-Ordovician, by analogy with HBC rocks elsewhere. There is general agreement that the HBC was tectonically emplaced into its current position. It is overthrust by Dalradian Southern Highland Group rocks to the north and is overlain unconformably to the SW by Old Red Sandstone facies rocks of Wenlock (Silurian) age.

BALMAHA AND ARROCHYMORE POINT (NS 419 915, 416 913 AND 412 919)

J. R. Mendum

Introduction

The Balmaha and Arrochymore Point GCR site at the SE end of Loch Lomond contains readily accessible exposures of altered ultramafic members of the Highland Border Complex (HBC). They range from massive and sheared serpentinite, locally silicified, to fragmental serpentinite, typically highly carbonated. Related fragmental rocks with a limestone matrix contain chitinozoa that suggest an Arenig age. The HBC is bound to the NW by the Highland Boundary Fault and to the SE by the Gualann Fault (Bluck, 1992). Precambrian Dalradian metagreywackes and slates occur farther to the NW, and Lower Old Red Sandstone conglomerates to the SE. The ultramafic rocks form two belts, some 800 m apart, termed the Northern and Southern belts (Henderson and Fortey, 1982). They are separated by lithic arenites, part of the 'Loch Lomond Clastics' (Figure 2.20), that are interpreted as a younger component of the HBC. The HBC rocks are overlain unconformably by Upper Old Red Sandstone red-brown conglomerates and sandstones (Rosneath Conglomerate) and just NE of the site these beds overlap the Highland Boundary Fault.

The serpentinites are the oldest exposed components of the HBC, and are probably of early Ordovician (Tremadoc or Arenig) age. They range from massive altered peridotite through to sheared, fragmental, highly carbonated and highly silicified varieties. The adjacent metasedimentary 'Loch Lomond Clastics' show mineralogical and geochemical evidence of derivation from both ultramafic/mafic and quartzo-felspathic units.

The area was mapped by the Geological Survey in the late 19th century and was included

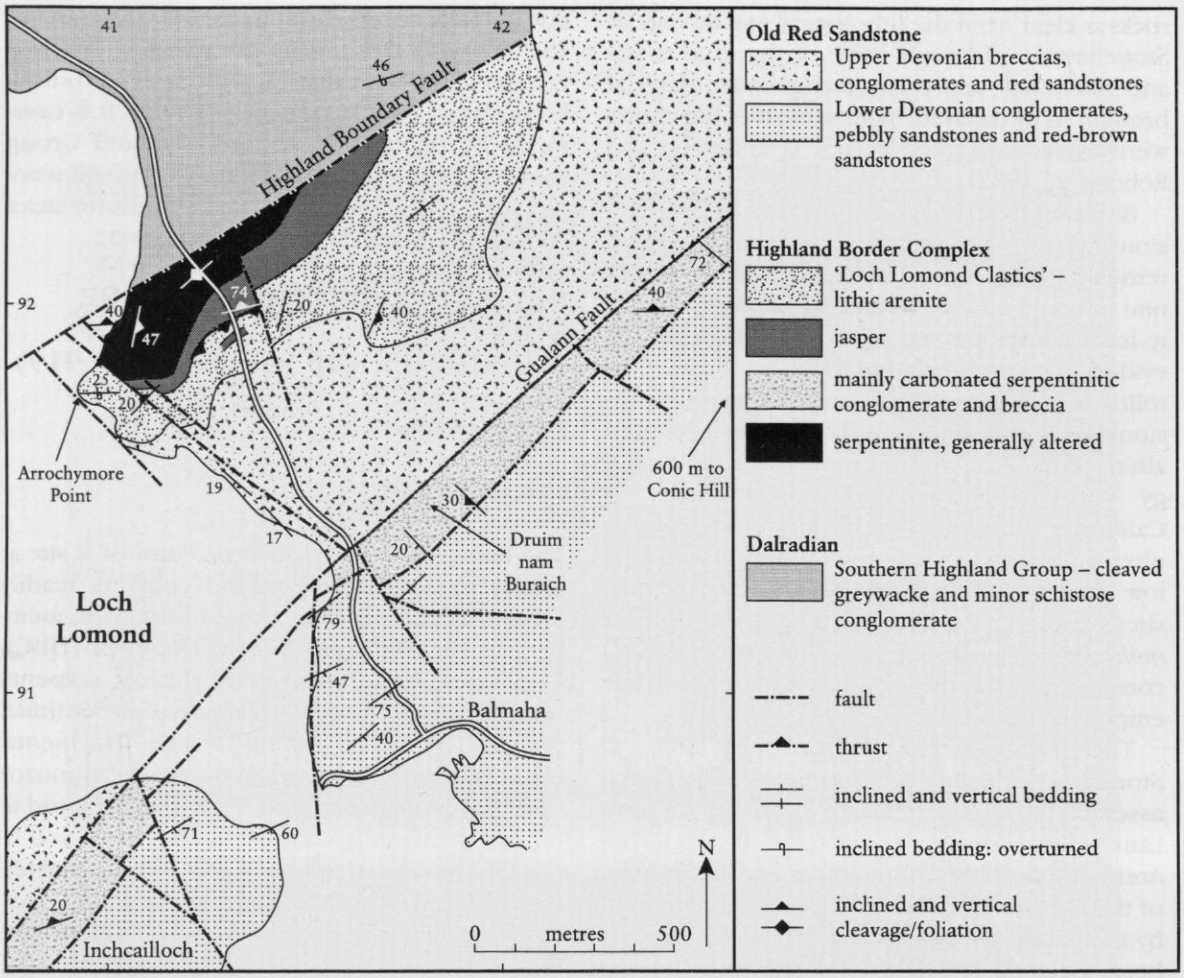

Figure 2.20 Map of the Balmaha and Arrochymore Point area (after Henderson and Fortey, 1982; Bluck, 1992).

in later studies of the HBC by Jehu and Campbell (1917) and by Anderson (1947). More recent work by Henderson and Fortey (1982), Henderson and Robertson (1982), Curry *et al.* (1982, 1984) and Bluck (1992) has described the nature of the HBC at Balmaha and in adjacent areas, notably in the Bofrishlie Burn section and the Lime Craig Quarry areas near Aberfoyle. These authors have discussed the relationships between the various component lithologies of the HBC and placed them in an overall stratigraphical and tectonic context.

Description

The southern serpentinite belt is exposed on Druim nam Buraich on the SW flank of Conic Hill (Figure 2.21). The serpentinite here is extensively altered, recrystallized, brecciated and veined by carbonate and silica. It is dominated by rounded to subangular pebble-like clasts typically 2–3 cm across and up to 15 cm long in places. In the least altered exposures (4254 9212) the clasts lie in a finer-grained (*c.* 2 mm diameter) matrix of altered serpentinite. Henderson and Fortey (1982) described these rocks as serpentinite conglomerate, derived by marine erosion from adjacent exposed intrusive serpentinite. Basic igneous rocks and quartz grains form minor clasts within the rock. Variations in fragment size and alignment of clasts is thought to reflect bedding with units typically about 50 cm thick. In other parts of the outcrop the texture of the rock is more variable and only a crude bedding structure is discernable. Only a weak gently dipping cleavage affects these lithologies. Thin-section studies (Henderson and Fortey, 1982) show that olivine 'mesh' and orthopyroxene 'bastite' textures are preserved in the larger clasts. Lizardite is the only serpentine mineral now present but tremolite, talc and chlorite are the dominant minerals and with further alteration they are replaced by ferroan dolomite and quartz. Chromite occurs both within the clasts and as separate detrital grains.

South-west along strike from Druim nam Buraich, exposures were noted by Bluck (1992) on the shore of Loch Lomond around 4156 9124. Normally below water-level, they comprise distinctive crystalline limestones and conglomerate containing clasts of dolerite, gabbro, spilite and well-rounded quartz arenite. The dolomitic limestone matrix has yielded chitinozoa suggesting an Arenig age. A similar lithology at Lime Craig Quarry near Aberfoyle (15 km to the NE) has yielded a silicified trilobite and ostracod fauna of early Arenig age from the Dounans Limestone. At this locality serpentinite is overlain by a serpentinite conglomerate with a dolomitic limestone matrix, that in turn passes upwards into the Dounans Limestone (Bluck *et al.,* 1984; Curry *et al.,* 1982; Ingham *et al.,* 1985). Similar serpentinite conglomerates are found on the Loch Lomond islands of Inchcailloch, at 4062 9010 adjacent to altered and brecciated basic volcanic rocks and serpentinite, and on Creinch (394 888). They also occur 90 km farther NE in the Highland Boundary Fault zone at Alyth, near Blairgowrie.

The northern serpentinite belt consists of variably foliated and altered serpentinite. Mesh, bastite and vein textures imply that it was initially a peridotite but the only non-serpentinized rock recorded, at a now-obscured locality near Arrochymore Point, is a pyroxenite composed mainly of chromian diopside (Henderson and Fortey, 1982). Antigorite, lizardite and chrysotile are present in the serpentinite seen elsewhere in the Arrochymore Point area. Deep-red altered chromite grains rimmed by opaque iron oxides and green, fine-grained, felted growths of chromium mica, chlorite and clay minerals are ubiquitous in the serpentinite. The foliation, defined mainly by ramifying sub-parallel fractures but locally by a mylonitic texture, is more marked adjacent to the northern faulted contact. The serpentinite was largely foliated prior to alteration to talc-magnesite and ferroan dolomite-quartz but, locally, subsequent deformation has caused the formation of cataclasite in the carbonate-quartz rocks.

Jasper, mainly dark red in colour, lies imme-

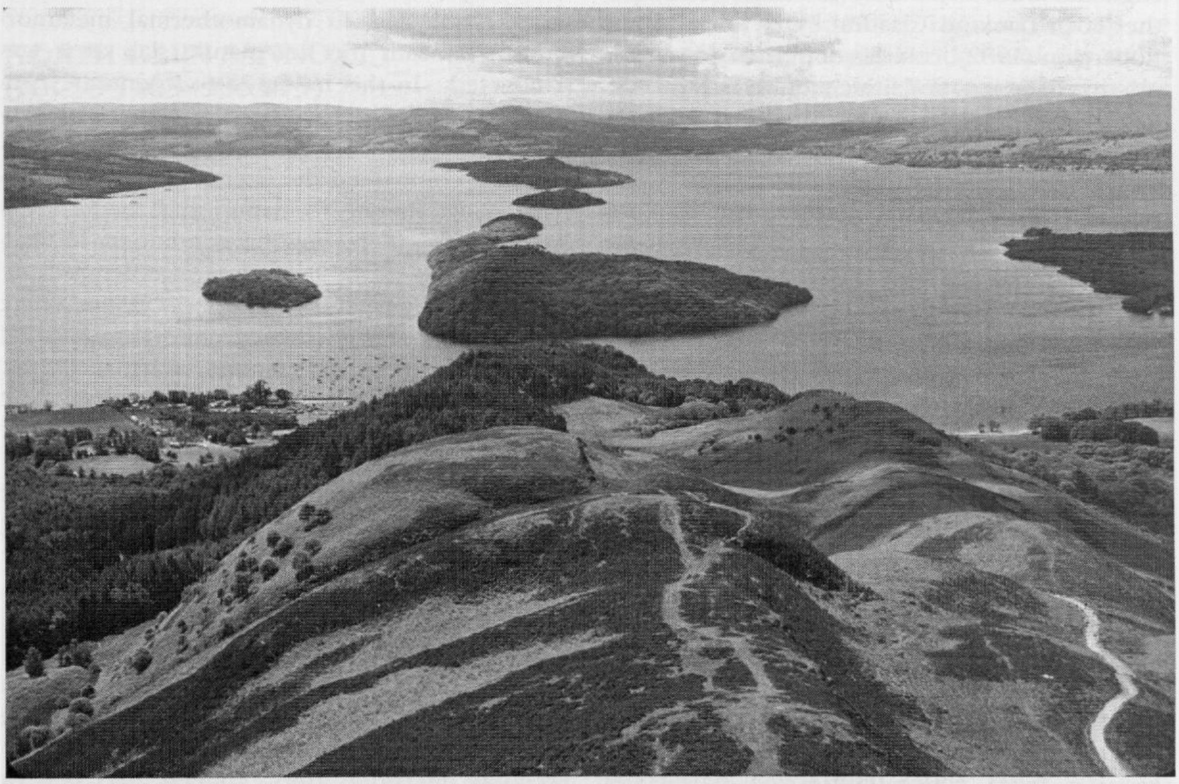

Figure 2.21 Looking SW from Conic Hill across Loch Lomond. Druim nam Buraich (foreground, right of centre) shows exposures of the southern serpentinite belt. (Photo: BGS no. D5406.)

diately SE of the serpentinite in the northern belt, but contacts are not exposed. Henderson and Fortey (1982) note that despite abundant brecciation and quartz veining relict pyroxene and olivine textures are visible, preserved by the fine-grained mosaic quartz. Early tectonic fabrics are also pseudomorphed by quartz and the jasper seems most likely to be the product of pervasive silicification of serpentinite. However, chert is commonly associated with black slates in other parts of the HBC and Bluck (1992) inferred that in these cases it is of sedimentary origin. In respect of the jasper, he commented particularly on the lack of interbedded red or black shales, the dearth of fossils, and the distinctive colour.

Immediately SE of the serpentinite and jasper forming the northern belt, cleaved gritty arenites show grading and immature grains indicative of deposition from turbiditic flows relatively proximal to their source area. The arenites are seen to be thrust over the jasper in the road cut at 4131 9194. Bluck (1992) reported that a soft green serpentinite conglomerate crops out just below the beach gravels on Arrochymore Point at 4095 9183, probably close to the arenite–serpentinite contact. This arenite unit, termed the Highland Border Grits by Tremlett (1973) and the 'Loch Lomond Clastics' by Henderson and Robertson (1982), consists of pink to grey, medium- to coarse-grained lithic arenite. In thin section Henderson and Fortey noted that it is composed of quartz, plagioclase feldspar, biotite, white mica and subsidiary chlorite, carbonate and accessory chromite. It also contains abundant grains and larger fragments of slate, chert, and ultramafic and mafic igneous rocks. Minor fine-grained acid volcanic clasts are also recorded. Development of kaolinite and extensive haematitic reddening was probably a late-stage alteration occurring in Early Devonian times. Cross-bedding is common and near Arrochymore Point, where the beds dip *c.* 25° to the NNW, it clearly shows that they are inverted. The cleavage in the arenites is marked by grain flattening, quartz pressure shadows and phengite mica trails. Henderson and Fortey (1982) noted that the angle between bedding and cleavage is very small so that the facing direction cannot be determined by that means. Bluck (1992) suggested that these rocks are Caradocian in age, correlating them with the similar Aberfoyle arenites from which chitinozoa have been obtained (Bluck *et al.*, 1984).

Interpretation

The Arrochymore Point outcrops are of massive, to foliated, serpentinized harzburgites and minor clinopyroxenites, commonly altered to talc-magnesite and ferroan dolomite-quartz assemblages. Pervasive silicification has altered part of the serpentinite to jasper. Farther south, the outcrops on Druim nam Buraich, near Balmaha, are resedimented serpentinite conglomerate, now almost entirely replaced by ferroan dolomite and quartz. Elsewhere these conglomerates apparently pass upwards into bedded limestones, although only limestone conglomerates are seen at Loch Lomond.

The serpentinite and serpentinite conglomerate outcrops are interpreted as the basal units of the HBC. They represent small fragments of sub-ocean-floor mantle and the detrital deposits formed by its erosion, under marine conditions, immediately following its thrust emplacment at the surface. The serpentinite assemblage is thought to be either Late Cambrian or early Ordovician in age. Basic intrusive rocks, now amphibolites, are associated with the ultramafic bodies near Aberfoyle and at Scalpsie Bay on Bute (Henderson and Robertson, 1982). They may represent original ocean-floor volcanic rocks subjected to dynamothermal metamorphism beneath the hot mantle slab as it was obducted. In this model the serpentinite conglomerates could have been generated by erosion at the active thrust front. Limestone deposition accompanied the later stages of serpentinite conglomerate formation and then continued under quieter marine conditions, in shallower water, as obduction was completed.

The overlying 'Loch Lomond Clastics', here represented only by lithic arenites, are a varied sequence of conglomerates, gritty arenites, shales and limestones. Robertson and Henderson (1984) used geochemical evidence to constrain the possible origin of the arenites. Analysis of a specimen of cleaved arenite from 200 m east of Arrochymore Point (4091 9178) showed high nickel (561 ppm) and chromium (1234 ppm) values in a rock with 77.7% silica. They concluded that the 'Loch Lomond Clastics' had been derived from erosion of ultramafic, mafic and quartzo-feldspathic rocks and hence deduced that they stratigraphically overlie the serpentinite, serpentinite conglomerate and succeeding limestone units. Indeed, they may be considerably younger than the serpentinite conglomerate and are possibly of Caradocian age.

Conclusions

The outcrops of the Highland Border Complex near Balmaha are of crucial importance in understanding its genesis. They provide examples of the oldest exposed parts of the complex, the serpentinite and serpentinite conglomerate, and reveal something of their relationships with overlying clastic strata. The serpentinite is generally regarded as of Late Cambrian or early Ordovician (Merioneth–Arenig?) age whereas the sedimentary rocks may range up to the late Ordovician (Caradoc?). Both the serpentinite bodies and adjacent clastic strata are fragments of the oceanic floor and sedimentary fill of small basins formed at the Laurentian continental margin. Much of the complex was emplaced both during collision of a large volcanic arc with the Laurentian margin in late Llanvirn times (*c.* 470 Ma), and by subsequent transcurrent faulting in the late Ordovician and Silurian. Some of the mafic and ultramafic elements may have been obducted during an earlier stage of this complex tectonic history.

NORTH GLEN SANNOX (NR 995465–995475)

D. J. Fettes

Introduction

The North Glen Sannox site presents an excellent section through the lava and black shale sequences of the Highland Border Complex (HBC) in north Arran. It illustrates one of the best and most varied developments of basic volcanic rocks within the complex. Downward-facing structures provide critical evidence for the study of the structural setting of the complex and its relationship to the Dalradian succession. The geology of the site is summarized in Figure 2.22.

Within this site the HBC outcrop forms a N–S strip that is about 2.5 km long, varies in width from 200 to 400 m and is broadly parallel to the regional strike. It is bound to the west by the turbiditic sandstones of the Dalradian Southern Highland Group and to the east by the Highland Boundary Fault and the Lower Old Red Sandstone succession (Figure 2.22). The strip is cut off to the south by the Palaeogene Northern Granite and to the north by the Highland Boundary Fault, which gradually transgresses northwards across the strike. The complex is composed predominantly of lavas with subordinate black schists, cherts and sandstones. Biostratigraphical control is poor but indicates an Arenig age.

The complex was first mapped by the Geological Survey around the beginning of the 20th century and was reported on in resultant memoirs by Gunn *et al.* (1903) and Tyrrell (1928). The area has been the subject of considerable subsequent study, notably by Anderson and Pringle (1944), Johnson and Harris (1967) and Henderson and Robertson (1982), most of whom were largely concerned with the structural setting of the HBC rocks and their relationship to the Dalradian succession.

Description

Excellent sections are provided through the Highland Border Complex outcrop by the North Sannox Burn (Figure 2.23) and by scattered exposures on the adjacent hillsides. In the burn, turbidites of the Southern Highland Group are exposed above the road-bridge and for about 130 m downstream to their contact with the HBC rocks. In the upstream part of this section massive pebbly sandstones predominate with quartz and feldspar clasts, but flaggy sandstones, siltstones and slates become more common downstream (towards the east). The bedding strikes *c.* NNE and is generally vertical or steeply dipping towards the east. Good evidence from graded beds indicates a consistent sense of younging, also to the east, in which direction the Dalradian strata are succeeded by the HBC assemblage. This may be divided into five units that comprise, from west to east:

1. black shales and cherts,
2. lower lava sequence,
3. black shales and cherts,
4. upper lava sequence,
5. green slates and sandstones.

