Excursion 25 Pinbain Block

Key details

Author	B.J. Bluck
Themes	1 Examination in detail of parts of the lava sequence from Kennedy's Pass to Pinbain with a view to determining the water depth in which they were deposited. 2 Examination of the contact between the lavas and the serpentinite. The lavas formed on the surface; the serpentinite formed at depths which may have exceeded 30 km, yet are both brought to the same level at Pinbain -implying that the Pinbain fracture has a considerable throw. 3 To evaluate the significance of the olistostromes at Pinbain, and to see how they may shed light on the overall evolution of the Complex.
Features	Lavas, pillows, hyalotuff deltas, volcanic breccias, faulting, olistostromes, cherts, black shales, dykes, contemporaneous faulting, slumping, soft sediment shearing.
Maps	O.S. 1:50 000 Sheet 76 Girvan B.G.S. 1: 50 000 Sheet 7 Girvan 1: 25 000 Sheets NX 08, 18 and 19 (in part) Ballantrae
Terrain	Rough shoreline with some scrambling
Time	8 hours; recommended short itinerary, 4 hours: localities 1, 3, 4, 12, 13.
Access	Fairly low tide recommended. (Coastal SSSI)

Itinerary

Locality 1. South end of Kennedy's Pass [NX 146 928]

Lavas and breccias (Figure 25.7), (Figure 25.8). Park cars south of Kennedy's Pass, walk 50 m to the north to dark-looking rocks on the foreshore; for coaches there is a larger car park on the seaward side of the road north of Kennedy's Pass. Beneath the unconformity which divides the Ordovician clastic sequence to the north from the Ballantrae complex to the south (see Excursion 30), are a series of tough, brittle, dark grey, orange and red lavas and breccias which form the local top to the Pinbain block. The Pinbain block comprises mainly albite-bearing altered basalts (spilites) most of which are pillowed.

The wave-washed surfaces at this outcrop are very polished and on them the dark lavas show indistict flow banding and occasional viscous, flow-banded folding. Some of the flow-banded lavas contain quartz, which in part is secondary; but some have >60% silicon dioxide, which is considerably higher than for an average basalt. The rocks are probably a mixture of dacitic and basaltic lavas which have been considerably altered by their contact with sea water. Acidic and intermediate lavas often brecciate when extruded into water, and this is a likely reason for the intense brecciation seen here.

Locality 2. [NX 143 923]: Interfingering lavas and sediments

This locality can be identified by the presence of an unofficial car park which is sited on the west side of the road, opposite the milestone 'Girvan 5; Ballantrae 8' (Figure 25.9). The outcrops are only barely seen at low tide. There are two types of lava exposed, the lower one is porphyritic, the upper is dark and aphyric. The dark lava is mainly massive, very fine grained and is partly replaced upwards (to the north) by conglomerates and breccias with a fine tuff matrix. Two thin units of pillow lava occur within these clastic rocks. Also, within the breccia unit there are layers of conglomerate, some having well rounded clasts which may be red in colour.

This locality shows the interaction between the sea and the lavas which flow into it. When lavas reach water they often brecciate, and the thick unit of breccia-conglomerate was probably produced in this way. The rounded clasts were produced during intervals when lava flowage had ceased and waves had time enough to work on the clast population before its burial. Some of these clasts are red because the lavas from which they have been derived have almost certainly been subject to sub-aerial weathering. During periods of strong lava flowage fragmentation of the lava flow takes place at the water's edge, and this may produce clasts so quickly that the waves have insufficient time to work on them and produce more rounded clasts. But if the extrusion is particulary rapid, then the lava extends beyond the water's edge and produces pillowed or massive lava flows (see (Figure 25.13) for explanation). This locality shows all of these features with respect to the dark, massive lavas.

The red tops to the lavas are almost certainly the result of subaerial weathering, having been eroded during a period of coastal retreat. Associated with these lavas are tuffs containing accretionary lapilli (Smellie 1984).

Locality 3. Slockenray headland [NX 149 919]: Hyalotuff deltas (Figure 25.10)

Park cars on the headland just above Slockenray Bay, at a bend in the road. Slockenray Bay is one of the most significant outcrops in the Pinbain Block. It was previously regarded as an Arenig vent, but is now thought to be a hyalotuff delta as it is clearly interstratified with the other lavas in the block and can be traced for some distance inland (Bluck 1981; (Figure 25.8)).

A detailed map of this headland is given (Figure 25.10), where the position of the car park is marked. From the car park the following may be observed; a steeply dipping distinctly porphyritic lava forms the headland: beneath this lava, and in the low ground of the bay to the south is the hyalotuff deposit which underlies the lava. The contact beween the two is within the low ground beneath the southern end of the headland. The lavas are replaced to the SSW by conglomerates, and they probably wedge out in this direction. The whole sequence is upward coarsening and is terminated by the porphyritic lava.

To the north of the headland (Figure 25.11) hyalotuff deposits overlie the porphyritic lavas, and interfinger with a dark aphyric lava-type.

Locality 4. Lavas (best seen at low tide). (Figure 25.10)

There are two lava types in this sequence, each of which comprises multiple flows. The main one is porphyritic with abundant phenocrysts >1 cm long of plagioclase (now mainly albite) arranged in a swirling fashion and suggesting alignment during turbulent viscous flow of the lava. The other is dark, aphyric and sometimes abundantly vesicular. This lava has also behaved in a plastic way; it has contorted margins against the porphyritic lava, which are lined with abundant vesicles left by the trapped gases; sometimes long deformed finger-like projections, and even detached irregular masses are totally enclosed in the porphyritc lava. It is clear that both lavas were extruded at the same Time at the boundary between them and (whilst they were flowing), each has injected into the other.

With Slockenray being near the contact between the two different types of lava, there must have existed on either side of this locality magma chambers each yielding a different lava. However, since both magma chambers were repeatedly producing lavas at the same time, they may have been responding to the same event-which is likely to have been structural.

Locality 5. Cross stratified hyalotuffs (Figure 25.10)

These deposits comprise rounded and angular clasts of dark basalt with a texture identical to the porphyritic lavas which immediately overlie them. Some of the clasts are whole pillows, some are very angular vesiculated fragments and both contain abundant phenocrysts. The clasts range in size from >80 cm to sand sized grains and have a matrix which is a brown coloured mass of chloritised volcanic glass with isolated long phenocrysts of labradorite and by townite. Some of these crystals have been broken and then welded by the glass implying explosive activity which fragmented the grains and the rapid invasion of the broken mass by the hot lava. From the extreme angularity of many of the clasts and the pristine nature of the crystals, it is clear that this tuff has suffered the minimum of reworking since it was produced by explosive activity.

As with the tuff crystals, the clasts also contain unaltered plagiocase of labradorite-bytownite type, despite being sourced from the overlying lava which contains phenocrysts of identical shape but composed of albite.

This deposit is clearly the product of the explosive breakdown of the overlying porphyritic lava. When the lava reached the sea it disintegrated into breccia and pillows, but at the same time its surface chilled to yield abundant glassy basaltic fragments. These were then transported into deeper water where they formed a platform over which the lavas could prograde see (Figure 25.11), (Figure 25.13). The sediment was probably laid down sometimes in a series of mass flows and sometimes as normal tractive currents. The former mechanism is evident in the abundant, poorly sorted deposits some of which have boulders strewn through the tuff; the latter is seen in the poorly developed large-scaled cross strata. By comparison with present day examples of hyalotuff deltaic deposits, sediment pro gradations were probably very rapid when lavas entered the sea. Pauses in sedimentation took place between periods of lava activity and are marked by the beds of clast supported (well sorted) conglomerates on the foreshore.