Gunn *et al.* (1903) originally ascribed an Arenig age to these rocks on the basis of their similarity to the rocks at Ballantrae. Subsequently Anderson and Pringle (1944) reported a find of

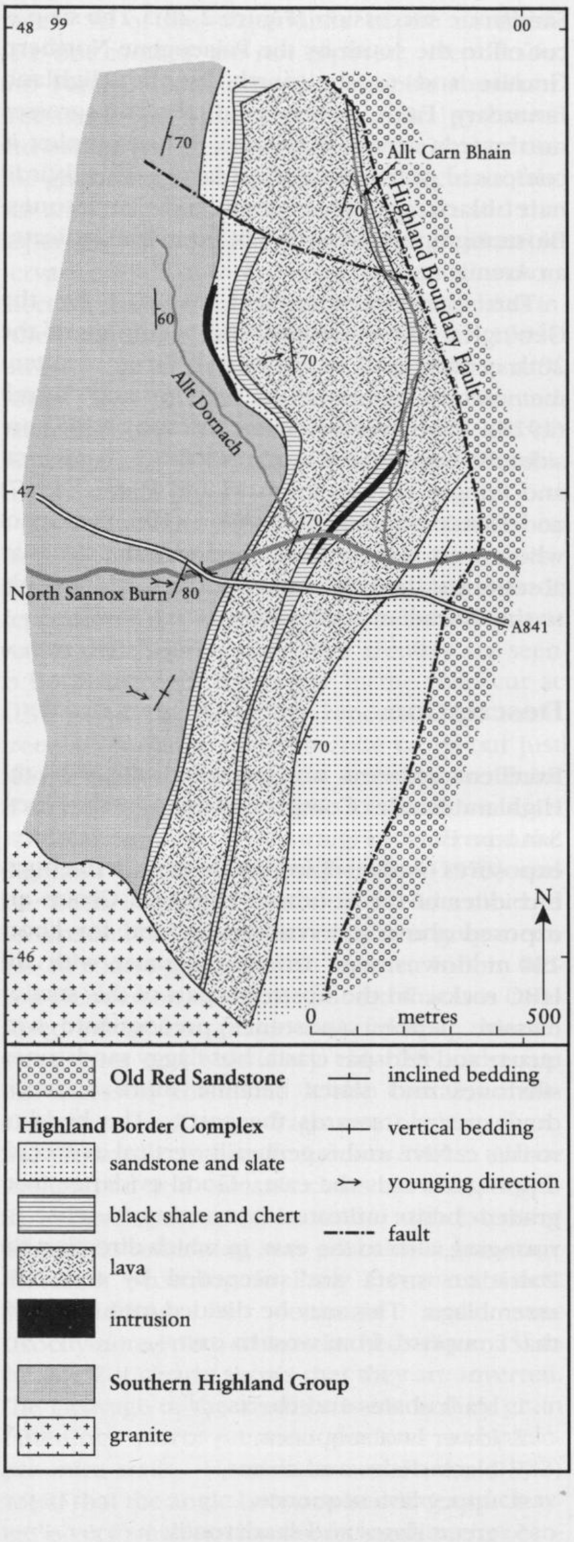

Figure 2.22 Map of the North Glen Sannox GCR site, in part after Johnson and Harris (1967) and Anderson and Pringle (1944).

some poorly preserved brachipods from the central black shale unit. The shells were referred to *Acrotreta* and compared with *A. nicholsoni* described from Aberfoyle by Jehu and Campbell (1917). Curry *et al.* (1984), although unable to trace the original specimens, accepted the age albeit with some caution.

The contact between the Dalradian and HBC rocks is exposed on the south bank of the burn but is only readily examined when water levels are low. Anderson and Pringle (1944) placed the contact at the junction between fine-grained sandstones and black shales. They reported that there is no evidence of interbedding and, although the shales are disrupted, there is no sign of significant faulting; the disruption being ascribed to 'slight movement' at the contact during the regional folding. Eastwards of the contact the black shales continue for a few metres followed by a heavily brecciated and carbonated lava which may represent an altered autobrecciated flow. This is succeeded by a further 2–3 m of black shale with chert bands.

Above the black shales lies the lower lava unit, which is here over 100 m thick and forms the first major development of volcanic rocks in the section. The lavas are greenish-grey, spilitic, commonly pillowed and locally contain zones of autobrecciation; the latter were described by Henderson and Robertson (1982) as 'clast-supported lava rudites, composed of angular to subrounded clasts 0.05 to 0.25 m in diameter'. The pillows have chilled margins and concentric amygdaloidal zones. Johnson and Harris (1967) described almost undeformed pillows from Torr na Lair Brice with flat bottomed bun-shapes, pear-shapes and wedge-shapes all indicating younging to the east. Henderson and Robertson (1982) drew attention to the absence of intercalated sedimentary units in contrast to the basic volcanic rocks in the HBC in the North Esk and at Stonehaven.

The succeeding black shale unit is about 40 m thick and contains numerous chert beds and some beds of coarse-grained sandstone which may in part be volcaniclastic. It was from this unit, where it crops out in the Allt Carn Bhain (Figure 2.22), that Anderson and Pringle (1944) recovered the brachiopods.

The upper lava unit is up to 170 m thick and broadly similar to the lower unit although pillow structures are less well developed. Locally the rock is strongly cleaved and particularly towards its top the unit becomes so 'schistose' that Gunn

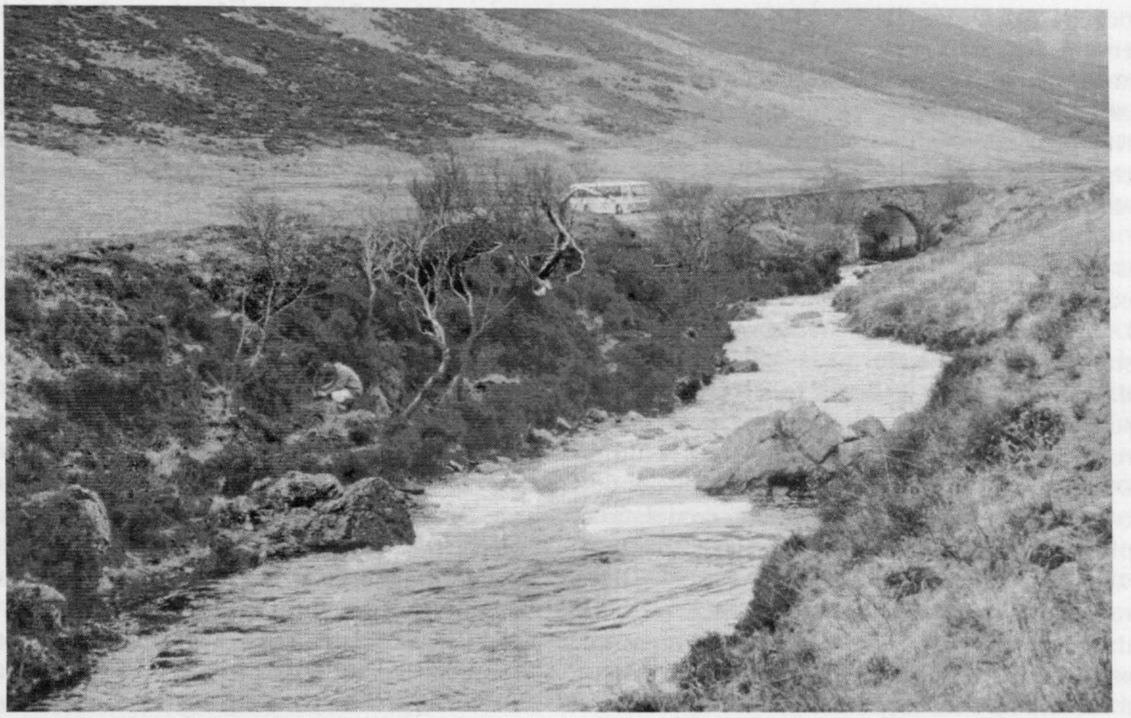

Figure 2.23 North Sannox Burn below the road bridge, a section through the lower part of the Highland Border Complex. (Photo: Nature Conservancy Council.)

et al. (1903) compared its appearance to that of the Green Beds in the Southern Highland Group of the SW Highlands.

Above the lavas and forming the easternmost unit of the complex is a sandstone and slate unit broadly similar to the Southern Highland Group turbidites. In the west the unit is dominated by green slates and phyllites and the sequence becomes more sandstone rich towards the Highland Boundary Fault. Graded beds indicate an easterly sense of younging. Although Gunn *et al.* (1903) described apparent interbedding between slates and lavas at the westerly contact of the unit, this interpretation was strongly challenged by Anderson and Pringle (1944) who argued against a transitional contact in favour of a high-angle fault.

Gunn *et al.* (1903) reported intrusive basic rocks at two localities. The first lies at the base of the HBC succession *c.* 450 m north of the burn, the second crops out in the Allt Carn Bhain. Anderson and Pringle (1944) subsequently identified a third occurrence of a 2.5 m band within the central black shale unit, exposed on the south bank of the burn. They suggested that this outcrop may be an along-strike correlative of Gunn *et al.*'s (1903) second locality. The intrusive rocks appear to be altered hornblende gabbros which, although still retaining ophitic textures, now exhibit carbonated assemblages of uralitized amphibole and albitized plagioclase.

The petrology of the lavas and the associated intrusions has been studied by Ikin (1983) and Robertson and Henderson (1984). The lavas still display igneous textures although the mineralogy has been almost wholly altered with the original olivine-pyroxene-calcic plagioclase associations now represented by albite-actinolite-chlorite-epidote assemblages. The chemistry of the rocks has been significantly affected by spilitization with strong Na-enrichment, Ca-depletion and substantial movement of many mobile elements. A study of the immobile elements, however, allowed Ikin (1983) to classify the lavas. He recognized two groups: the first covers the greater part of the lower lava sequence and has affinites to alkali basalts characteristic of within plate settings; the second, which covers the uppermost part of the lower lava unit, the whole of the upper sequence and includes the intrusions, has tholeiitic affinities and mid-ocean-ridge basalt (MORB) characteristics. Robertson and Henderson (1984) noted that the MORB chemistry is characteristic of the basic volcanic rocks developed within the HBC along the Highland edge, whereas lavas with alkali basalt chemistry are only known from Arran and Aberfoyle.

The bedding in the Dalradian and HBC rocks is congruent and dips steeply to the east with a constant sense of easterly younging. In the Dalradian succession a penetrative cleavage (S1) strikes broadly parallel to bedding and dips to the east at moderate angles to give a downward sense of facing. Johnson and Harris (1967) reported that D1 folds are rare. They recorded D2 structures evidenced by the presence of fine crenulations in the slate bands; the crenulations plunging at moderately steep angles to the SSW. Within the HBC Johnson and Harris (1967) recorded a similar structural sequence to that in the Dalradian rocks with a penetrative downward-facing S1, tight or sub-isoclinal D1 minor folds and rare D2 crenulations.

Ikin (1983) noted the presence of two high-strain zones of 'mylonitized' lavas and associated breccias, approximately at the base of the two lava units. Henderson and Robertson (1982) also described the presence of high strain zones marked, in part, by mylonitized spilites; they noted that the high-strain fabrics post-date asymmetric folds at a locality to the north of the current site, but that the fabrics themselves are corrugated by minor folds.

Interpretation

Gunn *et al.* (1903) regarded the Highland Border Complex as an integral part of the 'metamorphic series of the Highlands', remarkable only for its unusual lithologies. This view was largely supported and expanded by Anderson and Pringle (1944) and Anderson (1947) who argued that the Dalradian and HBC rocks had been subjected to the same history of deformation and low-grade metamorphism. They regarded the junction between the Dalradian and HBC rocks as a modified 'disconformity or slight unconformity' and the eastern sandstone unit as a down-faulted block of Dalradian rock. They concluded that the Dalradian succession is of pre-Arenig age and that the folding and metamorphism was post-Arenig. They further argued that the lavas had been extruded on to a subsiding surface of Dalradian rocks and are broadly comparable with the ophiolitic sequences of the Alps and elsewhere.

Johnson and Harris (1967), influenced by the strong structural congruity between the Dalradian and HBC successions, concluded that 'such would appear to be conclusive proof that the ?Arenig has been involved in the same sequence of movements as the adjacent Dalradian'. However, Ikin (1983) and Henderson and Robertson (1982), impressed by the presence of high-strain zones within the HBC succession, argued for tectonic emplacement of the complex. The latter authors further argued for 'the HBC outcrop in Arran as a thrust-stack with zones of recrystallized Dalradian and schistose rock at the sole of each thrust sheet'. They stated that, 'because cleavages cannot be traced continuously from the HBC into the Dalradian, we cannot correlate the cleavages in the two units'. They noted the presence of downward-facing structures in the HBC rocks and argued that the thrusting occurred before the Dalradian D3 event, probably during D2. Robertson and Henderson (1984) were less categorical about the type of emplacement suggesting a general strike-slip regime. They regarded the eastern arenite-dominated sequence as part of the HBC succession rather than as a faulted block of Dalradian rocks (Robertson and Henderson, 1984, table 1). A current view of the regional development of the Highland Border Complex (for example, Robertson and Henderson, 1984; Curry *et al.,* 1984) envisages the formation of the HBC in an oceanic basin marginal to a continental mass and subsequent tectonic juxtaposition with the Dalradian. Tanner (1995) however, in a discussion of the status of the HBC, emphasized the congruity of the younging evidence and minor structures between the Dalradian and HBC successions and concluded that the greater part of the HBC (at least including the Arenig rocks) was deposited in sequence with the Dalradian rocks.

On a regional scale the available biostratigraphical and radiometric age evidence for the formation of the HBC and the tectonism in the Dalradian rocks implies that the juxtaposition of the two successions could only have taken place after the initial Dalradian deformation events. It is these factors that are in conflict with the apparent congruity of the structural sequences in the Arran section. This dilemma has driven much of the recent research in the area and gives an added regional significance to the site.

Conclusions

The North Glen Sannox GCR site contains an excellent section through one of the major developments of basic volcanic rocks in the Highland Border Complex. It is of great importance in assessing the role of this enigmatic assemblage in terms of Caledonian orogenesis and terrane assembly. The lavas exhibit good pillow structures indicating eruption under water. Associated black shale and chert sequences have yielded fossils which suggest an Arenig age. These assemblages are characteristic of rocks generated on the ocean floor. Although early workers did not recognize a significant unconformity between the HBC and the adjacent Dalradian rocks, the subsequent recognition of highly sheared (mylonitic) zones within and at (or near the base of) the HBC, suggests that it was tectonically juxtaposed with the Dalradian early in the deformational history of the two assemblages. This may, however, be difficult to reconcile with regional evidence on the relative ages of the Dalradian and HBC deposition and deformation.

THE BALLANTRAE COMPLEX

BYNE HILL (NX 180 945)

P. Stone

Introduction

Byne Hill is formed by an intrusive body ranging in composition from gabbro at the margins, through dioritic lithologies to a leucotonalite core; the gabbroic margin is chilled against the host serpentinized harzburgite. The leucotonalite, one of the key components in the interpretation of the Ballantrae Complex as an ophiolite, was originally described as an oceanic 'trondhjemite'.

The site lies at the northern margin of the Ballantrae Complex where a composite intrusive body cuts the Northern Serpentinite Belt and forms a ridge of high ground running SW for about 4 km from the northern flanks of Byne Hill (185 950) to the south-western spur of Grey Hill (158 924). The majority of the ridge is composed of gabbro which is chilled against the host ultramafic rock, mostly serpentinized harzburgite. This intrusive and chilled relationship can be seen at intervals along the SE side of the ridge; the NW margin of the intrusive mass has been faulted so that the igneous rocks are juxtaposed against either volcanosedimentary components of the Arenig ophiolite complex or younger Llandeilian boulder conglomerate (the Benan Conglomerate Formation of the Barr Group). The intrusive rock becomes progressively more leucocratic away from the chilled (SE) margin so that the gabbro passes transitionally into diorite, quartz-diorite and ultimately into 'trondhjemite', a sodium-rich leucotonalite consisting of quartz and plagioclase (now mainly albite but originally more calcic) with accessory hornblende and/or biotite. The most complete section through this transition is preserved on the SE flank of Byne Hill, the summit of which is composed of leucotonalite. A useful field guide to the area is provided by Bluck and Ingham (1992).

The definitive work on the Byne Hill igneous assemblage is the petrological and geochemical study by Bloxam (1968). However, the passage of gabbro into diorite and granite was first noted by Peach and Horne (1899) and the term 'trondhjemite' was first applied by Balsillie (1932) in recognition of the complete absence from the granite of K-feldspar. Following the detailed study by Bloxam, the presence of 'trondhjemite' in the Ballantrae Complex was identified by Church and Gayer (1973) as a key factor in its interpretation as an ophiolitic assemblage of obducted oceanic crust. The 'trondhjemite' has also been pivotal in dating the complex, providing a well-defined, U-Pb zircon age of 483 ± 3 Ma, taken by Bluck *et al.* (1980) as the time of crystallization of the leucotonalite magma. In modern usage leucotonalite is preferred to trondhjemite.

Description

The outline geology of the Byne Hill area is summarized in Figure 2.24. The host rock to the intrusive body is serpentinized harzburgite of the Northern Serpentinite Belt; along the SE flank of the hill the intrusive contact is approximately vertical but it is offset by numerous minor faults. A dark-green to black, massive appearance is characteristic of the serpentinite, which was originally mostly olivine, with sporadic, lustrous bronze-coloured relics after orthopyroxene. Within the serpentinite are pale

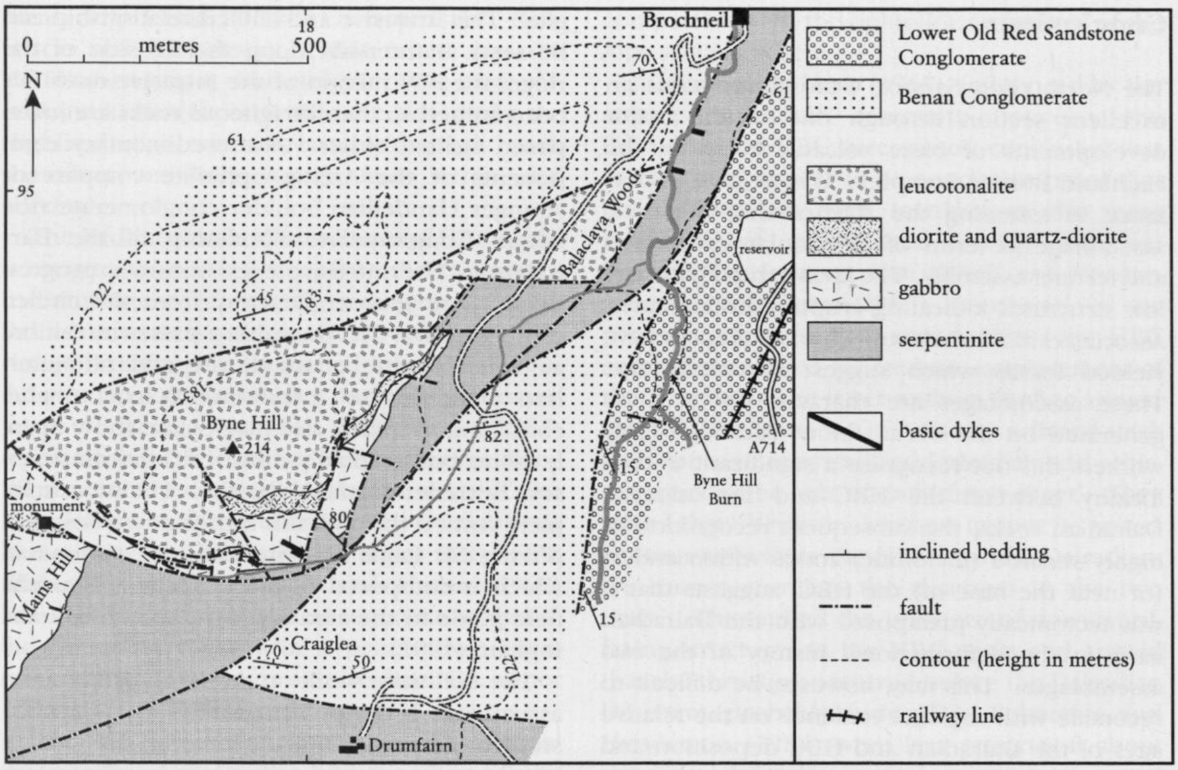

Figure 2.24 Map of the Byne Hill area (after Bloxam, 1968).

veins of calcium-rich secondary minerals such as hydrogrossular garnet and pectolite, believed to have been produced during the late stages of serpentinization. The veins cross the intrusive contact to penetrate the marginal gabbro establishing that intrusion was an early event preceding serpentinization. This deduction is reinforced by the local metasomatism of the marginal gabbro with the development of the same calcium-rich secondary minerals as seen in the veins.

The chilled margin of the intrusive body is doleritic but this coarsens to a gabbro within a few tens of centimetres of the contact. Olivine-gabbro forms the outermost facies but passes inwards into hornblende gabbro by the gradual loss of olivine and progressive mimetic replacement of clinopyroxene by brown hornblende. Parts of the gabbro are notably pegmatitic with individual crystals up to 5 cm across. The transition from gabbro, through diorite and into leucotonalite is relatively abrupt with the dioritic zone rarely wider than 10 m; the transitional lithologies contain distinctively lath-shaped albitized plagioclase, biotite, brown and green hornblende, with accessory quartz and apatite. Bloxam (1968) stressed that gabbroidal minerals and textures survive as relict features in the dioritic transition zone. The inwards increase in quartz is thence marked and, over only 50 cm or so, quartz-diorite passes into the leucotonalite which contains macroscopic quartz and forms the summit area of Byne Hill. It is a pale-pink or very pale orange-coloured rock that weathers white. Quartz and plagioclase (with albitized rims to original oligoclase) together comprise about 90–95% of the leucotonalite with accessory green hornblende, biotite and zircon. The grain size is generally 1–2 mm, ranging up to 3 mm in places. Zircons from the summit area were used by Bluck *et al.* (1980) to produce their U-Pb age of 483 ± 4 Ma. At Byne Hill, the NW margin of the leucotonalite is faulted against the Benan Conglomerate (Figure 2.24) but towards that margin there does seem to be an increase in the proportion of hornblende in the rock. This rather slim evidence may imply that originally the leucotonalite may have formed the core of an elongate body with a symmetrical transition to gabbro towards both the SE and NW margins.

A characteristic feature of the leucotonalite (and to a lesser extent of the quartz-diorite) is the pervasive network of cataclastic zones which traverse the rock. For the most part these are microfractures only clearly evident in thin section but locally they develop into mylonitized zones up to 2 cm across. A wide range of orientations is apparent but there is a preferential NE–SW trend to the macroscopic mylonite zones parallel to the long axis of the intrusive body. This trend is also parallel to the major fault forming the NW margin of the leucotonalite but there does not appear to be an increase in mylonitization towards that structure.