The breccia beds (Figure 25.10) resemble the deposits already described, but differ in that they contain only clasts of dark aphyric lavas. These have had a provenance in the dark lavas which interfinger with the porphyritic ones at Locality 3, where evidence also suggests that both lavas were extruded at the same time. A possible explanation of this interstratification is given in (Figure 25.11).

Locality 6. Graded hyalotuffs (Figure 25.10)

Graded and ungradedbeds of tuffs, some containing large clasts of vesicular lava, can be traced north into the coarse grained deposits already described. These are considered to be the deeper water, distal equivalents of the breccias and conglomerates. They sometimes have breccia bands amongst them, suggesting periods when coarse sediments by-passed the delta into deeper water, maybe as coarse-grained grain-flows. Some of the large clasts are thought to have been fragments of pumice which floated out into deeper water where they became water-logged and dropped into the regions where finer sediment was accumulating. The graded beds are probably turbidites generated in the delta region by slumping during periods of lava extrusion and rapid development of tuff.

These finer tuffs are in contact with underlying prophyritic lavas which form the floor to this sequence and make up the higher part of Pinbain Hill to the NE where they are thicker and comprise more massive flows with little interbedded tuff.

Locality 7. Red conglomerates (only at low tide; (Figure 25.10))

A thin, well stratified and sometimes cross stratified conglomerate containing clasts of basalt and spilite occurs at the very top of this sedimentary sequence. The clasts are often well rounded and are either dark grey or red in colour. The red basalt clasts have almost certainly been derived from the tops of lava flows (as seen at Locality 2), having been oxidised in subareal conditions. The matrix of this conglomerate is volcanic sand with a calcite cement: the glassy tuffs which characterize most of the other rudaceous rocks are conspicuously absent.

Seemingly, the conglomerate formed during a period when lava extrusion had ceased in this particular locality and the sea was transgressing over an inactive and subsiding stack of lavas, which were being eroded at the sea margin. Clasts derived from these lavas were rounded and assembled in fairly high energy conditions, probably at the shoreline. Within this and other volcanic blocks, similar thin conglomerates with well rounded clasts can often be seen to truncate the underlying lava sequence: they probably formed on marine erosion platforms cut into lavas which were slightly tilted before or during the marine transgression. At this locality, the porphyritic lava flow appears to be eroded by such a surface beneath the conglomerate and at locality 2 the conglomerate rests on the dark lava.

Locality 8. Sediments on top of the lavas (Figure 25.12)

Unless the tide is very low it is easier to return to the car park and then descend the cliff again to examine the upper part of this sequence exposed to the north of the headland. Although the lower part of the porphyritic lava sequence is massive, it is capped by a unit where large pillows of lava are enclosed in tuff. The top of the lava can be traced along the headland (Figure 25.10) and in the bays to the north (Figure 25.12). They are red in colour.

The presence of massive lava, pillowed lava and tuff can be explained by the rates of lava extrusion (see (Figure 25.13)). If the rate of flow is slow, then the lava will not advance seaward beyond the intertidal zone as its outer skin will be continously converted into glass which then cools and fragments into tuff and breccia. With increasing rates of extrusion, the lava will advance beyond the intertidal zone and when the outer skin of glass forms, zones of weakness in this carapace will inflate by the pressure of the lava, much like balloons, to form pillows. If the rate of extrusion is slow enough, then the whole of the lava may be converted to pillows or pillow-like feeder tubes (Figure 29.13) B. However, if the rate of lava extrusion is quite high then only the outer margin of the complete flow becomes pillowed: the interior remains massive (Figure 25.13) C.

It is thought therefore that the massive flows at Pinbain are the product of very rapid extrusion of lava, and this outcrop at the top of the lava sequence is a record of the chilling on the outer parts of the lava flow. Above these pillows is a thick sequence of hyalotuffs and conglomerates and breccias, with mixed clasts of porhyritic and aphyric lavas. This abundance of hyalotuff, as with that below the lavas, is the produced at times of low rates of lava extrusion. The fact that many of the clasts are complete pillows, suggests that the lavas occasionally advanced beyond the strand line into the subtidal zone to generate pillows which then detached and rolled out in front of the migrating lava flow.