Interpretation

Byne Hill is formed by part of a composite, intrusive igneous body transitional from gabbro at its margins towards a leucotonalite core. The host rock is serpentinized harzburgite and the intrusive contacts are very steep or vertical. From the intrusive contact with serpentinite the transitional passage from gabbro passes through diorite and quartz-diorite into the leucotonalite with no sharp contacts and with gabbroidal minerals and textures persisting through the dioritic zone and possibly into the marginal hornblendic phase of the leucotonalite. The principal mineralogical changes are reported by Bloxam (1968) to be progressive amphibolitization and albitization of the gabbro accompanied by increases in SiO_2 and Na_2O content. Bloxam considered that these features, coupled with the textural evidence, suggest that the dioritic transitional facies contains hybrid rocks produced by reaction between crystalline gabbro and silicic, Na-rich solutions related to the leucotonalite. However, Bloxam also entertained the possibility that the leucotonalite could be entirely metasomatic in origin. Subsequently, Bluck *et al.* (1980) examined zircons from the leucotonalite and considered them to be of magmatic origin, their age therefore dating the time of crystallization.

The presence of leucotonalite or 'trondhjemite' within the Ballantrae Complex is one of the key lithological features supporting its interpretation as an ophiolite assemblage formed from obducted oceanic crust (Church and Gayer, 1973). The U-Pb age of 483 ± 4 Ma obtained from the zircons by Bluck *et al.* (1980) is thus of great importance in terms of the evolution of the oceanic crust which became the Ballantrae Complex. The zircon age is broadly early Arenig and intrusion was thus penecontemporaneous with eruption of the extensive lava sequences. At the margins of the intrusive body gabbro is chilled against harzburgite of the Northern Serpentinite Belt which must therefore have cooled by about 483 Ma. However, Bluck *et al.* (1980) also presented K-Ar age data from amphibolites associated with the metamorphic sole of the ophiolite produced at the base of the Northern Serpentinite Belt during its obduction (see Knocklaugh GCR site). These show that the metamorphic sole was cooling at 478 ± 8 Ma but in terms of error overlap there is no appreciable difference between the ages of lava eruption, ophiolite obduction and leucotonalite intrusion; all occurred around the early to middle part of the Arenig epoch. This relatively short time lapse between generation and obduction of the Ballantrae ophiolite encourages interpretation in terms of a back-arc or marginal basin formed close to the site of its eventual obduction onto the Laurentian continental margin. Supporting evidence comes from the chemistry of the gabbro with data presented by Stone and Smellie (1990) compatible with its generation in a relatively mature, oceanic island-arc system.

A final discussion point in the interpretation of the Byne Hill body has been its use in constraining the Arenig time-scale. The U-Pb zircon age of 483 ± 4 Ma is accepted as a reliable data point and has been extrapolated by Bluck *et al.* (1980) to clasts of leucotonalite contained in mélange interbeds within the volcanosedimentary succession of the Ballantrae Complex. Bluck *et al.* regarded these as of middle Arenig age but subsequent biostratigraphical work by Rushton *et al.* (1986) suggests that a lower Arenig age is more likely. Allowing some time for unroofing and erosion it seems most unlikely that lower Arenig strata could contain igneous material with a cooling age as young as *c.* 483 Ma. Nevertheless, the petrographical similarity between the Byne Hill leucotonalite and some of the mélange clasts is remarkable, particularly so in an example from the south of the Ballantrae Complex (the Craig Hill Breccia Formation of Stone and Smellie, 1988). Their relationship and its overall significance remains uncertain.

Conclusions

The Byne Hill site is of great importance in any assessment of the origin and tectonic role of the

Ballantrae Complex within the Caledonian Orogen. The gabbro–leucotonalite intrusion is one of the key lithological components supporting the interpretation of the complex as of ophiolitic origin. The intrusion has a marginal facies of gabbro, chilled against the serpentinized harzburgite host rock, with an inwards transition to leucotonalite through a narrow dioritic zone. Zircons have provided a reliable U-Pb date of 483 ± 4 Ma for cooling of the leucotonalite magma, thus constraining the age of the complex as a whole. The chemistry of the intrusive body, particularly the gabbro, is compatible with an origin at an oceanic volcanic island arc rather than within oceanic crust (*sensu stricto*).

SLOCKENRAY COAST (NX 135 911 AND 136 914–143 924)

P. Stone

Introduction

The Slockenray Coast GCR site includes a part of the Northern Serpentinite Belt, its faulted margin with the Pinbain volcanosedimentary block, and an olistostrome, lavas and volcaniclastic strata within that block. The site falls into three parts each of which contains an important component of the Ballantrae Complex ophiolite. The northern sector forms the major part and provides an extensive section through an Arenig volcanosedimentary sequence of lava and intercalated volcaniclastic sedimentary rocks. The central part of the site is occupied by a structurally complex zone with a coarse conglomeratic olistostrome faulted against serpentinite. The isolated southern outlier contains a remarkable pegmatitic gabbro intruded into the serpentinite, which is also traversed by veins of coarse-grained pyroxenite. A general account of the geology is given by Stone and Smellie (1988).

The good accessibility and exposure have made it one of the most intensively studied parts of the Ballantrae Complex so that it has been central to several of the historical controversies which have raged over interpretation of the regional geology. Initially there was debate over a metamorphic versus an igneous intrusive origin for the complex and crucial evidence in support of the latter was obtained by Bonney (1878) from the pegmatitic gabbro at the southern end of the section. To mark this breakthrough the name 'Bonney's Dyke' was coined for the locality by Balsillie (1932) and it is generally known by this colloquial name in the geological literature. The gabbroic dyke has a striking black and white appearance, created by large plagioclase and altered clinopyroxene crystals; it displays a range of textures within an overall coarse pegmatitic style. The host rock is serpentinized harzburgite of the Northern Serpentinite Belt which at this locality contains numerous veins of coarse pyroxenite.

A short distance north from Bonney's Dyke, in the vicinity of Pinbain Bridge at the southern end of the main GCR sector, the Northern Serpentinite Belt is faulted against an unusual conglomeratic unit. This comprises a variety of clasts, both in terms of size and composition, contained in a black shaly matrix that is pervasively foliated in places. It was originally interpreted in volcanic terms but, with the development of the ophiolitic model, Church and Gayer (1973) noted its similarity to the extensive mélange deposits associated with the large-scale ophiolite bodies of Newfoundland. In this interpretation, masses of brecciated pillow lava, apparently interbedded with the conglomerate, may be rather viewed as large blocks within an original olistostrome.

The olistostrome is most pervasively foliated adjacent to its southern margin and the fabric becomes less intense towards the northern margin where there is another faulted sliver of serpentinite. This narrow zone provided the focus for much of the argument over a proposed intrusive origin for the serpentinite. The critical relationships were interpreted by Anderson (1936) in favour of intrusion of the serpentinite into the lava sequence while Balsillie (1937) was using information from elsewhere in the Ballantrae Complex to deduce that the serpentinite was much older than, and formed a basement to, the lavas (see particularly the Knockormal GCR site report). Bailey and McCallien (1957) re-examined the Pinbain locality, having been previously disposed towards the intrusive interpretation, and decided that the apparently intrusive veins are of secondary origin and are probably related to serpentinization of an ultramafic precursor. This conclusion is now generally accepted although Bailey and McCallien's broader interpretation, that the serpentinite originated as an ultramafic, submarine lava flow, has now been superseded by the ophiolite model.

Northwards from the olistostrome–serpentinite fault slice the coastal section comprises a relatively continuous volcanosedimentary succession. The beds are vertical or steeply inclined and young consistently northwards. Despite the abundant minor faulting this is the least structurally disturbed part of the Ballantrae Complex and many sedimentological and volcanic features are well displayed. Two aspects are of particular importance: a graptolite fauna collected from the southern end of the section and a remarkable lava-front delta assemblage in the central part at Slockenray itself. The graptolites were recovered and described by Rushton *et al.* (1986) who deduced an early Arenig age. The Slockenray relationships have been described in great detail by Bluck (1982) who emphasized their support for eruption of the lava sequence into relatively shallow water. The lack of spilitization of some of the Slockenray rocks was also stressed by Bluck following earlier debate over their exceptional lack of alteration. This was sufficiently striking that Bailey and McCallien (1957) admitted to briefly entertaining the idea that they were contained in a volcanic neck of Palaeogene age.

Description

The outline geology of the Slockenray–Pinbain section, which encompasses the Slockenray Coast GCR site, is shown in Figure 2.25. The Balcreuchan Group lithostratigraphy shown is based on that proposed informally by Bluck (1982).

At the southern extremity of the section, within the Northern Serpentinite Belt, a striking pegmatitic gabbro body (Bonney's Dyke) protrudes from the ultramafic rocks on the foreshore and just above the high-water mark. It is an elongate mass up to about 5 m wide, with an overall WNW trend made arcuate by the cumulative offsets of several minor sinistral faults (Figure 2.26). The gabbro is a pale-coloured rock formed almost exclusively of plagioclase and clinopyroxene, the latter in a rather fibrous variety of diopside and/or augite termed 'diallage' in the older literature. These minerals are now extensively altered, the plagioclase to a calcium-rich assemblage of pectolite, prehnite and hydrogrossular garnet and the pyroxene to an aggregate which includes actinolite and chlorite; some serpentine minerals are also present, which may indicate original olivine. Grain size is very variable (Figure 2.27) with fine-grained patches alternating irregularly with zones in which the plagioclase laths and pyroxene crystals range up to 2 cm. Xenoliths of serpentinite up to about 50 cm across are abundant locally. The margins of the gabbro body generally dip moderately to the south but three types of contact relationship are displayed: sharp contacts against serpentinite (including the xenoliths); more diffuse margins abutting coarse pyroxenite veins; and sheared margins that have recrystallized to a very fine-grained secondary mineral assemblage through calcium metasomatism. It is noteworthy that none of the gabbro margins are chilled unequivocally against the ultramafic rock.

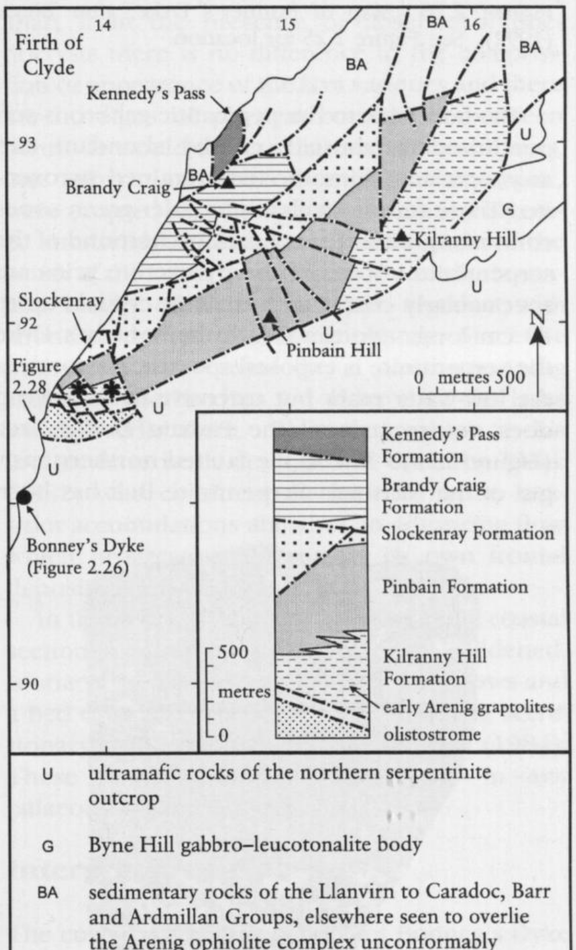

Figure 2.25 Map and stratigraphy of the Pinbain Block, the northernmost part of the Ballantrae Complex, after Bluck (1982) and Stone and Smellie (1988). * The area marked thus on the map was included within the Pinbain Formation by Bluck (1982), but is more closely related lithologically to the Kilranny Hill Formation.

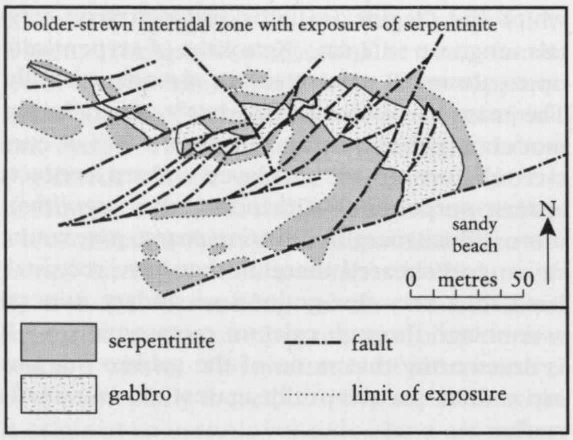

Figure 2.26 Map of Bonney's Dyke, after Bluck (1992). See Figure 2.25 for location.

The host rock to the pegmatitic gabbro is serpentinized harzburgite cut by abundant and anastomosing veins of coarse-grained pyroxenite. These stand out clearly as paler-green zones contrasting with the very dark background of the serpentinite. Some of the pyroxenite veins are spectacularly coarse with diallage crystals up to 10 cm long. Northwards from Bonney's Dyke the serpentinite is exposed sporadically towards the low-water mark but extensive rock outcrop does not recur until the Pinbain Bridge area (Figure 2.25). There, the faulted northern margin of the Northern Serpentinite Belt has been

Figure 2.27 Textural variation in the Bonney's Dyke pegmatitic gabbro between plagioclase (pale) and pyroxene (dark). The long axis of the sample is 170 mm. (Photo: BGS no. MNS4007.)

intruded by a dolerite dyke of Palaeogene age which has baked and hardened the adjacent ultramafic rock. Accordingly, this marginal zone of serpentinite has proved unusually resistant to erosion and now stands proud of the surrounding beach and rock platform.

Northwards, beyond the Palaeogene dyke, is an extensive outcrop of the Pinbain olistostrome. This has been described in great detail by Bluck (1978a, 1992) and is illustrated in Figure 2.28. The southern part of the olistostrome, adjacent to the serpentinite and dyke, consists of a mélange containing a great variety of clasts set in a pervasively foliated, black shaly matrix. The clasts range considerably in shape and size with some boulders several metres across. An equally large lithological range is present; basalt tends to form the larger clasts with some appearing to be intact pillows of lava, large pale clasts of carbonate may well be highly altered ultramafic rock but small clasts of dark-green serpentinite are also present. The other smaller clasts have a remarkably large lithological range with records of shale, tuff, chert, basalt, dolerite, gabbro, pyroxenite and, very rarely, glaucophane- and garnet-glaucophane schist (Bailey and McCallien, 1957). A large mass of brecciated pillow lava at the northern margin of the mélange may be an interbedded unit or may be an exceptionally large clast caught up in the olistostrome. Farther north still, the lithology becomes more conglomeratic with a more restricted clast size range and a generally unfoliated shale matrix. Granitic clasts, quartz, and fragments of quartz-cemented breccia appear at this level. The northern margin of the Pinbain olistostrome is formed by a complex fault zone introducing slivers of serpentinite and intruded by Palaeogene dolerite dykes. The olistostrome is thus structurally isolated from the volcano-sedimentary sequence forming the main part of the Slockenray section.

The Slockenray coastal section provides a cross-strike traverse through the Pinbain block, the northernmost and least structurally confused of the three principal volcanic units which together form the Balcreuchan Group within the Ballantrae Complex. Bluck (1982) divided the approximately 1500 m of lava and volcaniclastic sedimentary strata into five formations and this lithostratigraphical scheme, slightly modified by Stone and Smellie (1988, see especially table 4), is applied in Figure 2.25. The age of the sequence is fixed by the early Arenig graptolite

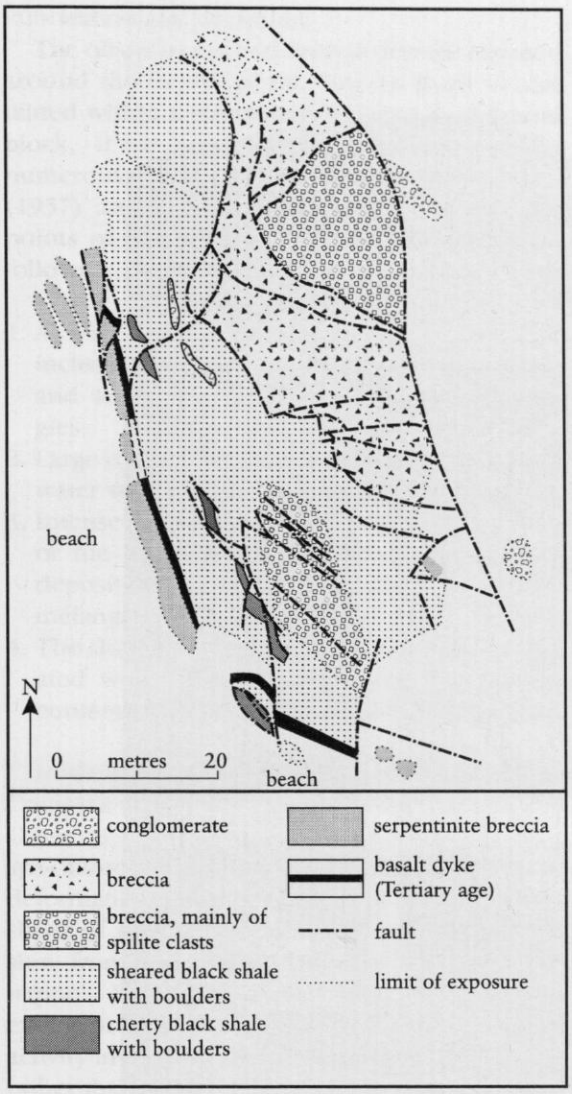

Figure 2.28 Map of the SW margin of the Pinbain olistostrome, after Bluck (1978a, 1992). See Figure 2.25 for location.

fauna described from the southern extremity of its outcrop by Rushton *et al.* (1986). The fossils were recovered from thin shale laminae within volcaniclastic sandstone where the beds strike NE–SW and dip steeply to the NW. This is a fairly characteristic bedding attitude throughout the section and since younging is consistently to the NW the early Arenig age applies to the base of the sequence. A variety of volcaniclastic sedimentary rocks are interbedded with lavas and lava breccias. The lavas are variably porphyritic and aphyric, although even the aphyric lavas commonly contain scattered microphenocrysts. Plagioclase is the most common of the phenocryst minerals and there are also rare augite and olivine phenocrysts, all set in a matrix dominated by plagioclase and augite. Spilitization and low-grade metamorphism have created widespread secondary minerals including albite, chlorite, titanite, epidote and prehnite.

An area of particular interest is centred on Slockenray itself where there has apparently been mixing of two lava types, one aphyric and the other strikingly porphyritic (Figure 2.29). In the latter, large plagioclase laths are aligned in swirling patterns indicative of flow orientation and the zones of phenocrysts have relatively sharp margins against aphyric lava. However, apart from the presence or absence of phenocrysts there is no difference in the composition or appearance of the lava varieties and there are no signs of physical boundaries between them. It would appear that two different lavas, one aphyric and the other plagioclase-phyric, were erupted simultaneously and mixed together immiscibly during flow. Immediately to the south of the lavas, hyaloclastite breccia beds contain blocks of both lava types together with abundant detrital plagioclase crystals. Since the beds are sub-vertical in a north-younging sequence the breccias are older than the lavas which overlie them. The relationships at this locality have been described in detail by Bluck (1982) who considers the breccias to be lava-front accumulations ahead of an advancing flow which progressively over-rode its own frontal deposits.

In the northern part of the Slockenray coastal section features of interest include reddened, scoriaceous top surfaces of some lava flows and a bed of tuffaceous sandstone containing accretionary lapilli described by Smellie (1984). These features are of importance in any palaeoenvironmental analysis.