Locality 9. Lava tubes (Figure 25.12)

Thin finger-like units of aphyric lava occur within the hyalotuff. The origin of these is not readily apparent, but they are possibly the solidified tubes of lava which have advanced rapidly ahead of the migrating lava front and out over the platform or plinth of hyalotuff. They have mamilliferous outer chilled surfaces which may represent the incipient development of pillow buds on the skin of the lava. These features are thought to be similar to the lava tubes described from hyaloclas tic deltas seen forming at the present-day.

Locality 10. Contact with the overlying porphyritic lavas (Figure 25.12)

Porphyritic lavas with red and white phenocrysts are seen to rest on the hyaloclastic deposits. These lavas are massive with some zones of pillows. They represent rapid lava extrusions, probably once again at the strand-line.

Origin of the Slockenray sequence

The Slockenray sequence is considered to be the product of migration of a hyaloclastic or hyalotuff delta (Figure 25.11). There are two lava types present on the headland at Slockenray: a dark aphyric lava and a distinctive porphyritic lava. They were both extruded at the same time. This simultaneous extrusion is confirmed by the presence of dark as well as porphyritic lava clasts in the breccias and conglomerates of the hyaloclastic deltas, suggesting that the two types of lavas were hot and being broken up at the shoreline to produce their distinctive debris. In this way two cones of tuff from distinctive lava types overlapped each other and produced the interstratified sequence as seen at Slockenray.

The sequence at Slockenray is therefore significant in that it demonstrates that part, at least, of the Pinbain lava pile formed in intertidal conditions. However the whole of the Pinbain sequence is made up of lavas and breccia-conglomerates. These have been seen at Localities 1–9 and continue down the sequence to the Pinbain Fault, where they can be seen on the north outcrops of (Figure 25.15). Well over 50% of the Pinbain sequence comprises volcanogenic sediment, and throughout the sequence there are clasts which are well rounded. This implies that the complete thickness of about 1.5 km was deposited in fairly shallow water, and therefore that accumulation kept pace with subsidence.

Not only at Slockenray, but also elsewhere in the Pinbain sequence, there is evidence for advances and retreat of the lava. Conglomerate with the well rounded clasts (as seen at Localities 2 and 7 for example) probably represents an initial transgression on a subsiding block of lava (Figure 25.14). With increasing water depth the conglomerate is replaced by tuff and then by cherts, but when the lavas begin to advance again the cherts are replaced by tuffs and then by breccias and then by lavas (Figure 25.14). The cherts and tuffs associated with the deepening of the trough can be seen at the base of the sequence near the Pinbain Fault.

Locality 11. [NX 137 916]: Pinbain Fault and associated features ((Figure 25.15) a, b, c, d, e,)

The section begins at the north end of the beach to the north of Pinbain Burn (Figure 25.15). Lavas and associated sediments of the Pinbain Block strike NE-SW but are truncated at the base by the Pinbain Fault which, at the coast near Pinbain Burn is an almost E–W, 30–60 m wide fault zone. This zone contains sheared serpentinite, spilite, gabbro and a variety of other rocks. To the south of this shear belt lie the olistostromes and mass flow deposits of Pinbain.

(a). The sequence to the immediate north of the Pinbain Fault (a on (Figure 25.15)) contains tuffs, Ethic arenites, cherts and black shales which are overlain by breccias and agglomerates with accretionary lapilli. The cherts sometimes contain thin light-coloured laminae of feldspar grains; these are crystal tuffs and are probably of air-fall origin. The cherts are interstratified with thin black mudstones and shales from which, on the roadside above this exposure, Rushton et al (1986) have recovered fragments of trilobites, brachiopods and graptolites suggesting a Lower Arenig age. They are interstratifed with graded and ungraded beds, about 10–20 cm thick, which are probably the result of turbidity currents and grain-flows.