Interpretation

The coarse pegmatitic gabbro of Bonney's Dyke was quoted by Church and Gayer (1973) as representing one of the characteristic components of an ophiolite assemblage. Such lithologies may be found in the basal part of the gabbro unit or as a late intrusion into the upper part of the ophiolite 'stratigraphy'. However, while the lithology might be appropriate, the setting of the Bonney's Dyke example is incompatible with either of these styles of occurrence. If it origi-

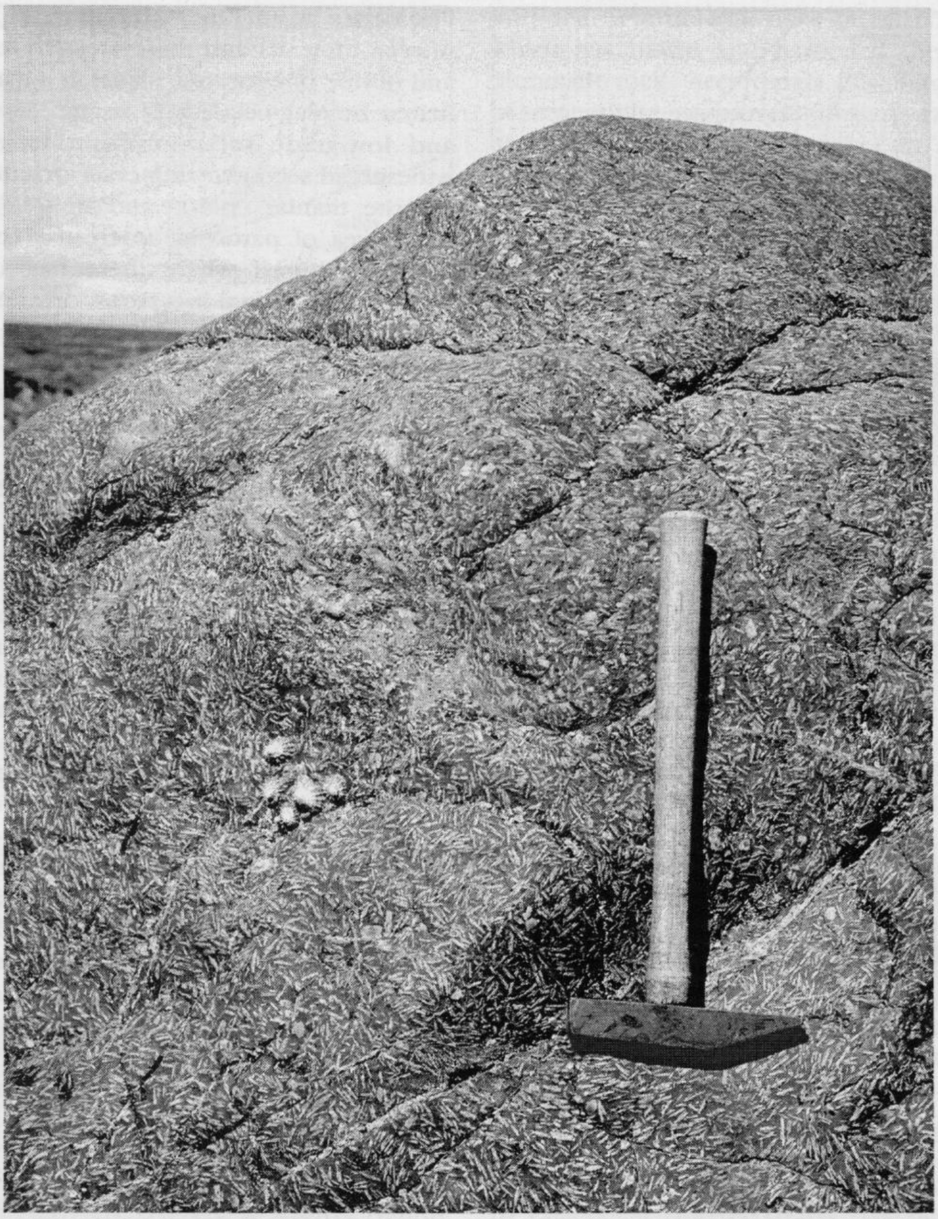

Figure 2.29 Massive, porphyritic basalt lava at Slockenray (NX 1403 9197). The pale feldspar phenocrysts are tabular and range up to 2 cm in length. (Photo: BGS no. D4239.)

nated in the basal gabbro unit its present isolated situation would require interpretation as a clast or xenolith within the serpentinite. This is not supported by the marginal relationships or by the presence of serpentinite xenoliths in the pegmatite; these features all favour an intrusive origin. The intrusion is not into the upper levels of the ophiolite 'stratigraphy' as the model would predict but instead the pegmatite has been emplaced into mantle harzburgite, the lowest part of the ideal sequence. Thus, assuming that the pegmatite rose on intrusion, the implication of its current position is that the Ballantrae Complex ophiolite was structurally imbricated while the final stages of its igneous generation were still in progress. The calcium-metasomatism of the dyke margins shows that intrusion occurred before serpentinization of the ultramafic rock, further compressing the time-scale of ophiolite development, structural imbrication and alteration.

The olistostrome and breccia deposit exposed around the mouth of the Pinbain Burn is contained within a structurally isolated fault-bound block. It has been described and discussed by numerous authors, notably Bailey and McCallien (1957) and Bluck (1978a, 1992). The salient points of interpretation can be summarized as follows.

1. A wide variety of clast types are present, including the full range of ophiolitic rocks and some high-grade metamorphic lithologies.
2. Large clasts were slumped into relatively deep water where black shale was accumulating.
3. Intense shearing was focused into some parts of the unit during or immediately after its deposition to create a pervasively foliated mélange.
4. The slumped lithologies are intimately associated with tuff and pillow lava indicative of contemporaneous volcanicity.

In their original comparison of the Ballantrae Complex with the ophiolite model, Church and Gayer (1973) described the olistostrome as a syn-obduction deposit 'developed as a result of deformation beneath a forward moving thrust sheet of material previously produced by erosion from the leading edge of the self same nappe'. Bluck (1992) preferred an origin in an extensional environment, citing the volcanic activity in support of an association with normal rather than compressional faults. These brought the deeper levels of the ophiolite to the erosion level in fault scarps, the faults themselves probably rooting in the fissile shaly lithologies and there producing the localized intense ductile deformation. Such structures would almost certainly have been re-activated as thrusts during subsequent obduction-related basin inversion.

A mid- to late-Arenig graptolite was recorded from the olistostrome by Peach and Horne (1899, p. 442). Amphibolite *in situ* elsewhere in the Ballantrae Complex and lithologically identical to clasts in the olistostrome has been dated radiometrically at 478 ± 8 Ma. These relationships have been used by Bluck *et al.* (1980) to suggest a constraint on the Arenig time-scale. However, the graptolite was re-examined by Rushton *et al.* (1986) who concluded 'that it is quite undeterminable and even its graptolitic nature is questionable'. Stratigraphical deductions based on the Pinbain olistostrome should therefore be treated with caution.

The Pinbain volcanosedimentary sequence is approximately 1500 m thick, which suggests considerable synvolcanic subsidence during its accumulation. This point is emphasized by the abundant evidence for shallow-water sedimentation and possible subaerial eruption. Bluck (1982) described the Slockenray breccias in terms of an upward coarsening sequence built up as a hyalotuff delta, ahead of lava flows, in the shallow sub-tidal or intertidal zone. He summarized the situation as 'the lavas advanced over the sediments so a sequence was generated where these sediments have a source in lava flows which were eventually to overlie them'. Syn-eruption subsidence would seem implicit in this interpretation. Higher in the Pinbain Block succession the reddened tops of lava flows and the presence of accretionary lapilli (Smellie, 1984) are both indicative of subaerial eruption with the lapilli falling into shallow water.

Several cycles of lava advance have been described by Bluck (1992). Slow eruption rates produced a pillowed flow whereas rapid eruption produced a more massive lava; the lavas were continuously eroded into breccias and hyaloclastites and the sea may have transgressed the flow top between eruptions. The mixing of plagioclase-phyric and aphyric lavas, well displayed at Slockenray itself, requires the contemporaneous eruption of different magma types in close proximity. Flow segregation of the feldspar phenocrysts may be a possible alternative explanation but has not found much support from previous investigators.

The geochemistry of the Pinbain–Slockenray lavas has been generally interpreted in terms of within-plate, oceanic island eruption (Wilkinson and Cann, 1974; Thirlwall and Bluck, 1984; Stone and Smellie, 1990). This is most readily reconciled with the evidence for subsidence and shallow water in a marginal or back-arc basin. In this respect the sequence compares closely with that developed in the Bennane Head sector of the Balcreuchan Port to Port Vad GCR site but at Pinbain a geochemical curiosity occurs at the base of the succession. Here, volcaniclastic sandstones form the oldest stratigraphical level and are interbedded with graptolitic shales of early Arenig age (Rushton *et al.*, 1986). The sandstones are formed exclusively of volcanic material and on analysis prove to have the chemical characteristics of island-arc lavas (Stone and

Smellie, 1990; Smellie and Stone, 1992). If this association is accepted the back-arc model is strengthened since it provides an environment in which arc-derived sands could be overlain by within-plate lavas erupted during a phase of significant subsidence. Evidence of this sort, provided by the Slockenray Coast GCR site, is of fundamental importance in the interpretation of the Ballantrae Complex as a whole. A polygenetic source for the ophiolite seems increasingly likely with island-arc and within-plate components now structurally juxtaposed.

Conclusions

The Slockenray Coast GCR site contains three different geological components, each of which is fundamental to the ophiolitic interpretation of the Ballantrae Complex. Their characteristics are exceptionally well displayed in a section that has received much historical study.

At the southern end of the GCR site the coarse gabbro body forming Bonney's Dyke shows a spectacular development of very coarse-grained, pegmatitic texture. The lithology is an integral part of the ophiolite assemblage but its position here, intruded into serpentinite cut by pyroxenite veins, is somewhat anomalous. It suggests early structural disordering of the Ballantrae Complex while ophiolitic magmatism was still in progress.

The foliated conglomeratic (mélange) deposit exposed around the mouth of Pinbain Burn has been the subject of much debate. It contains clasts with a wide size and composition range, the latter including both ophiolitic lithologies and exotic, high-grade metamorphic rocks. An origin by large-scale slumping is now generally accepted.

The accessible and well-exposed Slockenray coastal section forms most of the GCR site; it contains the most complete and least structurally disturbed volcanosedimentary succession within the Ballantrae Complex. The lavas are believed to have been erupted in an oceanic island, within-plate environment and graptolite fossils from intercalated shale at the base of the sequence fix the age at early Arenig. Some of the lavas apparently resulted from the mixing of different varieties during flow and may have been erupted into relatively shallow water. At one locality the lava–sediment relationships have been interpreted in terms of a lava front delta deposited in an intertidal environment.

KNOCKLAUGH (NX 168 920)

P. Stone

Introduction

The Knocklaugh GCR site provides a section through a sequence of dynamothermal metamorphic rocks developed from shale and spilitic lava marginal to the Northern Serpentinite Belt of the Ballantrae Complex ophiolite. Lithologies range from chlorite and epidote schists, prograding into amphibolites towards the ultramafic belt, to the highest-grade metamorphic rock seen, which is a garnet metapyroxenite. However, since this lies just within the serpentinite it may be unrelated to the schist–amphibolite assemblage. The aureole of metamorphic rocks is generally about 40 to 50 m wide and can be traced discontinuously along the serpentinite margin for over 5 km.

The unusual metamorphic assemblage was first noted by Peach and Horne (1899, pp. 456–9) and subsequently described in some detail by Anderson (1936) who proposed an origin by tectonism during the intrusion of hot ultrabasic magma. Later, the development of plate tectonic concepts and the recognition of ophiolites as obducted oceanic lithosphere allowed the re-interpretation of the Ballantrae Complex in those terms. Church and Gayer (1973) considered the Knocklaugh rocks to be part of a metamorphic aureole formed at the base of an obducting slab of still-hot mantle material. They drew attention to the similarities with dynamothermal aureoles present beneath the large, well-preserved ophiolite complexes of Newfoundland and noted the inversion of the metamorphic succession, with the highest grade rocks at the top (closest to the over-riding hot slab) of an originally sub-horizontal metamorphic layer. More recent and detailed studies of the aureole rocks (Spray and Williams, 1980; Treloar *et al.*, 1980) have confirmed the broad interpretation while stressing the range of temperature and pressure conditions required for formation of the different components.

Description

Detailed descriptions of the metamorphic aureole rocks at Knocklaugh, in both petrographic and structural terms, are provided by Spray and

Williams (1980) and Treloar *et al.* (1980). A Sm-Nd isotopic age of 505 ± 11 Ma has been reported by Hamilton *et al.* (1984) for the high-grade metapyroxenite whereas Bluck *et al.* (1980) gave a K-Ar age of 478 ± 8 Ma for the lower-grade amphibolites. The geological setting is illustrated in Figure 2.30.

The Knocklaugh GCR site provides a section wherein the Arenig lavas and interbedded cherts and shales (oceanic upper crust) of the Balcreuchan Group become increasingly sheared and schistose towards the SE margin of the Northern Serpentinite Belt. Epidote schists are seen as slaty rocks with streaked epidote augen and granular boudins, up to a few millimetres across, dispersed within finer-grained, dark-green laminae of actinolitic hornblende, chlorite and albite. Lighter-coloured, yellow-green layers of fine-grained epidote and titanite are also present in places together with sporadic metasedimentary quartz-albite-epidote-muscovite-chlorite schists and small pods of recrystallized carbonate. Closer to the serpentinite the epidote schists pass abruptly into coarser-grained, dark-grey amphibolites composed largely of plagioclase and green or brown hornblende. The more highly foliated of the amphibolites may contain garnet, and pyroxene occurs as an accessory in amphibolites adjacent to the serpentinite margin. Hornblende-bearing garnet metapyroxenite also occurs within the contact zone but in some cases is separated from the amphibolites by thin slivers of serpentinite. This may be due to structural imbrication of the schistose aureole and adjacent serpentinite during the final stages of thrust movement. Alternatively, the garnet pyroxenite may have originated as segregations near the base of the ultrabasic precursor to the serpentinite and so would not be part of the dynamothermal aureole *sensu stricto*. The width of the aureole assemblage is generally 40 to 50 m. Locally this may reduce to 20 m and in places the entire aureole has been removed tectonically; conversely the epidote schist component may broaden locally to nearly 200 m.

Structurally, the epidote schists and amphibolites of the aureole are dominated by a slaty schistosity or fine gneissose foliation with a locally developed mineral aggregation lineation. These fabric elements dip or plunge steeply to the NNW. Within the schistose fabric tight and asymmetrical, reclined interfolial folds close both SW and NE. They are refolded by small,

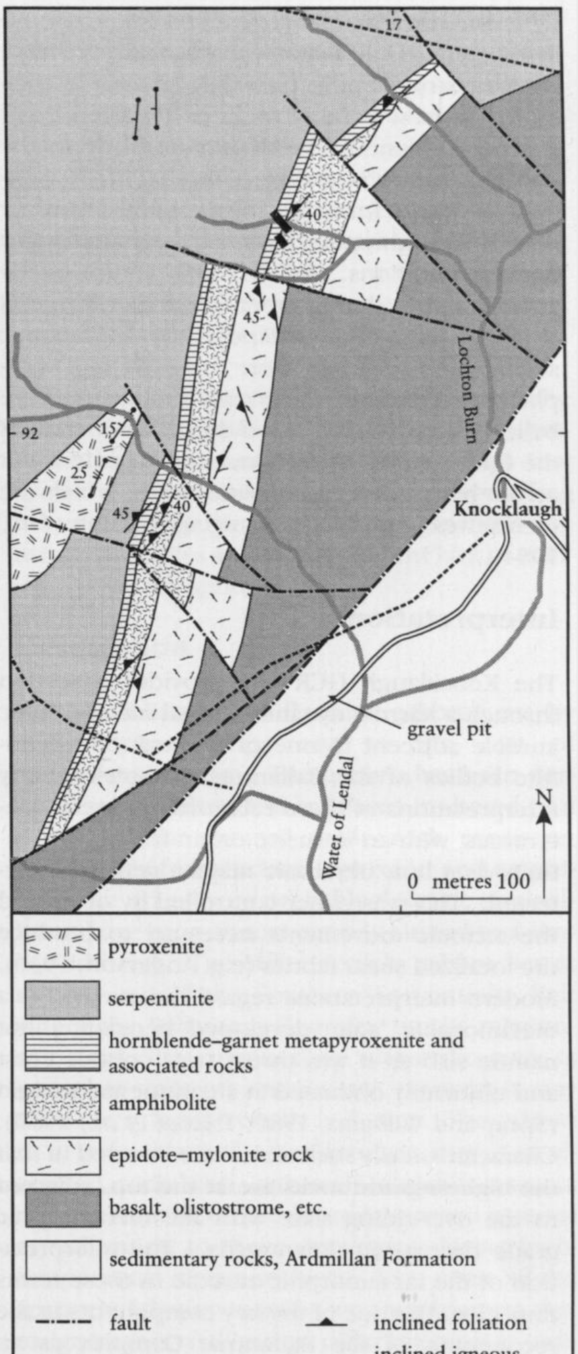

Figure 2.30 Map of the metamorphic aureole developed adjacent to serpentinite at Knocklaugh, after Treloar *et al.* (1980).

tight and angular, straight-limbed folds with an associated axial-planar crenulation cleavage. Such a history of polyphase deformation restricted to a narrow band of rocks is characteristic of a mylonite zone. The dominant fabric in the metapyroxenite is rather different, more compatible with deformation by prolonged laminar flow under granulite facies or perhaps upper mantle conditions. The relatively abrupt transitions between the different lithologies suggests that they have been juxtaposed by the tectonic slicing of an originally more extensive metamorphic aureole. All of these structural features are believed to have developed during obduction of the Ballantrae Complex since the aureole is cut across by unsheared doleritic dykes, which are themselves of probable Arenig age (Holub *et al.*, 1984).

Interpretation

The Knocklaugh GCR site provides a section through a narrow dynamothermal metamorphic aureole adjacent to one of the major serpentinite bodies of the Ballantrae Complex. Early interpretations of these relationships were contentious, with an intrusive origin for the serpentinite as a hot, ultrabasic magma generally preferred. This was either controlled by or caused the tectonic movements necessary to produce the localized shear fabrics (e.g. Anderson, 1936). Modern interpretations regard the aureole as a metamorphic 'sole' developed beneath a hot mantle slab as it was thrust over oceanic crust and ultimately obducted at a continental margin (Spray and Williams, 1980; Treloar *et al.*, 1980). Characteristically such 'soles' are inverted in that the highest-grade rocks are at the top, adjacent to the over-riding slab, with the metamorphic grade decreasing downwards. The interpretation of the metamorphic aureole in these terms thus identifies one of the key components in the recognition of the Ballantrae Complex as an obducted ophiolite.

The temperature–pressure conditions necessary for the generation of the various parts of the aureole are clearly varied. The highest-grade rocks have attracted the most interest with texture, mineralogy and mineral chemistry used to estimate the conditions of metamorphism. Treloar *et al.* (1980) considered the garnet metapyroxenite to have recrystallized under upper granulite facies or mantle conditions with $T = 900 \pm 70°C$ and $P = 10–15$ kbar. This lithology may not be part of the aureole *sensu stricto* but it rather formed as original mantle segregations where the same P–T conditions would apply. However, broadly comparable conditions were determined by Spray and Williams (1980) who considered that formation of the foliated garnet amphibolite required a minimum of $T = 850°C$ and $P = 7$ kbar. At these temperatures the adjacent ultramafic rock would have been unserpentinized suggesting that the onset of metamorphism was an early event in the assembly of the Ballantrae Complex, a deduction compatible with the available radiometric dates.

Various lines of evidence consistently indicate that the upper part of the metamorphic 'sole', i.e. that part closest to the serpentinite, formed under upper mantle conditions. Conversely, the lower-grade components of the aureole formed under very much less extreme conditions, perhaps as low as the greenschist facies, providing the characteristic inverted metamorphic profile. The full range of metamorphic lithologies are developed across only about 40 to 50 m, which represents an impossibly steep temperature gradient; further, the upper parts of the 'sole' were generated at much higher pressures than the lower parts. The limited radiometric age data suggests that the highest-grade rocks may have formed significantly earlier than the lower-grade components. Clearly the aureole assemblage as seen was not produced as a single entity and must have accreted piecemeal under decreasing pressure and temperature conditions. This would be in keeping with its interpretation as an obduction 'sole' wherein the metamorphic components would be both composite and tectonically transported. Treloar *et al.* (1980) envisaged 'a thin, transported and telescoped aureole made up of tectonic pieces of successively-formed metamorphic rocks that welded onto the base of an upward-moving and progressively-cooling peridotite slab'. Overall, it seems most likely that metamorphism was initiated in a deep thrust zone within oceanic lithosphere and continued as the thrust sheet moved upwards to shallower depths. The present array of aureole lithologies is thus the result of progressive incorporation at the base of the thrust sheet during its tectonic rise and emplacement.

Conclusions

Within the Ballantrae Complex ophiolite, at the

SE margin of the Northern Serpentinite Belt, a range of high- to low-grade metamorphic lithologies was produced during obduction. Epidote schist, amphibolite, garnet amphibolite and possibly garnet metapyroxenite form an intermittent aureole to the serpentinite body ranging in width from 20 m to almost 200 m. The garnet metapyroxenite occurs just within the serpentinite and is succeeded, in a south-easterly direction, first by foliated and garnet-bearing amphibolite and then by slaty, banded epidote schist. The regular arrangement of lithologies within the metamorphic aureole arose by the progressive addition of metamorphic slivers onto the base of a hot mantle peridotite body during its thrusting from oceanic lithosphere onto continental crust. This crucial evidence for the obduction of the Ballantrae Complex is well preserved in the Knocklaugh section.

MILLENDERDALE (NX 177 905 AND 177 906)

P. Stone

Introduction

Mafic dykes rich in hornblende and pyroxene and with a granular texture are seen intruded into foliated gabbro in a series of rocky knolls about 500 m east of Millenderdale Farm. The foliation in the gabbro is generally steeply inclined with a variable, approximately NE–SW trend. The dykes range up to about 3 m across and comprise about 25% of the outcrop; for the most part they cross-cut the gabbro foliation but locally they swing into a parallel orientation. The relationships seen are unique within the Ballantrae Complex and are controversial in terms of its interpretation as an ophiolite assemblage.

The unusual characteristics of the gabbro–dyke assemblage were first described by Teall (in Peach and Horne, 1899) and discussed in some detail by Balsillie (1932) and Bloxam (1955) before its interpretation by Church and Gayer (1973) as representing a part of the sheeted-dyke component seen in an ideal ophiolite sequence. However, that interpretation has not been widely accepted and the problem has been summed up by Bluck (1978a) as follows.

> 'These outcrops have been interpreted as part of a sheeted dyke complex ... but the proportion of dykes to country rocks is considerably lower than is found in other ophiolites, e.g. 90–100% of the sheeted dyke complex is dyke rock at Troodos .... Dyke intrusion in a tectonically active regime where gabbroic rocks were forming and cooling might well fit the model of a spreading ridge, but on the exposed evidence the outcrop is merely gabbro cut by a few dykes ... a sheeted dyke complex is either poorly represented or not present at all in the Ballantrae ophiolite.'

Despite the sheeted-dyke controversy the Millenderdale gabbro–dyke assemblage preserves unusual textures and relationships that probably originated at an oceanic spreading ridge. Of particular interest is the granofels texture in the dykes and their apparent lack of chilling against the gabbro.

Description

Variably foliated and flasered gabbro, cut by mafic dykes, crops out to the east of Millenderdale Farm. The gabbro has been pervasively hornfelsed and in places is altered to a dark, granular-textured mafic rock consisting essentially of plagioclase, pyroxene and amphibole; the dykes were probably doleritic originally but have been almost entirely altered to a similar lithology. The Millenderdale gabbro lies at the NE end of the Southern Serpentinite Belt within the Ballantrae Complex but its relationship with the surrounding ultramafic rock is uncertain since no contacts are exposed; it may form a very large tectonic inclusion. The gabbro is a pale-coloured, medium- to very coarse-grained rock with extremely variable textures ranging from those produced by normal igneous crystallization to those developed during high-temperature metamorphism. That some of the apparent foliation must be a primary feature due to igneous crystallization differentiation is indicated by cryptic variation in pyroxene composition and the presence in some mafic bands of relict olivine.

The dykes are medium grained and dark-grey, locally weathering to a rusty-brown colour. Most have a granular, saccharoidal texture and are not chilled against the host gabbro, indicating that it was still hot at the time of dyke intrusion. The dykes have a very wide range of orientations and no dominant trend is evident; indeed, cross-cutting relationships between dykes are fairly common, perhaps indicating a protracted period of intrusion. Some dykes follow a meandering and anastomosing path and may enclose xenoliths of foliated gabbro, whereas others cut straight across the gabbro foliation (and earlier dykes) with no deviation.

Most dykes that lie parallel to the NE–SW gabbro foliation are foliated themselves, with augen structure developed in places, so that their margins appear gradational with the gabbro. The foliated dykes may then be cut by either of two further dyke varieties. Completely non-foliated dykes generally cross-cut the foliation at a high angle but a considerable range of intersections are seen and some non-foliated dykes lie parallel to the gabbro foliation; in some examples the gabbro foliation bends to become conformable with the dyke margin. A few of the dykes have a very faint internal foliation that bears no simple relationship to the main gabbro fabric; this variety may cut either foliated or non-foliated dykes but may in turn be cut by other non-foliated dykes.

The metamafic texture and composition of the dykes and some zones within the gabbro is a striking feature shared with numerous other doleritic enclaves within the Southern Serpentinite Belt. However, only at Millenderdale can the range of intrusive relationships be seen. The metamafic dyke rocks were described as beerbachites by Bloxam (1955) and this name has remained in common usage although beerbachite (*sensu stricto*) is a leucocratic hornfels dominated by plagioclase and containing little or no amphibole. The Millenderdale 'beerbachites' have a fine-grained (0.1–0.3 mm), equidimensional and polygonal granofels texture with only rare porphyroclasts of plagioclase or pyroxene, which results from the almost complete recrystallization of original igneous minerals (Figure 2.31a). They are high-grade, thermally metamorphosed rocks. Plagioclase, generally andesine, makes up between about 30% and 60% of the rock accompanied by variable quantities of pyroxene and brown amphibole, the latter forming by alteration of the former. The pyroxene is variously salite, diopside or hypersthene, the amphibole is Ti-rich ferroan pargasite, and olivine is a very rare accessory. A detailed mineralogical and chemical description is given by Jelinek *et al.* (1980). With the development of a foliation in the 'beerbachite' dykes the distribution of the different minerals becomes increasingly heterogeneous until the development of discrete mafic and leucocratic domains gives the rock a streaky appearance (Figure 2.31b).

Interpretation

Geological relationships preserved at Millenderdale are unique within the Ballantrae Complex and influence its interpretation as an ophiolite assemblage. It therefore forms an important complement to the GCR sites elsewhere in the Ballantrae district which illustrate other ophiolitic aspects. Unusual igneous and metamorphic features are also seen. The gabbro–dyke relationships require a protracted period of intrusion commencing while the gabbro was still in a relatively hot and plastic state. Development of foliations between phases of intrusion implies that the site of intrusion was tectonically active. The textural evidence indicates that much of the hornblende in the 'beerbachites' has replaced granoblastic pyroxene developed in original doleritic and gabbroic lithologies. This requires a post-magmatic metamorphism at about 900 to 1000°C (Jelinek *et al.*, 1980). Cooling from this event is probably recorded by the K-Ar age of 487 ± 8 Ma obtained from the flaser gabbro by Bluck *et al.* (1980) which is within error of most other indicators of the age of formation of the Ballantrae Complex. Geochemical data presented by Jelinek *et al.* (1980) for the gabbro and beerbachite dykes was interpreted in terms of a genetically related, ocean-floor tholeiite suite with the more evolved gabbros originating as the upper differentiates of a layered assemblage. In this context, although not representing a true sheeted-dyke complex, the Millenderdale gabbro–dyke association uniquely represents an important component of ophiolite 'stratigraphy'. However, data from the gabbro alone was thought by Stone and Smellie (1990) to be more indicative of within-plate, ocean-island magmatism and so the nature of the original intrusive setting remains controversial.

Quite apart from uncertainties over the geotectonic setting of intrusion there has also been debate over the circumstances and mechanics of incorporation of the metamafic rocks into the ultramafic belt. Various possibilities have been considered ranging from intrusion into a mass of hot peridotite with subsequent disruption and metamorphism as xenoliths, to metamorphism resulting from repeated intrusion in the same a

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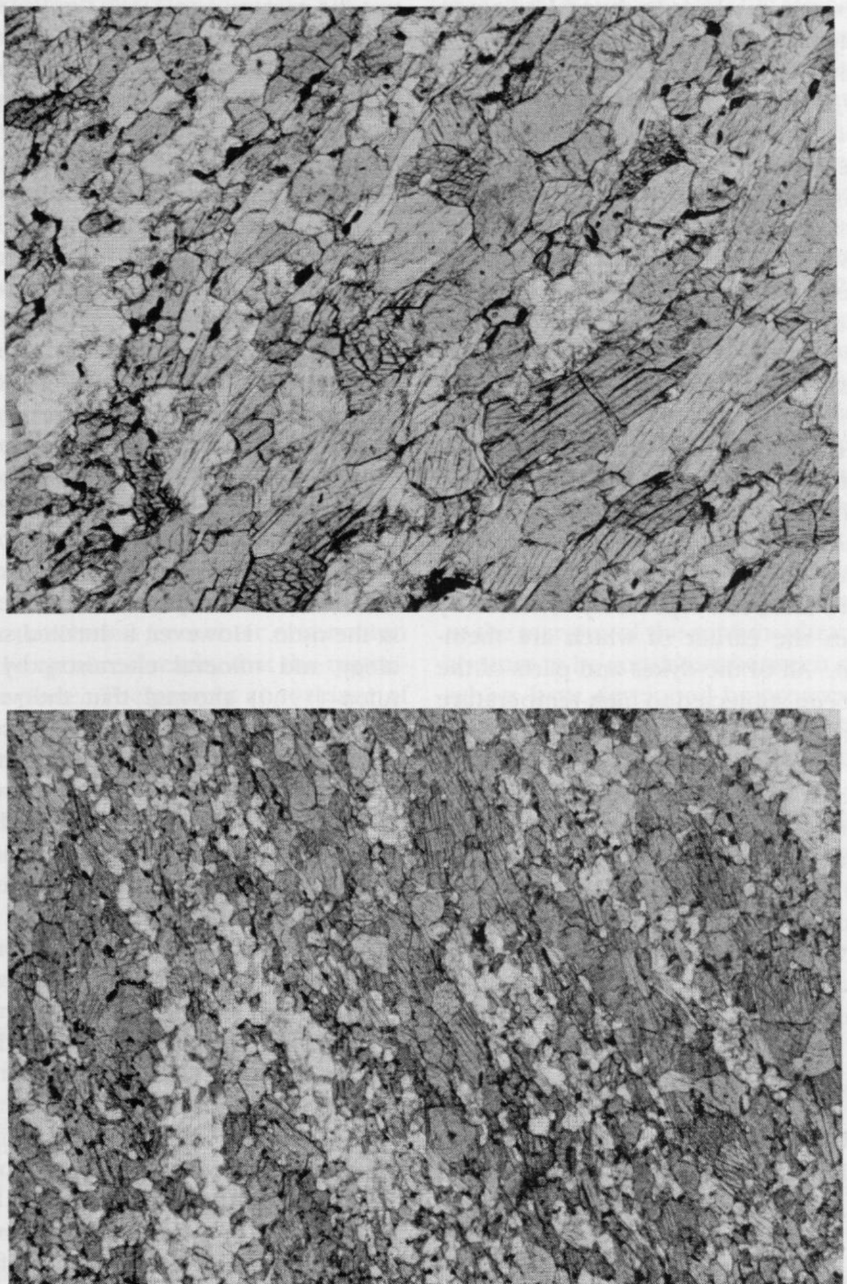


Figure 2.31 Photomicrographs illustrating textures seen in 'beerbachite' dykes near Millenderdale:
(a) Polygonal, granoblastic texture developed between feldspar and brown amphibole. Plane polarized light, ×40. (Photo: BGS no. PMS469);
(b) Development of mafic and leucocratic domains. Plane polarized light, ×20. (Photo: BGS no. PMS470.)

area with the gabbro–dyke blocks incorporated into the serpentinite during its cold, diapiric rise. Currently the most satisfactory explanation is probably the faulted spreading ridge model proposed by Jelinek *et al.* (1980). They envisaged a part of the ridge displaced by transform-fault movement such that spreading and fault movement in opposing directions combined to keep a ridge segment adjacent to the high heat-flow zone. Such an environment would allow both polyphase intrusion and high-temperature metamorphism in a tectonically active region.

Conclusions

The Millenderdale GCR site contains one of the more enigmatic components of the Ballantrae Complex, the interpretation of which has an important bearing on models for development of the complex as an ophiolite. A banded and foliated gabbro has been repeatedly intruded by doleritic dykes the earlier of which are themselves foliated. All of the dykes and parts of the gabbro have been subjected to high-temperature metamorphic recrystallization and are now granular-textured metamafic rocks ('beerbachites' *senso lato*) rich in pyroxene and amphibole. The explanation of these phenomena may lie in the development of the gabbro–dyke assemblage at or near an oceanic spreading centre; faulting is invoked to maintain the assemblage in the high heat-flow environment and allow thermal metamorphism. If such an interpretation is correct the Millenderdale rocks uniquely illustrate an aspect of ophiolite lithology not seen elsewhere in the Ballantrae Complex.

KNOCKORMAL (NX 134 885, 137 890, 138 890 AND 143 891)

P. Stone

Introduction

That part of the Ballantrae Complex ophiolite exposed between Knockormal and Carleton Mains farms spans an anastomosing fault zone trending NE–SW through the central of the three volcanic belts (Figures 2.5 and 2.32). The area has long been famous for the reported occurrence within the fault zone of 'eclogite' (garnet clinopyroxenite with, in this case, accessory hornblende and spinel) and glaucophane-crossite schist. At its SW extremity this zone cuts out the metamorphic sole developed at the base of the Northern Serpentinite Belt and farther NE it introduces slivers of serpentinite into the upper Ordovician sedimentary sequence, which elsewhere overlies the lower Ordovician (Arenig) ophiolitic lithologies unconformably. A polyphase history of post-obduction movement is evident with cataclastic and brittle fabrics imposed locally.

The first extensive study of the Knockormal rocks was by Balsillie (1937) who identified the essential paradox; these are very high-grade metamorphic rocks requiring extremely high pressure for their formation, yet they are in close proximity to the extensive Ballantrae lava successions, which have experienced only very low-grade regional metamorphism. Balsillie's solution was to regard the high-grade rocks as relics of an ancient crystalline basement onto which the lavas were erupted, a controversial proposal at the time. However, a detailed study of mineralogy and mineral chemistry by Bloxam and Allen (1960) showed that the various glaucophane and other blue-amphibole rocks had been derived from spilitic lavas identical to those widely seen elsewhere in the Ballantrae Complex. Transitional lithologies were also discovered and an origin through in-situ faulting and shearing was favoured. The eclogite was also studied by Bloxam and Allen and described as a likely segregation within a host peridotite (now serpentinized), possibly forming as a primary component of a layered ultramafic body.

With the identification of the Ballantrae Complex as an obducted ophiolite, speculation was renewed and comparisons made with better exposed analogues elsewhere in the world. Church and Gayer (1973) introduced the possibility that all of the exotic lithologies at Knockormal could be contained within a large-scale, tectonized slump deposit formed during the obduction process. Bloxam (1980) suggested that the glaucophane schist could form tectonic inclusions contained within the serpentinite and derived from an older source, an echo of Balsillie's original proposal although differently effected. Nevertheless, it was the supposed association of eclogite with glaucophane schist and their joint implication of exceptionally high-pressure metamorphism that continued to attract most attention. Hence the inclusion of Ballantrae in that category on such regional compilations as the UNESCO-sponsored

Metamorphic Map of Europe (Fettes, 1978). With this background the Knockormal GCR site is of great importance both for interpretations of the Ballantrae Complex itself and for assessments of its large-scale regional relationships.

Description

The general distribution of lithologies in the Knockormal area is shown in Figure 2.32. The first descriptions of the blueschist and eclogitic lithologies were given by Balsillie (1937) who referred to 'a large knoll of foliated hornblendite and smaragdite-eclogite' with adjacent garnetiferous glaucophane schists 'derived from hornblendites'. Subsequently, in their extensive and detailed study, Bloxam and Allen (1960) located several zones of glaucophane schist in the vicinity of a low ridge capped by 'a prominent crag of eclogite'. Bloxam and Allen admitted the possibility that the eclogite 'crag' 'may not be precisely *in situ*' but concluded from their observation of other small exposures nearby that the material is indeed in place. The present appearance of the 'crag' clearly confirms it as a glacial erratic and there are currently no other exposures of the same lithology nearby. In 1975, a series of shallow boreholes was drilled in the vicinity of blueschist and eclogite 'exposures' and proved till with large boulders to a depth of at least 4 m, thus raising further doubt over the provenance of the schist and eclogite (J.E. Dixon, reported in Smellie and Stone, 1984). However, Smellie and Stone also reported in-situ material revealed in trenching operations: a drainage trench to the west of the Lendalfoot–Colmonell road (Trench 1, Figure 2.32) exposed a sequence of glaucophane-bearing schists while a garnet clinopyroxenite similar to the eclogite was recovered from a temporary excavation a short distance east of the crag/erratic (Trench 2, Figure 2.32).

The field relationships between the various ultramafic lithologies are uncertain but the contacts between the ultramafic rocks and the blueschists all appear to be faulted. The blueschists are disposed along the sheared and faulted margins of a complex tectonic lens, which contains dolerite and gabbro together with serpentinized ultramafic rock, wherlite and various pyroxenites including the 'eclogite'. The schistose foliation trends approximately parallel to the margins of the gabbro–ultramafic lens and appears to be consistent between outcrops. The schists have developed by progressive deformation of spilitic lavas which form the bulk of the outcrop in the Knockormal–Carleton Mains area. Locally, for example 400–500 m WSW from Knockormal farmhouse, the spilitic lavas are interbedded with black siliceous mudstone and recrystallized limestone. For the most part the

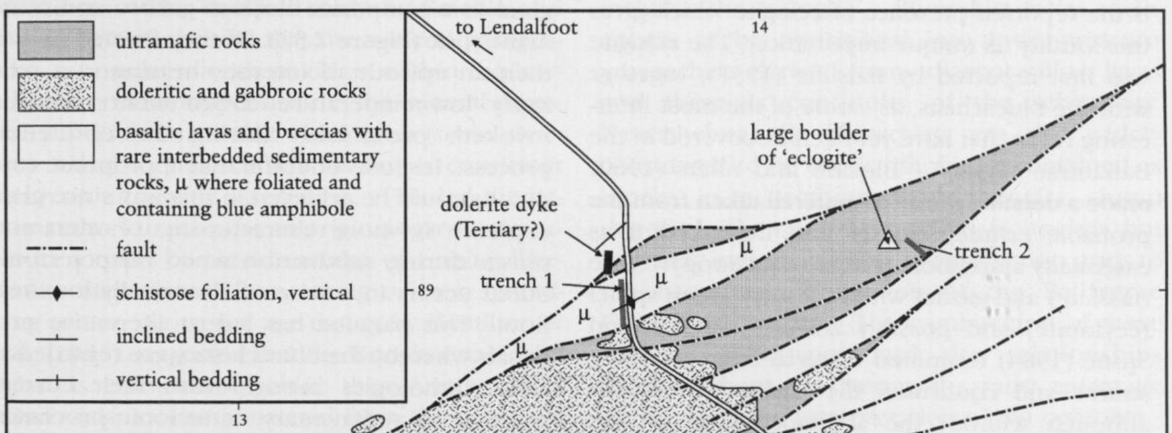


Figure 2.32 Map of the fault zone between Knockormal and Carleton Mains, after BGS (1988).

schists are of the greenschist facies but in some examples wisps of pale-blue glaucophane and/or crossite appear, apparently as replacements of actinolite or chlorite; rarely these may develop into macroscopic blue bands of fibrous crossite. These lithologies have been referred to as 'transitional glaucophane schists' by Bloxam and Allen (1960) and are the most common blueschist lithology. True glaucophane schist is rare and is restricted to narrow zones within the drainage trench and exposures 'on the crags near the Lendalfoot road' as reported by Bloxam and Allen (1960). In these examples the glaucophane forms a continuous foliation, apparently replacing chlorite and enclosing grains of epidote, titanite and albite with rare garnet present in places. In hand specimen the glaucophane schist is a hard, blue-grey rock with a silky lustre. Crossite amphibolites are interbanded with the glaucophane schists and are generally coarser grained with the crossite forming ragged fringes around crystals of green hornblende. A comprehensive account of the mineralogy and mineral chemistry of the Knockormal blueschists is provided by Bloxam and Allen (1960). All of the schistose lithologies show evidence of refolding, cataclasis and brittle deformation subsequent to the formation of the blue amphiboles.

Ultramafic lithologies exposed in the Knockormal area include serpentinized harzburgite, wherlite and clinopyroxenite. However, it is the reported presence of eclogite which gives this locality its unique importance. The eclogite was first reported by Balsillie (1937), together with the blueschists, as 'some of the most interesting rocks that have yet been discovered in the Ballantrae region'. Bloxam and Allen (1960) made a detailed study of material taken from the probable erratic boulder and described it as essentially amphibole (pargasite), clinopyroxene (fassaite) and garnet with accessory green spinel (ceylonite) and possibly zoisite. Smellie and Stone (1984) examined material from the same source and confirmed the earlier description although without the accessory spinel. The nearby excavation of a comparable lithology (Trench 2, Figure 2.32) was also reported by Smellie and Stone and is of importance in view of the uncertain origin of the originally described eclogite crag/boulder. A garnet clinopyroxenite from the NW end of Trench 2 consists of dark-green clinopyroxene (partially replaced by chlorite) and pale-pink garnet (generally fresh). Microprobe analyses presented by Smellie and Stone show that the mineral chemistry of the specimens from the erratic boulder and the trench is indistinguishable and suggests that the rocks had a common origin. The amphibole is tschermakitic hornblende, the pyroxene is fassaitic diopside and the garnet is almandine-rich pyrope. Small exposures close to the erratic boulder are of either serpentinized harzburgite or clinopyroxenite. A severely brecciated clinopyroxenite recovered from the SE end of Trench 2 consists of fresh, pale-green pyroxene occurring as large, ragged plates with strained and bent cleavage. This rock has apparently been affected by intense brittle fracture probably related to movement on the nearby fault (Figure 2.32).

Interpretation

The association of glaucophane schist and supposed eclogite has led to speculation that the Ballantrae Complex contains the remains of a high-pressure metamorphic belt (e.g. Fettes, 1978), with major implications for any interpretation of its origin. However, the ambiguous field relationships between the various lithologies and some unusual petrographical and chemical features have allowed the development of a number of alternative models.

The blueschist occurrences are all contained within an anastomosing fault zone and are marginal to a composite block of gabbro and ultramafic rock (Figure 2.32). If they formed *in situ* then an episode of intense shearing in a relatively low-temperature environment must be invoked, presumably during the obduction process. It seems doubtful that appropriate conditions could be achieved in this way since glaucophane is more characteristic of metamorphism during subduction when deep tectonic burial occurs in a zone of abnormally low heat flow. This paradox has led to alternative proposals whereby the blueschists were regarded as exotic lithologies brought into their current position by sedimentary or tectonic processes. Church and Gayer (1973) suggested that the blueschists (and the neighbouring 'eclogite') could be exotic clasts incorporated into a mélange which 'developed as a result of deformation beneath a forward moving thrust sheet of material previously produced by erosion from the leading edge of the self same nappe in an olistostrome environment.' These authors were particularly impressed by similarities between the Knockormal situation and more extensive analogues in the Newfoundland ophiolites. A different exotic origin was preferred by Bloxam (1980) who described the blueschists as tectonic inclusions contained within serpentinite; 'samples of Precambrian crust dismembered and caught-up in the serpentinite during its cold diapiric rise and tectonic emplacement.' The similarities of these proposals with the original deductions of Balsillie (1937) are striking and perhaps gain further support from the occurrence of blueschist clasts in undisputed mélange deposits elsewhere in the Ballantrae Complex (e.g. the Pinbain section within the Slockenray Coast GCR site).