In terms of the fluctuating coastline shown in (Figure 25.14), these beds are thought to represent the deeper water or more tranquil part of the basin. The overlying breccias are recording the seaward advance of the lavas.

(b). Exposures of a sheared gabbro-breccia faulted against a sheared serpentinite breccia which occupies most of the sandy foreshore. This breccia zone marks the position of the Pinbain Fault.

(c, d, e). The olistostromes (isolated clasts in a fine grained matrix) and breccia-conglomerates of Pinbain are exposed in the raised-beach cliff sections to the east side of the road, and are seen to interfinger with black mudstones along the middle and upper foreshore, but are almost totally replaced by black mudstones on the lower foreshore (c, d, e of (Figure 25.15)) and in the sub-tidal outcrops. This whole outcrop is therefore at the interfingering boundary between breccia-conglomerate/olistostromes and the black mudstones.

The boundaries between the fingers of breccia-conglomerate and the mudstones (d) of (Figure 25.15)) are nearly always sheared: the former may have dark zones adjacent to the black mudstones, where mud from the mudstone unit has invaded their fabric. In some instances flames of dark mudstone have penetrated into the brecciaconglomerate, thus adding further evidence for the view that the mudstone was not dewatered when the breccia-conglomerate was deposited.

A prominent outcrop at low tide at (c), (Figure 25.15), comprises folded black shales made up of an early sequence of small, pyrite-bearing folds which are refolded by a later large-scale fold. The folding took place when the shale was ductile and it seems probable that both these phases of folding took place when the sediment was still plastic. In the outcrops surrounding (c), the black shales are often well exposed showing the sometimes quite intensive ductile shearing. The nature of the shearing is particularly well seen at (e) (Figure 25.15), where siliceous beds are interstratified with cherty shale beds and light grey, silicic tuffs.

The following important points can be drawn from an examination of this outcrop:

1. The shales are made up of black and light grey tuffs which are sheared into each other, producing a feather-like contact between the two lithologies.
2. The black shales are often tightly folded into cm scale folds, and the light grey tuffs are sometimes sheared up the cleavages, resembling flame structures.
3. Black cherry shales have either resisted deformation or have been brecciated, with mud infilling the breccia matrix.
4. Pyrite growth in the shales pre-dates the deformation in some instances, but appears to have overgrown the deformation in others.

It is concluded that the deformation in the shale sequence took place when it was unconsolidated. Clasts in the breccia-conglomerate are sometimes quite large; some of the blocks of pillow lava exceed 20 m in diameter (Locality 12; a of (Figure 25.16)), forming boulders. There are clasts up to boulder size of the following types: pre-existing conglomerate, carbonates, turbidites comprising dark volcanic rich lithic arenites, various sized clasts of granite (including trondhjemite), sheared rocks rich in epidote, serpentinite, amphibolites. Some of the fine conglomerates contain clasts of blue-schist.

Locality 12. [NX 1372 9145]: Olistostromes and their contact with the serpentinite (Figure 25.16)

The olistostromes are exposed on the rock platform at (b) (Figure 25.16), where boulders are surrounded by sheared shales. Some of the shale beds are siliceous, fairly well stratified and contain microscopic radiolaria: they are normally tougher than the unsilicified shales and have consequently resisted the otherwise pervasive shearing.