The 'eclogite' has received even more attention than the blueschists and has generated a range of interpretations. These include: an ultramafic lens within serpentinized harzburgite (Bloxam and Allen, 1960), subsequently refined as a high-pressure phase of a layered peridotite sequence (Bloxam, 1980); part of a high-grade metamorphic belt (e.g. Fettes, 1978); a tectonically isolated sliver of a dismembered metamorphic sole to the ophiolite comparable to the in-situ examples seen at the Knocklaugh GCR site (Spray and Williams, 1980; Treloar *et al.*, 1980); an exotic clast derived from a mélange deposit (Church and Gayer, 1973) comparable, as with the blueschists, to fragments of garnet-clinopyroxene rock seen in other Ballantrae Complex mélanges. The detailed study by Smellie and Stone (1984) showed that the mineral paragenesis is consistent with an origin within the lowermost crust or upper mantle; the mineral chemistry is dissimilar to that of eclogites (*sensu stricto*) found in blueschist belts but is consistent with an ultramafic association. This led the latter authors to propose an origin as garnet pyroxenite formed by partial melting of a rising lherzolite diapir beneath an Arenig oceanic spreading centre. However, age dating by Hamilton *et al.* (1984) suggested that the Knockormal 'eclogite' is considerably older than Arenig (early Ordovician). Using material from the originally described crag/boulder (which they considered to be a clast within a mélange) these authors obtained a Sm-Nd garnet–pyroxene–whole rock isochron of 576 ± 32 Ma, about Early Cambrian. This is the oldest age yet determined for any component of the Ballantrae Complex and strengthens those interpretations that see the 'eclogite' as an exotic clast derived from pre-existing oceanic crust.

Despite a plethora of research the origin and relationships of the Knockormal blueschist-'eclogite' assemblage remain uncertain. The current consensus favours an exotic origin as clasts within a mélange deposit but the relative importance of sedimentary versus tectonic processes has not been established and the ultimate origin(s) of the clasts (if that is what they are) remains unknown. However, it does seem clear that there is no genetic link between the blueschists and the 'eclogite' and their present juxtaposition must therefore be regarded as fortuitous.

Conclusions

The 'eclogite' and glaucophane schist (blueschist) lithologies at Knockormal are internationally known as unusual components of the Ballantrae Complex. The 'eclogite' is most likely to have originated as a garnet(-hornblende) clinopyroxenite segregation within upper mantle harzburgite. It may now occur as a tectonic inclusion within serpentinite. It is the oldest rock so far identified within the complex (*c.* 576 Ma) and may represent a fragment of the oceanic crust which formed the basement to the younger (*c.* 500–490 Ma) components. The high-pressure blueschists occur at the sheared and faulted margins of a lensoid, composite block of gabbro and ultramafic lithologies that includes the 'eclogite'. Blueschist facies rocks appear to be transitional into lower-pressure greenschists derived from adjacent pillow lavas and there is continuity of the schistosity. Nevertheless, the blueschists are also widely considered to be exotic lithologies contained as blocks in a tectonized sedimentary slump deposit. Their ultimate origin is unclear but they may indicate a phase of subduction prior to the late-Arenig obduction of the Ballantrae Complex ophiolite. The combination of these unusual lithologies and the relationships between them, are among the most enigmatic problems in the interpretation of the ophiolite.

GAMES LOUP (NX 103 880–107 882)

P. Stone

Introduction

At Games Loup a sub-vertical fault trends NE–SW and juxtaposes two components of the Ballantrae Complex ophiolite: pillow lavas, with the geochemical characteristics of an oceanic island arc, and a serpentinite body in which red and green varieties show an interfingering relationship. The ultramafic rock lies to the NW of the fault and forms part of the Northern Serpentinite Belt whereas, to the SE, the sequence of pillow lavas is assigned to the Balcreuchan Group (Figure 2.33). A very large displacement is required to bring together these two units, which originated at very different structural levels – the serpentinite as mantle ultramafic rock and the pillow lavas as sea-floor volcanic flows. However, the fault exposed is a brittle structure that probably formed late in the assembly of the complex as part of a plexus of faults that cut through its central part. Faults within this zone affect Old Red Sandstone strata confirming a continuing history of movement well after the early to mid-Ordovician obduction of the ophiolite. In this respect the fault exposed at Games Loup, which forms the local margin to the Northern Serpentinite Belt, contrasts starkly with the structure seen in the Knocklaugh GCR site farther NE which also illustrates the marginal relationships of the Northern Serpentinite. At Knocklaugh the faulted margin preserves slivers of high-grade metamorphic lithologies produced during obduction; these have been removed by the later generation of faults as represented by the Games Loup example.

The serpentinite to the NW of the fault has a number of distinctive features: Balsillie (1937) first drew attention to the foliation seen at a number of localities, including Games Loup, while Stone and Smellie (1988) described the veined relationships between green and red varieties. On the SE side of the fault the lava sequence provides good examples of pillow structure with only local brecciation and very little intercalated sedimentary rock. The geochemistry of the Games Loup lavas has been extensively studied and quoted in support of polygenetic interpretations of the Ballantrae Complex. The consensus view is that the lavas were erupted at an oceanic island arc (Wilkinson and Cann, 1974; Lewis and Bloxam, 1977; Thirlwall and Bluck, 1984; Stone and Smellie, 1990). In addition, Thirlwall and Bluck gave a Sm-Nd radiometric age of 476 ± 14 Ma for the lavas, compatible with other evidence of an early Ordovician age for the complex as a whole. The

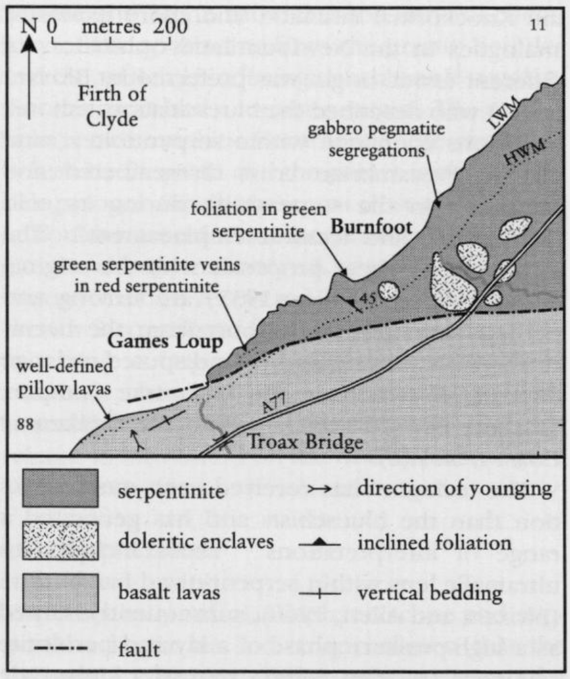


Figure 2.33 Map of the Games Loup area, after BGS 1:25 000 special sheet, Ballantrae (1988) and unpublished data.

outcrop of the Games Loup lava sequence continues southwards and its southern margin is included within the Balcreuchan Port to Port Vad GCR site.

Description

The serpentinite at Games Loup forms a series of low, rocky sea stacks and foreshore exposures on the NW side of the prominent fault gully. Adjacent to the fault the largest of the rocky areas consists mainly of red serpentinite but it is cut by prominent veins of green serpentinite up to about 15 cm across (Figure 2.34). The red serpentinite is slightly finer grained than the green variety but, that apart, the only difference is the dissemination of haematite in the serpentine crystal mesh, which imparts the red colouration. Both varieties were originally dunites consisting mainly of olivine with only a very sparse scattering of relict orthopyroxene and chrome-spinel grains. Running parallel to the vein margins are subordinate, very thin (mm-scale) veinlets containing fibrous chrysotile; these form an outer zone extending out for several centimetres from the principal vein margins (Figure 2.34).

Northwards, over about 50 m away from the fault, the proportion of green serpentinite increases until it is the red variety that appears to form sporadic veins. Other changes over the same interval include an increase in orthopyroxene relics (i.e. a more harzburgitic protolith) and the appearance of a faint and patchy foliation striking about NW and dipping moderately to the SW. Farther on, towards Burnfoot (108 882), highly altered doleritic enclaves several metres across occur within the green serpentinite, which there is also host to rare segregations of gabbroic pegmatite similar in lithology to that forming Bonney's Dyke within the Slockenray Coast GCR site.

There is no fabric development or change in the appearance of the serpentinite close to the fault zone, which forms a shingle-filled gully several metres across. The SE side is a substantial cliff of brecciated lava. Some of the brecciation might be of original autoclastic origin and some volcaniclastic material may be present but much of the texture seen probably arose as a brittle response to faulting. The fault seen here was probably initiated late in the structural history of the Ballantrae Complex and was not part of the deep-seated ductile fault system developed during ophiolite obduction.

A few metres south, away from the fault zone, most evidence of brecciation is lost and well-defined pillow shapes dominate the lava sequence. The pillows are generally relatively small with long axes ranging up to about 1 m. There are no sedimentary interbeds but the regular trend of the pillows defines an original horizontal layering, now steeply inclined, striking approximately NE and younging towards the NW. Vesicles are generally sparse. The lava is tholeiitic basalt, now extensively spilitized. A fine-grained matrix contains a scattering of small plagioclase and unusually fresh augite phenocrysts which, in some specimens, form as much as 15% of the rock. Sporadic pseudomorphs of chlorite and serpentinite, usually reddened by iron oxide, are probably the remains of original olivine phenocrysts. Other secondary minerals present include titanite, epidote and prehnite, probably generated by low-grade bur-

Figure 2.34 Green serpentinite veins cutting a red serpentinite host at Games Loup. Chrysotile veinlets are developed parallel to the vein margins. (Photo: BGS no. D3345.)

ial metamorphism. Since all of the margins of the Games Loup lava sequence are either faulted or unexposed it is not possible to estimate the original thickness; at least 250 m are now preserved.

Interpretation

The fault at Games Loup separates spilitic pillow lavas from ultramafic rock and as such forms the local margin of the Northern Serpentinite Belt within the Ballantrae Complex ophiolite. Elsewhere in the complex, as for example the Knocklaugh GCR site, this margin is marked by a zone of intense ductile faulting generated during the obduction of the ophiolite and the resulting juxtaposition of mantle and upper crustal lithologies. At Games Loup only brittle deformation is seen and this particular structure can best be regarded as a post-obduction fault locally cutting out the original structural boundary between the two Ballantrae Complex components.

The green serpentinized harzburgite and dunite seen in the Games Loup section are fairly typical of the common rock types within the Northern Serpentinite as a whole; the veined relationship with the red variety is, however, quite unusual. It may reflect original differences in the ultramafic protolith that have survived or been accentuated by serpentinization or alternatively it may be an entirely secondary effect developed during serpentinization. The subsidiary chrysotile veins might support the latter since their fibrous nature would require formation of the serpentine minerals in a dilational environment. The significance of the sporadic foliation is also uncertain. It is picked out by an elongation of the serpentine crystal lattice but it is not clear whether this reflects an original fabric in the protolith (hence perhaps a mantle tectonite) or whether subsequent deformation has been imposed following serpentinization.

The Games Loup lava sequence is exposed as a strike section south-westwards for about 750 m to the Balcreuchan Port area where it is terminated by a major fault zone. This part of the sequence contains an important lithological variation and falls within the Balcreuchan Port to Port Vad GCR site. Exposure is very poor inland but within 500 m of the coast volcaniclastic breccia and sandstone, containing only aphyric clasts, appear to be dominant. This transition may be stratigraphical, with the Games Loup lavas overlying an older clastic succession, or a fault may intervene. In a wider context the relative freshness of the lavas has allowed their use in geochemical characterization of geotectonic settings of eruption during the evolution of the Ballantrae Complex. Overall this has proved a controversial exercise but for the Games Loup lavas, almost uniquely, a succession of studies based on different element combinations has produced a consistent result (see the Balcreuchan Port to Port Vad GCR site report). Eruption of the lavas at an oceanic island arc was first proposed by Wilkinson and Cann (1974) based largely on Ti–Zr–Y–Nb relationships. A similar conclusion was then reached by Lewis and Bloxam (1977) using rare-earth element distributions. Thirlwall and Bluck (1984) used multi-element comparisons coupled with an assessment of Sr and Nd isotope ratios to interpret the lavas as primitive island-arc tholeiites similar to modern examples from the South Sandwich Islands. Data presented by Stone and Smellie (1990) are almost indistinguishable from those of Thirlwall and Bluck. Overall there is a selective enrichment in large ion lithophile elements (e.g. Rb, K, Ba, Sr) and a marked depletion in the high field strength elements (e.g. Y, Zr, Ti, Nb).

A Sm-Nd age of 476 ± 14 Ma was obtained by Thirlwall and Bluck (1984) from internal isochrons on the lavas and their separated clinopyroxene phenocrysts. The relatively large error overlaps with most other age determinations from the Ballantrae Complex and gives a time range compatible with its generally accepted early to mid-Arenig generation.

Conclusions

The Games Loup GCR site exposes several important geological features pertinent to interpretation of the Ballantrae Complex as an ophiolite. At Games Loup itself, the SE margin of the Northern Serpentinite Belt is a late brittle fault that juxtaposes the ultramafic rock against pillow lava. This structure has locally cut out the major ductile shear zone, which originally brought the two lithologies together during obduction of the ophiolite. To the north of the fault, serpentinized harzburgite and dunite occur as distinct red and green variants that have an interfingering relationship. In the largest exposures, adjacent to the fault, red serpentinite is dominant and is cut by veins of green serpentinite; to the north the proportion of the green variety increases sharply. South of the fault a sequence of basaltic pillow lavas has the tholeiitic geochemical characteristics of a primitive, oceanic island arc. The lavas have suffered relatively little alteration which has allowed their use in producing a Sm-Nd radiometric age of 476 ± 14 Ma, compatible with the Arenig age of the Ballantrae Complex as a whole.

BALCREUCHAN PORT TO PORT VAD (NX 100 878–093 869)

P. Stone

Introduction

Balcreuchan Port and Port Vad are both small embayments, backed by steep cliffs, on the northern flank of Bennane Head. The coastal section in this area affords magnificent exposure through part of the volcanosedimentary Balcreuchan Group in the central zone of the Ballantrae Complex ophiolite. The features preserved were fundamental in persuading Bonney (1878) that the rocks have a volcanic origin and an extensive description in those terms was provided by Peach and Horne (1899). A sequence of early Ordovician pillow lavas contains the very unusual lithology boninite, erupted only at primitive oceanic island arcs. This is the only unambiguous occurrence of boninite lava so far recorded from Britain. The island-arc lavas are faulted against a more disparate lava sequence believed to have originated at a within-plate, oceanic hot spot. Pillow structure is particularly well developed in the within-plate lavas, some of which are remarkably plagioclase-phyric. Interbedded sedimentary strata are sparsely fossiliferous, establishing an early to mid-Arenig age range and allowing structural imbrication of the sequence to be proven. Quite apart from its geological attractions Balcreuchan Port also features in Scottish legend as the lair of Sawney Bean and family, the infamous cannibal tribe finally brought to account and executed *en masse* in 1604 by personal command of King James VI.

The ophiolitic interpretation of the Ballantrae Complex (e.g. Church and Gayer, 1973) regarded the lavas as oceanic lithosphere and much subsequent work has been geochemically orientated in attempts to test this hypothesis. Wilkinson and Cann (1974) pioneered this approach, concluding that the Bennane Head section contains island-arc (at Balcreuchan Port) and ocean-island, hot-spot (south from Balcreuchan Port towards Port Vad) basalt lavas; a faulted relationship was inferred. Lewis and Bloxam (1977) also found a preponderance of island-arc lava types in the section but were also impressed by the apparent conformity of the sequence, placing it within a 4 km-thick lava pile. They considered this great thickness to be most compatible with an island-arc eruptive environment and regarded the hot-spot basalts as anomalous components within a single succession. A more sophisticated geochemical study by Thirlwall and Bluck (1984) confirmed the original conclusions of Wilkinson and Cann and re-established the fault juxtaposition of two very different lava sequences. The island-arc associations of the Balcreuchan Port rocks have also been confirmed by the discovery of characteristic boninite lavas (Smellie and Stone, 1992; Smellie *et al.*, 1995).

Graptolites had been recorded from sedimentary interbeds by Peach and Horne (1899) and assigned a generally mid-Arenig age. Further discoveries and a review of old collections allowed Stone and Rushton (1983) to establish biostratigraphical control on the lava sequence, which they showed to range from the early to the mid-Arenig. Despite the apparent conformity and uniform younging sense the biostratigraphy shows repetitions, and the structural imbrication of an originally relatively thin lava sequence seems the most likely explanation.

A detailed description of the Bennane Head section is given by Stone and Smellie (1988) and a field guide is provided by Stone (1996).

Description

The steep sea cliffs surrounding Balcreuchan Port expose an extensive array of pillow lavas intercalated with sporadic more massive lava flows. There is some autobrecciation of the pillows but intercalated sedimentary rocks are notable by their absence. The eastern margin of the bay coincides with a major N–S fault which, at sea level, juxtaposes the pillow lavas in the cliff against highly altered ultramafic rock forming the foreshore exposures. Alteration in the ultramafic rock lessens across the bay so that on the SW side serpentinized dunite and harzburgite are exposed in the intertidal zone. The cliff sections show no signs of ultramafic rock and a sub-horizonal structural contact is envisaged, between serpentinite below and pillow lava above, cut off by the N–S fault on the east side of the bay (Figure 2.35). An added point of interest in the foreshore exposures is a well-exposed Palaeogene dyke trending just east of north; it is about 50 to 60 cm wide with amygdales concentrated into zones parallel to the dyke margins.

The pillow lavas forming the Balcreuchan Port cliffs range up to about 1.5 m across but most are relatively small (less than 1 m). They are only sparsely vesicular. Small phenocrysts of plagioclase are much in evidence, accompanied by clinopyroxene phenocrysts (probably augite) and rare pseudomorphs after olivine; the matrix is dominated by altered glass. Spilitization has caused the alteration to albite of the originally more calcic plagioclase while low-grade metamorphism has produced the secondary mineral assemblage titanite-epidote-prehnite. The lavas are tholeiitic basalts forming a continuous sequence with those farther NE along the coast at the Games Loup GCR site. There, the relative freshness of the clinopyroxene phenocrysts allowed Thirlwall and Bluck (1984) to obtain a Sm-Nd age of 476 ± 14 Ma.

One very important lithological variation occurs in the sequence adjacent to the fault on the east side of Balcreuchan Port. Around the high-water mark a thickness of 5–10 m of lava shows only poorly defined pillows and is less porphyritic than most of the sequence. In thin-section the rock proves to have more glassy matrix and the scarcity of plagioclase microphenocrysts is confirmed. Conversely, clinopyroxene and olivine phenocrysts are slightly more abundant. Geochemical data show relatively high SiO_2, MgO, Cr and Ni but relatively low Al_2O_3 when compared to the normal tholeiites. These lavas are boninites (Smellie and Stone, 1992; Smellie *et al.,* 1995), a rare lithology that has no other unambiguous occurrence in Britain.

Westwards from Balcreuchan Port the lava sequence continues, with an increasing proportion of brecciated flows, as far as a second major N–S fault (Figure 2.35). Beyond this fault there is a marked change in the nature of the volcanic succession, which becomes lithologically much more variable. Adjacent to the fault about 30 m of sedimentary strata form the base of this varied sequence. Sandstone, shale, chert and conglomerate are all present and a graptolite fauna has been recovered that gives an early Arenig age (Stone and Rushton, 1983). The beds have a

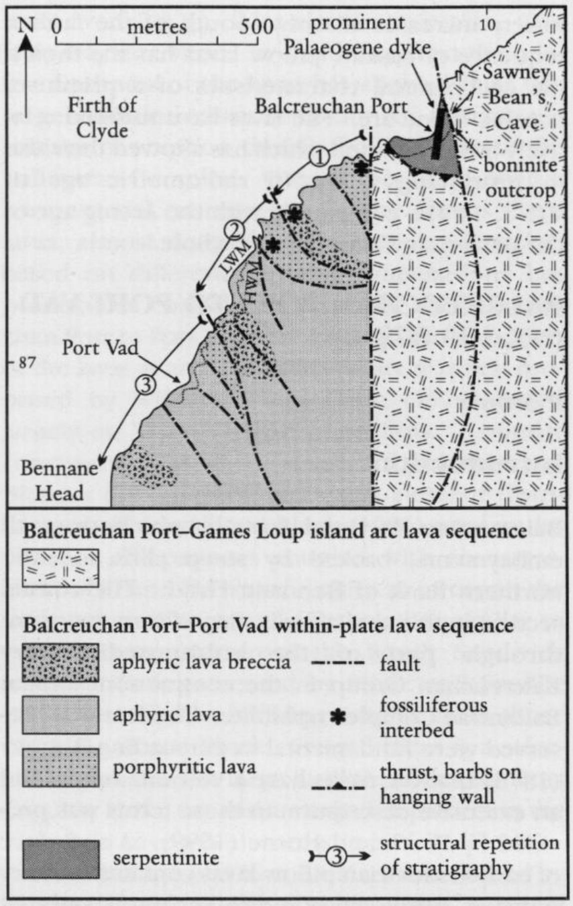


Figure 2.35 Map of the Balcreuchan Port to Port Vad area, after BGS 1:25 000 special sheet, Ballantrae (1988) and Stone and Smellie (1988).

N–S strike and are sub-vertical with sedimentary structures showing that they young towards the west. They are overlain conformably by exceptionally well-formed pillows of markedly porphyritic lava in which large plagioclase phenocrysts, up to 1 cm long and tabular in form, are contained in a pervasively reddened fine-grained matrix. The consistent asymmetry of the pillows (Figure 2.36), which range up to 2 m across, is consistent with the N–S strike, steep dip and westward younging deduced from the underlying sedimentary strata. The red, feldspar-phyric lavas form a unit about 150 m thick that is cut in places by thin dykes (less than 50 cm) of unreddened, aphyric basalt. These are the feeders for the overlying unit, which consists of about 150 m of dark grey-green aphyric lava in well-formed pillows generally smaller than those of the reddened lava type. Despite the presence of numerous minor faults in the section a conformable relationship can be seen between the two lava units, with aphyric pillows resting directly on reddened feldspar-phyric pillows. There is no intervening sedimentary material. Up sequence the aphyric pillows become increasingly more brecciated and interbeds of clastic rock appear. For one of these sedimentary intervals, just SW from the mouth of the Bennane Burn, a graptolite fauna indicates a mid-Arenig age (Stone and Rushton, 1983).

In the Bennane Head sector the succession described above (sedimentary beds (early Arenig)–red feldspar-phyric pillow lavas–aphyric pillow lavas–aphyric lava breccia–sedimentary interbeds (mid-Arenig)) is then terminated by faulting. However, biostratigraphical and lithostratigraphical comparisons suggest that it is repeated to the SW in a series of imbricate fault slices (Stone and Rushton, 1983). Throughout the imbricated succession the overall strike and dip and the younging direction remain constant. One repetition extends from red feldspar-phyric pillow lava just SW of Bennane Burn to brecciated aphyric pillow lava NE from Port Vad. At this point faulting re-introduces the red feldspar-phyric lava and another repetition of the sequence continues thence around Bennane Head (Figure 2.35). From Port Vad this highest structural repetition continues farther up the sequence with a much greater thickness of aphyric lava breccia preserved. It is this lithology (Figure 2.37) that forms the steep cliffs of Bennane Head itself. Ultimately the stratigraphy continues southwards to the Bennane Lea GCR site where the youngest part of the succession is exposed.

Interpretation

The lava sequence in the Balcreuchan Port to Port Vad area can be divided into two parts on lithostratigraphical grounds. The north-eastern sector consists exclusively of pillow lava and lava breccia with little obvious lithological variation and a complete absence of sedimentary interbeds. In contrast, the south-western sector contains an alternation of aphyric and reddened feldspar-phyric pillow lavas with abundant sedimentary intercalations. From the available evidence their ages are indistinguishable: the north-eastern lavas have a Sm-Nd radiometric age of 476 ± 14 Ma, the south-western lavas are early to mid-Arenig on biostratigraphical grounds. They are divided by a major N–S fault.

In a series of geochemical studies the Balcreuchan Port to Games Loup (north-eastern) lavas have consistently been interpreted as the

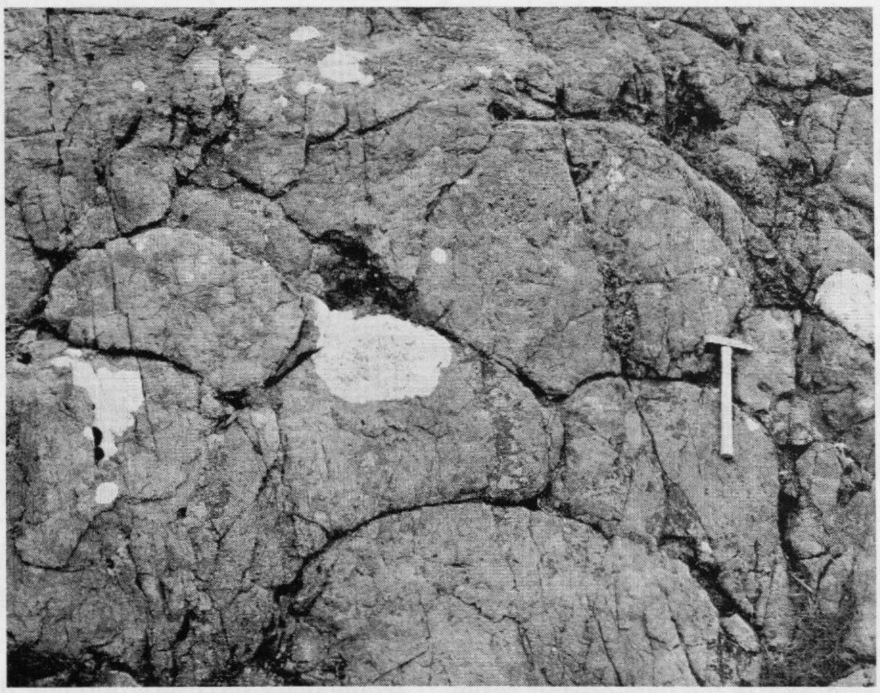


Figure 2.36 Large, well-formed pillows of reddened, plagioclase-phyric basalt exposed SW of Balcreuchan Port. (Photo: BGS no. D3585.)

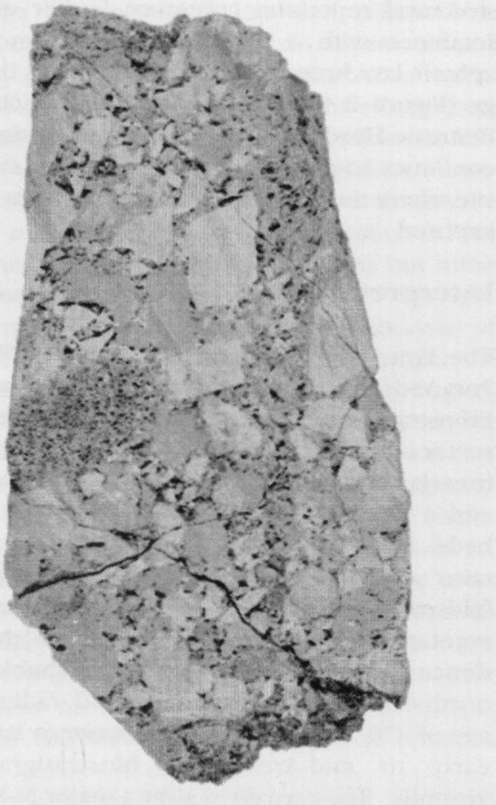


Figure 2.37 Volcaniclastic breccia of aphyric and vesicular lava clasts from Bennane Head. The long axis of the sample is 165 mm. (Photo: BGS no. MNS3838.)

products of eruption at an oceanic island arc (Wilkinson and Cann, 1974; Lewis and Bloxam, 1977; Thirlwall and Bluck, 1984). Most of the analyses published by these authors were derived from the Games Loup end of the section, now designated as a separate GCR site. More recent results from the Balcreuchan Port area (Stone and Smellie, 1990; Smellie and Stone, 1992; Smellie *et al.*, 1995) have confirmed the bulk of the lava sequence as of low-Ti tholeiitic character, and have also established the presence of the boninitic lavas. Examples of the geochemical discriminations used, based on abundances and ratios of trace elements such as Ti, Y and Zr, are shown in Figure 2.38a.

Despite their relative scarcity the boninites are of great importance in interpretation of the Ballantrae Complex because of the unique combination of circumstances involved in their petrogenesis. Important factors include: they are found exclusively in the intra-oceanic realm; their major element contents resemble those of primary magmas derived by partial melting of strongly depleted sources; the partial melting may occur at abnormally high geothermal gradients at relatively shallow levels. In most modern examples a general association with supra-subduction zone extension is apparent and in this respect the interbedding of the boninites with 'normal' island-arc tholeiites is important. Hence, arc-splitting during the initiation of a back-arc basin was invoked by Smellie and Stone (1992). After more detailed analytical work Smellie *et al.* (1995) noted that the low-Ti tholeiitic rocks, an unusual suite in their own right, shared some of the boninitic characteristics, such as the high Cr content (Figure 2.38b) and could not be definitively distinguished from basalts erupted in back-arc basins. A solution was proposed whereby the distinctive features of the combined sequence arose from mantle source heterogeneity caused by metasomatism above a subducting slab. Whatever the detailed mechanism the association with an intra-oceanic island arc seems unequivocal.

The more varied volcanosedimentary succession extending SW from Balcreuchan Port towards Port Vad has also been included in a number of geochemical studies (Wilkinson and Cann, 1974; Lewis and Bloxam, 1977; Thirlwall and Bluck, 1984; Stone and Smellie, 1990). The geochemical results have been consistently indicative of eruption in a within-plate environment above a mantle hot spot with the lavas forming a Hawaiian-type ocean island (Figure 2.38a). Controversy was introduced into the interpretation by Lewis and Bloxam who regarded the whole Ballantrae Complex succession as essentially conformable and abnormally thick, over 4 km and dominated by island-arc lavas. This apparently great thickness could be best accommodated in an island arc environment and so the Balcreuchan Port to Port Vad results were taken to be anomalous. The dilemma was resolved by recognition of structural imbrication of the within-plate sequence (Stone and Rushton, 1983). The nature of both the intercalated sedimentary rocks and those farther south at the top of the sequence (within the Bennane Lea GCR site) suggests deposition in fairly deep water and the reddening of the porphyritic lavas may have been caused by lengthy sea-floor weathering. This contrasts with the shallow-water environment deduced for the volcanosedimentary succession forming the Slockenray Coast GCR site, which contains geochemically indistinguishable (but largely unreddened) lavas

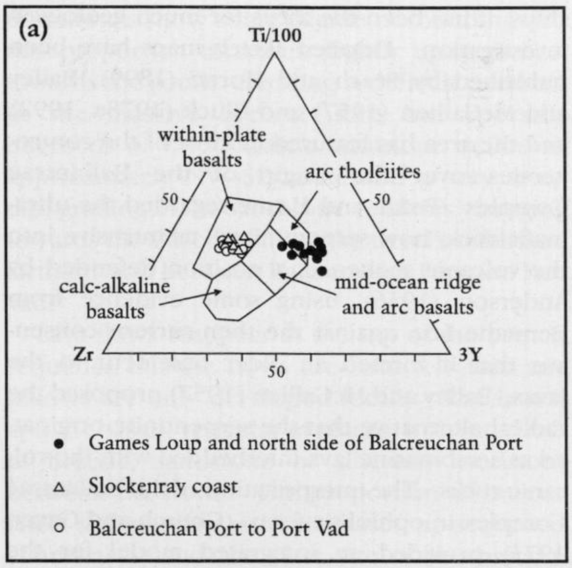
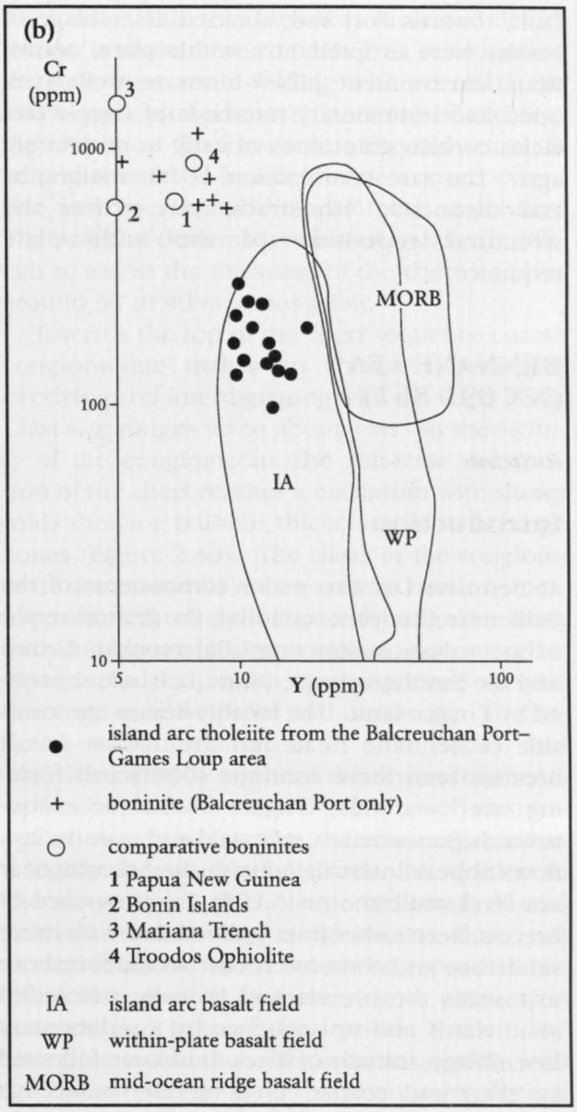


Figure 2.38 Geochemical discrimination diagrams for Ballantrae Complex basalt lavas:
(a) Ti–Zr–Y, fields from Pearce and Cann (1973).
(b) Cr–Y, showing comparison with representative boninites after Smellie and Stone (1992); fields from Pearce (1982).



of the same early Arenig age. It would seem reasonable to regard these two successions as facies variants formed coevally at the same ocean island.

Between Balcreuchan Port and Port Vad the N–S fault zone separating the two lithologically and geochemically distinct lava sequences (Figures 2.35 and 2.38) intersects the coast at 0971 8751, about 100 m west of the SW corner of Balcreuchan Port. It has had a polyphase history of movement but is unlikely to be the original structure formed during the imbrication of the lava sequence, which was probably a low-angle thrust. Some trace of this may be preserved to the immediate east of the main fault in a small and complex subsidiary fault-block. On the west side of the main fault a pronounced swing of strike suggests a sinistral movement laterally. However, in regional terms the fault is part of a plexus forming the eastern margin of a largely offshore Permo-Triassic basin and so has a significant downthrow to the west. Taking this latter movement alone, the island-arc sequence at Balcreuchan Port may have originally been at a greater structural depth within the obducted ophiolite thrust stack than the juxtaposed within-plate sequence forming Bennane Head.

Conclusions

The Balcreuchan Port to Port Vad sector of the Ballantrae Complex provides crucial data for its interpretation and is important internationally for the more general assessment of ophiolite generation and obduction models. Within the site, two dissimilar lava sequences are separated by a major N–S fault. To the east of the fault, at Balcreuchan Port, pillow lavas and lava breccias of uniform tholeiitic basalt were generated in an oceanic island arc. Rare intercalated boninite lavas are definitive of island-arc eruptions; they are a rare but petrogenetically important lithology and no other unambiguous examples are currently known from Britain. To the west of the fault, towards Port Vad, a varied assemblage of basalts were erupted in a within-plate, ocean-island environment; pillow-forms are well developed and sedimentary interbeds of deep-water facies contain graptolites of early to mid-Arenig age. The rare combination of biostratigraphy and distinctive lithostratigraphy proves the structural repetition of the within-plate sequence.

BENNANE LEA (NX 091 861)

P. Stone

Introduction

At Bennane Lea two major components of the Ballantrae Complex ophiolite, the central sector of the volcanosedimentary Balcreuchan Group and the Southern Serpentinite Belt, are separated by a major fault. The locality lies on the south side of Bennane Head and the massive basalt breccias seen there continue southwards forming steep sea cliffs. Higher in the succession, towards Bennane Lea, pillowed and massive lava flows appear intercalated with the breccias near sea level while the main cliffs are composed of breccia, here rather finer grained and with more sandstone and chert interbeds. In the foreshore exposures the breccia and lava are succeeded southwards and up-sequence by a sedimentary assemblage, initially of black mudstone followed by chert and coarse conglomerate; tuffaceous sandstone interbeds occur sporadically throughout the succession. The major fault, trending approximately E–W, cuts out the higher sedimentary beds and juxtaposes serpentinite against the chert and conglomerate. Coinciding with this abrupt lithological change is an equally abrupt change in topography. The softer serpentinite has been more readily eroded so that the steep sea cliffs and restricted foreshore typical of the volcanic rocks are replaced by a broad sandy beach and an extensive raised beach backed by relict sea cliffs cut in glacial till. Within the broad, sandy foreshore small exposures show that here Permian red sandstone (the fringe of a major, mainly offshore basin) overlies the serpentinite unconformably.

Since the coastal and foreshore exposures at Bennane Lea contain such a wide variety of lithologies with complex structural relationships, it has been the focus for much geological investigation. Detailed sketch maps have been published by Peach and Horne (1899), Bailey and McCallien (1957) and Bluck (1978a, 1992) and the area has featured in most of the controversies over the origin of the Ballantrae Complex. Peach and Horne regarded the ultramafic rock, now serpentinized, as intrusive into the volcanic sequence; a position defended by Anderson (1936), using some evidence from Bennane Lea, against the then-current consensus that it formed an older basement to the lavas. Bailey and McCallien (1957) proposed the radical alternative that the serpentinite originated as a submarine lava interbedded with the volcanic rocks. The interpretation of the Ballantrae Complex in ophiolitic terms (Church and Gayer, 1973) provided an integrated model for the igneous rocks, with the sedimentary assemblage regarded as a deep-water, slumped olistostrome. Detailed work by Bluck (1978a) built on this hypothesis, with evidence from the Bennane Lea conglomerates supporting his proposition that the complex had suffered considerable structural disruption prior to its final obduction. Bluck envisaged an island-arc–marginal basin environment for its original generation.

Description

Basalt breccias form the steep sea cliffs of Bennane Head and from there extend south towards Bennane Lea where they are succeeded by a mixed assemblage mainly composed of chert and conglomerate. A detailed geological sketch map of the Bennane Lea area is shown in Figure 2.39.

The Bennane Head breccias overlie aphyric pillow lavas seen at the southern margin of the Balcreuchan Port to Port Vad GCR site and extend that volcanosedimentary sequence southwards. A thickness of between 200 and 300 m is present and, for the most part, the breccias are very coarse with clasts up to 2 m across and commonly in the 2–10 cm range. They are composed exclusively of aphyric basalt although there is considerable difference in the vesicularity, with some clasts completely devoid of vesicles whereas others are scoriaceous (Figure 2.37). Although some clasts are partially reddened there are no clasts of the distinctive, reddened porphyritic basalt seen lower in the sequence towards Balcreuchan Port. The clasts are generally angular although a small proportion always show some degree of rounding and this proportion increases slightly up-sequence, towards the south. Bedding is difficult to detect in the massive breccia units but sporadic interbeds of volcaniclastic sandstone define an approximately NW–SE strike and a very steep dip; grading in the sandstone confirms younging towards the SW, continuing the trend established farther north between Balcreuchan Port and Port Vad.

Lava flows reappear at the top of the breccia sequence but the flows are relatively thin and pillows are only developed locally. The lava is aphyric basalt, vesicular in places, with interbeds of breccia, sandstone or siliceous mudstone between the flows. The proportion of intercalated sedimentary rock increases up-sequence, southwards, and the clast size of the breccias reduces. Bluck (1992) reported that the composition of the breccias also becomes more variable with the inclusion of acid volcanic lithologies. Near the top of this lava-dominated unit, which is about 100 m thick, black siliceous mudstone is particularly well developed and contains a mid-Arenig graptolite fauna (Peach and Horne, 1899; Stone and Rushton, 1983). The mudstone contains chalcopyrite and fracture surfaces within it are coated by green, secondary copper minerals.

Minor faulting complicates this part of the sequence and across the faulted zone there is a change of strike to almost E–W, the dip remaining steep. However, the sedimentary interval continues and any stratigraphical break is thought to be fairly small. It is also possible that there is a minor intra-formational unconformity at this point but the evidence is obscure and inconclusive. Immediately south of the fault pale-green volcaniclastic sandstone is interbedded with chert and the relationship between the two lithologies suggests that the sandstone remained fluid after the chert had become partially lithified. The resulting soft-sediment deformation structures were first described and illustrated by Bailey and McCallien (1957).

Southwards and up-sequence the proportion of chert increases sharply. It is reddish-brown in colour and individual beds rarely exceed 10 cm in thickness; radiolaria were described by Aitchison (1998) and Bluck (1992) reported that locally the chert may contain glass shards. Boudinage and other features characteristic of soft-sediment deformation are widely present and increase in both size and frequency towards the south. This trend towards increased deformation is made more complex by a concomitant increase in tectonic folding with several large fold hinges, plunging steeply seawards, affecting the cherts on the foreshore. The old sea cliffs behind the raised beach provide a more extensive and spectacular view of these structures. The range of deformation features makes it difficult to assess the thickness of the chert unit but around 30 to 40 m seems likely.

Towards the top of the chert sequence coarse conglomerate makes an abrupt appearance, overlying and interdigitating with the chert beds. Clast size ranges up to about 1 m. In the vicinity of the conglomerate the soft-state deformation of the chert reaches a maximum with slump folds showing bulbous thickening of their hinge zones (Figure 2.40). The clasts in the conglomerate are mainly of either aphyric basalt (generally rounded) or chert (generally angular), both with an obvious local provenance, but rarer clast types include a pinkish, coarse-grained syenitic lithology, massive pyrite and carbonate. The latter is of some interest in that Bailey and McCallien (1957) reported dissolving fragments in hot acid and recovering grains of chrome-spinel; the carbonate was therefore regarded as a highly altered ultramafic rock. In addition to the above Bluck (1992) reported clasts of acid volcanic rock. The conglomerate does not extend inland from the foreshore exposures but Bluck described it as increasing in proportion towards and beyond the low-tide mark. The shoreline would therefore seem to coincide with an interfingering zone between two distinct lithofacies.

The southernmost of the conglomerate beds is only a few metres short of the major fault juxtaposing this volcanosedimentary succession and ultramafic rock of the Southern Serpentinite Belt. In this interval lies an enigmatic lithology that has some characteristics of an intrusive dolerite and some of a volcaniclastic sediment. Its northern margin appears to be in conformable contact with chert but its origin remains uncertain; Bluck (1992) refered to it as 'dolerite-tuff' and speculated that it might be a high-level sill, despite an absence of peperitic margins. The southern margin of this body is faulted against serpentinite with a thin zone of quartz-carbonate alteration along the contact. The serpentinite itself is pervasively reddened and contains large enclaves of altered gabbroic rock. These form small bosses between which expo-

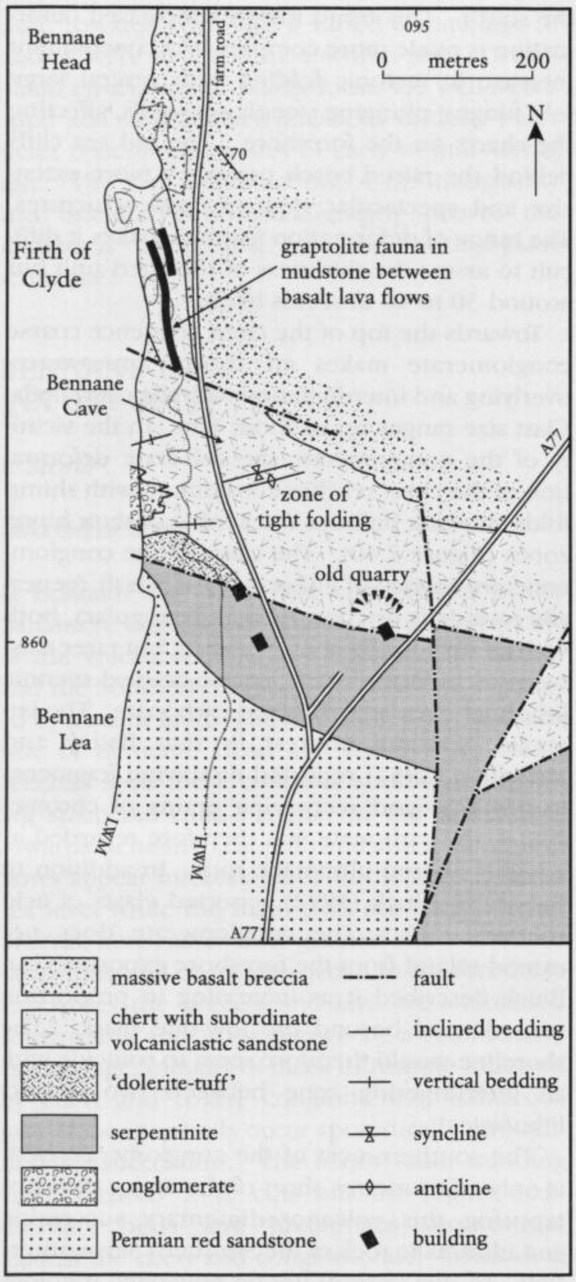


Figure 2.39 Map of the Bennane Lea area, after BGS 1:25 000 special sheet, Ballantrae (1988) and Stone and Smellie (1988).

sure of the serpentinite is only sparse. A few metres farther south the outcrop of ultramafic rock ends at the unconformable contact with Permian red sandstone.

Interpretation

The Bennane Lea section illustrates the dominantly sedimentary and volcaniclastic upper part of a Balcreuchan Group lava sequence. Coarse, oligomictic basalt breccia in the north of the section probably accumulated as flow-front talus and in terms of their petrography and geochemistry the clasts are identical to the subjacent lavas seen farther north at Port Vad. The presence of interbeds of graded volcaniclastic sandstones suggests deposition in relatively deep water but the rounding of some lava clasts probably arose through wave action in shallow water. The most likely combination of events would seem to be the eruption of aphyric basalt lava into shallow water, autobrecciation of the lava front (and perhaps some wave erosion) with partial rounding of some fragments by wave action, followed by avalanching of the talus accumulation into deeper water to cover basalts erupted earlier in the same eruptive episode. A few lava flows then extended across the accumulated breccia.

The association of these later lava flows with graptolitic mudstone is also indicative of a deep-water depositional environment but the evidence of the overlying chert–conglomerate sequence is ambiguous. In straightforward lithological terms the chert might be regarded as a deep-sea deposit but its extensive soft sediment deformation requires slumping to have occurred, presumably on an unstable slope. The interdigitating conglomerates with their rounded clasts appear to have slumped from a shallow-water environment and the glass shards in the chert indicate relatively proximal and probably subaerial, contemporaneous volcanicity. Cobbles and boulders in the conglomerate are derived from ophiolitic lithologies, including altered ultramafic rock, which suggests that obduction was already in progress when this part of the Ballantrae complex was deposited. Overall, a tectonically unstable environment is indicated, interpreted by Bluck (1978a) in terms of obduction within an active marginal (back- or intra-arc) basin. Developing this theme, Bluck (1992), emphasized the evidence for acid volcanicity and the abrupt facies changes, concluding that the Bennane Lea succession represents the rift facies developed during the splitting of a volcanic arc. The lavas to the north were regarded as part of the original arc structure. This interpretation was also influenced by the thickness of the succession which, at over 2 km, Bluck thought anomalous for anything but an island-arc environment. However, this figure was reached by including within the same suc-

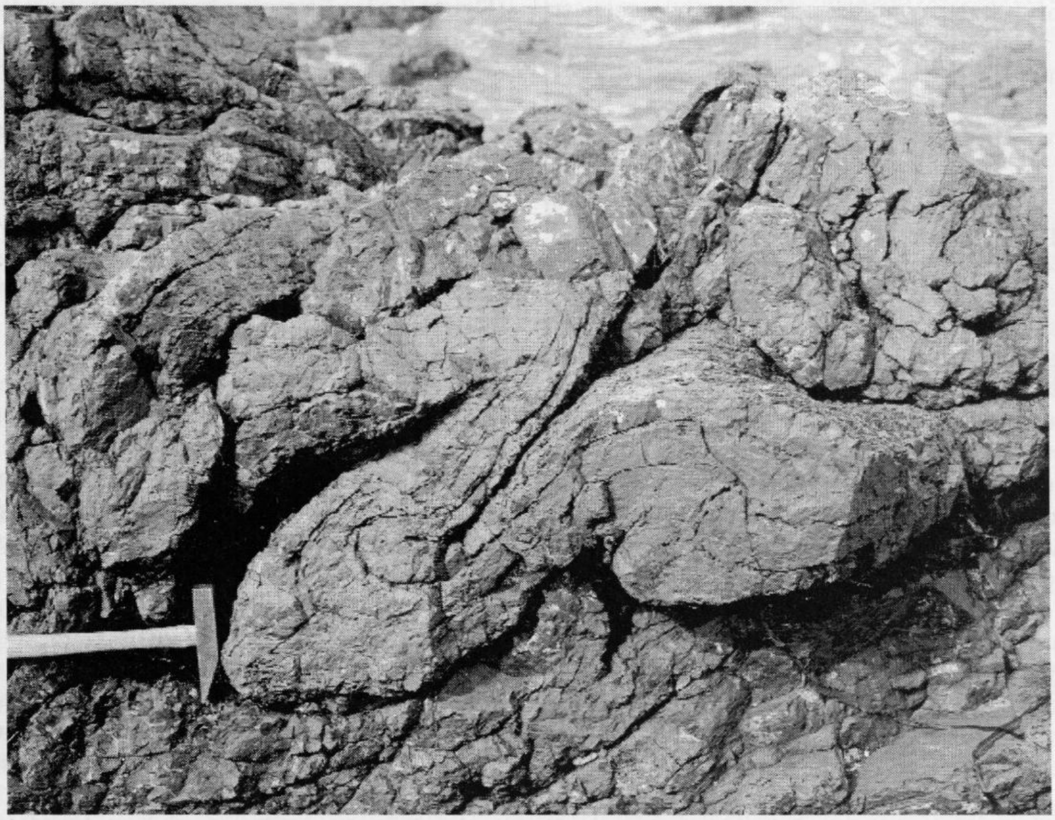


Figure 2.40 Slump fold in chert interbedded with conglomerate and volcaniclastic sandstone at Bennane Lea. (Photo: BGS no. D3333.)

cession the whole coastal outcrop between Bennane Lea and Games Loup, which a series of geochemical studies has shown to be polygenetic (Wilkinson and Cann, 1974; Thirlwall and Bluck, 1984; Stone and Smellie, 1990). On the geochemical evidence (Figure 2.38) the Bennane Lea strata overlie within-plate, ocean-island basalts with both the breccia clasts and the lavas above the breccia showing the same geochemical characteristics; on this basis none of these lavas can have any direct connection with an island arc. Further, when the evidence for structural imbrication presented by Stone and Rushton (1983) is taken fully into account the exposed thickness of the succession seen in part at Bennane Lea does not exceed 900 m (Stone and Smellie, 1988). The dilemma remains unresolved but the problem cannot be addressed by evidence from the Bennane Lea GCR site taken in isolation; comparison with the contiguous outcrop, as represented by the Balcreuchan Port to Port Vad and Games Loup GCR sites is important. It is also useful to compare the Bennane Lea succession with the within-plate, ocean-island volcanosedimentary rocks of the Slockenray Coast GCR site, which show some remarkable similarities.

Whatever the correct interpretation of the Bennane Lea volcanosedimentary sequence, at its southern margin it is faulted against serpentinized ultramafic rock. The fault is one of the major structures of the Ballantrae Complex juxtaposing mantle rocks of the Southern Serpentinite Belt and upper crust of the Balcreuchan Group, represented here by the chert–conglomerate–volcaniclastic assemblage. Movement on this scale seems unlikely to be achieved in a single phase of faulting and a history of periodic re-activation is probable. However, in this context the narrow zone of quartz-carbonate alteration adjacent to the fault is significant. This developed as a side-effect of serpentinization at a relatively early stage in the history of the complex. Its presence suggests that the Bennane Lea Fault was also an early-formed feature.

Conclusions

The Bennane Lea GCR site reveals several unique features of great importance in the interpretation of the Ballantrae Complex as an ophiolite. Some aspects remain controversial. The exposed strata form the highest preserved beds of a volcanosedimentary sequence dominated elsewhere, in its lower part, by basalt lava. These lavas have the geochemical characteristics of within-plate, ocean-island eruption, a feature shared by large clasts within breccias and sporadic lava flows within the Bennane Lea sector. However, the unusual chert–conglomerate–volcaniclastic rock assemblage has been interpreted, on mainly sedimentological grounds, as being more compatible with a volcanic arc environment. The compositional range of clasts in the conglomerate suggests that ophiolite obduction was already in progress when it was deposited; graptolites in subjacent mudstones give deposition a maximum age of mid-Arenig. At the southern margin of the site, volcanosedimentary upper crustal rocks are faulted against serpentinized mantle rocks, across one of the major lithological breaks within the Ballantrae Complex. Bennane Lea provides rare exposure across a fault of this magnitude.

SOUTHERN UPLANDS

SGAVOCH ROCK (NX 075 810)

P. Stone

Introduction

The coastal sections on the mainland opposite the offshore Sgavoch Rock provide the finest array of pillow lavas to be seen in Britain. Individual pillow shapes range from almost spherical, through elliptical 'bolster' shapes into sheet flows; lava tubes can also be clearly identified with well-preserved pillow buds at their margins. The lava sequence is structurally confined between the Stinchar Valley Fault to the north and the Dove Cove Fault to the south (Figure 2.41). Extensive coastal exposure is continuous from the Sgavoch Rock area southwards to Downan Point but the outcrop narrows inland as the two faults converge towards the NE. These lavas have traditionally been associated with the Arenig (early Ordovician) Ballantrae ophiolite complex with which they are juxtaposed across the Stinchar Valley Fault. However, recent interpretations have given more weight to relationships at the southern margin of the Sgavoch–Downan lavas where the faulted base of the sequence is intimately associated with chert and shale containing a *gracilis* Biozone (Caradoc, mid- to late Ordovician) graptolite fauna. On this basis the lavas should be regarded as the earliest accreted unit within the Southern Uplands imbricate thrust belt. A poorly constrained radiometric (Sm-Nd) age of 468 ± 22 Ma (Thirlwall and Bluck, 1984) does not differentiate between the two alternatives. A summary of the debate is given by Stone and Smellie (1988) who introduced the lithostratigraphical name Downan Point Lava Formation.

The spectacular pillow structures within the Downan Point Lava Formation (DPF) were first noted and illustrated by Peach and Horne (1899) and subsequently discussed in more detail by Bloxam (1960). A more recent description of the Sgavoch Rock locality was given by Bluck (1992). The pillow lavas exposed preserve a range of features characteristic of submarine eruption and are arguably the best British examples of their kind; they also occupy an important position in terms of the regional geology of southern Scotland. To the north, the ophiolitic Ballantrae Complex was generated and obducted onto the Laurentian continental margin during the early Ordovician. To the south the Southern Uplands Terrane was sequentially accreted at the Laurentian margin during the late Ordovician and early Silurian. The DPF represents renewed volcanicity within or marginal to the Iapetus Ocean immediately prior to the initiation of Southern Uplands accretion and in this context the DPF represents the most extensive volcanic fragment preserved within the Southern Uplands Terrane.

Description

The geological setting of the Sgavoch Rock area is shown in Figure 2.41. The lavas and the associated breccias, cherts and shales accumulated in a submarine environment and it is the process of eruption under water that produces the characteristic pillow shapes. The pillows form as lava is squeezed out from points of weakness in the walls of lava tubes, a phenomenon known as budding. Each emerging tongue of lava is rapidly chilled by contact with the water and by the time it has grown to pillow size the hardening skin prevents further growth. The pillow may then break free of the lava front or be overtaken by new tongues, which similarly swell into pillows. Since the pillows accumulate in a semi-solid state they settle and mould around each other, finally solidifying into interlocking patterns. In an ideal example the top surface of each pillow is a convex dome whereas the base may be flat or irregular, moulded to the shape of the underlying surface; vesicles are typically concentrated towards the top of the pillow and in many of the examples seen they form prominent concentric zones. These features are illustrated in Figure 2.42.

The full array of pillow lava features is most spectacularly displayed on the small rocky headland due east from the Sgavoch Rock; note that the latter feature is covered at high tide. On the headland the pillow attitude shows bedding to be steeply dipping or vertical with an approximately NE–SW strike; the asymmetry of the pillows establishes that the original top of the lava pile lies towards the NW. Since the lavas were erupted onto a probably sub-horizontal sea floor they have therefore been rotated through 90° and now become sequentially younger seaward. The Sgavoch (DPF) pillows range from slightly elliptical with long axes of 20–50 cm through to larger, elongate 'bolster' shapes up to 2 m across. Within the pillow lava sequence there are other sheet-like lava bodies which probably represent original lava tubes carrying magma forward to the eruptive front. In some cases lava pillows can be seen budding from the extremities and top surfaces of the sheets in a remarkable illustration of their mode of formation. The proportion of unpillowed sheet-flows in the succession increases southwards towards Downan Point.

Several types of pelagic sediment are intimately associated with the lavas. Lenses of laminated black chert, up to 3 m long and 30 cm thick, occupy what were probably hollows in the lava pile topography and indicate local breaks in lava accumulation. Red chert or dark siliceous mudstone may fill gaps left between pillows and

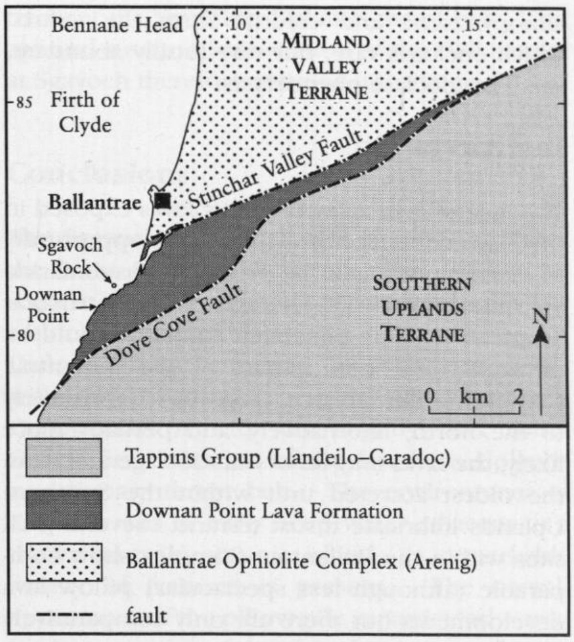


Figure 2.41 The position of the Sgavoch Rock GCR site within the Downan Point Lava Formation and its relationship to the Ballantrae Complex and the Southern Uplands Terrane.

in some cases the partial draining of pillows has left spaces that are now filled with either chert or laminated carbonate. Euhedral calcite crystals are seen growing from the pillow margins into the inter-pillow sediment and appear to have done so while the sediment was still soft, displacing sedimentary laminae. Carbonate is also the most common filling for the original vesicles.

Lava breccias are interspersed in the volcanic sequence and are usually dominated by small pillows as little as 10 cm in diameter. These show intact chilled margins and so are not clasts resulting from the disintegration of larger pillows although such clasts are also present. The breccias are commonly rich in fine-grained matrix and may have been emplaced as debris flows from the lava front. However, they are also pervasively bleached to a pale yellowish-green colour, probably by hydrothermal alteration, and the restriction of this alteration to the clastic zones means that an origin as intrusion breccia cannot be entirely ruled out.

The pillows all have green chloritic rims, which are the result of alteration of the original glassy chilled margins; the basalt forming the pillow cores is tholeiitic, fine grained, mostly aphyric and generally vesicular. Plagioclase, clinopyroxene and chlorite form the matrix, which encloses rare, but very locally abundant, small plagioclase phenocrysts.

Interpretation

The magnificent array of pillow lavas exposed in the Sgavoch section provides a rare opportunity to examine in detail the products of submarine volcanic eruptions. The age of this particular eruption remains uncertain: the lavas could be of Arenig age and a part of the Ballantrae Complex ophiolite that crops out immediately to the north; alternatively, and perhaps more likely, the lavas may be of Caradoc age and form the oldest accreted unit within the Southern Uplands imbricate thrust terrane. Several GCR sites within the Ballantrae Complex show comparable (although less spectacular) pillow lava developments but there are only comparatively meagre equivalents within the Southern Uplands. The DPF lava pile at Sgavoch has been rotated to a sub-vertical attitude but has otherwise escaped deformation. There is no penetrative cleavage and no evidence that the pillow shapes have been modified tectonically. The consistent westward younging of the steeply inclined flows militates against structural complexity introduced by folding or faulting. The metamorphic grade is very low. This fortunate state of affairs contributes to the value of the Sgavoch section as an ideal site for the investigation of submarine volcanism.

The submarine environment of eruption is not in dispute and broadly similar lava flow morphologies have been observed forming at mid-ocean ridges and around Pacific islands. The alternation of pillowed and sheeted flows reflects varying rates of extrusion on an unstable sea-floor, the more rapid the advance of the lava

Figure 2.42 A spectacular array of pillow lavas from the Downan Point Lava Formation exposed on the coast adjacent to the offshore Sgavoch Rock. The lavas are steeply inclined and slightly overturned. (Photo: BGS no. D1572.)

front the less opportunity there is for the production of pillows. Submarine slopes are indicated by a preponderance of jumbled, elongate pillows with the steeper slopes producing breccias as lava pillows and fragments cascaded downwards. The interpillow chert may represent either background pelagic sediment or be an essentially hydrothermal deposit; the laminated chert lenses are most likely to have a sedimentary origin and may even indicate intermittent turbidity current activity. A fairly deep water setting seems probable.

Several geochemical studies of the DPF pillow basalts have been carried out and the results were summarized and assessed by Thirlwall and Bluck (1984). A consensus view considers that they are most closely comparable to modern oceanic island, 'Hawaiian-type' basalts and thus a within-plate geotectonic setting has been proposed for their eruption. The source of the magma erupted at Sgavoch was therefore a mantle plume of some sort. Lavas of closely similar composition are found within the Ballantrae Complex at the Slockenray Coast and Balcreuchan Port to Port Vad GCR sites but are also known from small, scattered outcrops contained within fault zones defining the northern tracts of the Southern Uplands thrust belt. The age and structural association of the DPF basalts at Sgavoch therefore remains uncertain.

Conclusions

The Sgavoch Rock GCR site contains arguably the finest array of submarine pillow lavas to be seen in Britain. It is a spectacular locality of major volcanological interest. The lavas are of Ordovician age but may be associated either with the Ballantrae Complex ophiolite (early Ordovician) or with the slightly later (mid-Ordovician) initiation of the Southern Uplands accretionary thrust belt. The environment of eruption was in fairly deep water adjacent to a 'Hawaiian-type' oceanic island. A remarkable variety of features is present. The overall appearance of the pillow pile is particularly striking but a wealth of detail is also preserved: variable pillow shape clearly indicating the original top of the sequence, pillows budding from lava tubes, drained pillow cores filled with sediment, marked zonation by vesicles (relict gas bubbles) and interbedded lava breccias to list only the more obvious.