The following points are thought to be significant in an interpretation of the outcrops of sedimentary rocks described at Localities 11 & 12:

1. Large clasts were displaced into quiet waters where the shales accumulated; this in turn implies that steep slopes developed on the side of the basin in which muds formed, in order to displace repeatedly the blocks into it.
2. Considerable shearing took place during and immediately after the deposition of the sediments, shears which can now be seen in the shale-chert sequences. The shales probably acted as detachment horizons for major shears created on the basin margins: or at least during episodes of tectonic activity within the basin.
3. The abundant tuff bands suggest that the shale basin was relatively proximal to an area of volcanic activity, but not so close that pyroclastic flows entered the basin. The considerable exposures of pillow lava may have been intruded into the wet muds of the basin but could also have been detached from cliff-like faces in older lava flows and then slid into the basin of deposition. Those clasts of pillow lava which are rounded to varying degrees, probably acquired this rounding at a shoreline and were then displaced into deeper water.
4. The source of the conglomerates comprised a wide variety of ophiolitic rocks which were formed in quite disparate environments; fragments of blue-schist from high pressure, low temperature conditions: sepentinites from the mantle: amphibolites from fairly high temperature metamorphic regimes: and trondhjemites, gabbros and dolerites from quite high in the oceanic lithosphere. Some rocks were already deformed by the time they entered the conglomerate. All this implies that a great deal of the ophiolite, including rocks which have a provenance deep within the lithosphere and the mantle such as blue-schists, amphibolite and serpentinite, were exposed at the time, and had all been mixed before arriving at the basin of sedimentation.

Since black shale sedimentation accompanied tuff accumulation, it is reasonable to assume that the basin opened during extension accompanied by volcanicity and that the conglomerates are associated with normal rather than compressional faults. The faults probably detached within the black shale and chert horizons, causing the intense ductile-style deformation there. The acidic-intermediate tuffs together with the pillow lavas, indicate accumulation in a submarine to subareal volcanic regime and the presence of ophiolitic debris and the absence of terrigenous detritus suggest a volcanic complex founded on pre-existing oceanic crust.

The whole exposure at Pinbain is terminated in the south by a prominent dyke (Figure 25.16), possibly of Tertiary age, south of which there is a sandy beach with a few exposures of breccias and serpentinite, seen only at low-tide. The breccias which outcrop for a few metres to the immediate south of the dyke (Figure 25.16) are associated with sheared black shales and sheared blocks of ophiolitic rock and at this point probably constitute a mélange. To the south of the breccia there is a wide outcrop of normal serpentinite, which makes up most of the low ground extending to Lendalfoot.

There is clearly major displacement between the serpentinite and the breccia-olistostrome sequence, as each formed at totally different crustal levels. The sediments formed on the surface; the serpentinite has been brought up from depths exceeding 10 km. The timing of the juxtaposing of these two lithological units is uncertain: the serpentinite may have been uplifted before the deposition of the sedimentary units so providing a source for the conglomerates.

These sediments are not typical of layer1, ocean basin sediments in the ocean crust sequence. They are clearly deposited in basins where there was a copious supply of coarse clas tic sediment. For this reason they are not thought to be associated with an ocean ridge, although it is possible that they are part of a fracture-zone sequence; also they are not likely to have formed as part of a hot-spot, as the composition of the conglomerates indicates a source which included metamorphic rocks and serpentinite which are not found in hot-spot oceanic islands; but in all details they resemble the sediments produced during arc-rifting and marginal basin formation (see Introduction).

Locality 13. Bonney's Dyke [NX 1347 9113]: Gabbro pegmatite (Figure 25.17)

Bonney's Dyke was named by Balsillie (1932) after the famous petrologist T.G. Bonney. The feature is not a dyke but a steeply dipping, sill-like sheet intruded into serpentinite, comprising a coarse gabbro pegmatite of altered diallage and feldspar, most of the latter being replaced by white prehnite and pectolite. The following points should be noted: the sheet thins towards the west: it is sinistrally shifted by a number of faults and as such provides an excellent strain marker within the serpentinite: it has a hydrogrossular northern margin and contains rafts of serpentinite. The serpentinite to the immediate south of the sheet contains large crystals of diallage: to the immediate north is a norite with enstatite crystals enclosed in plagioclase; the norite is only visible at low tide.

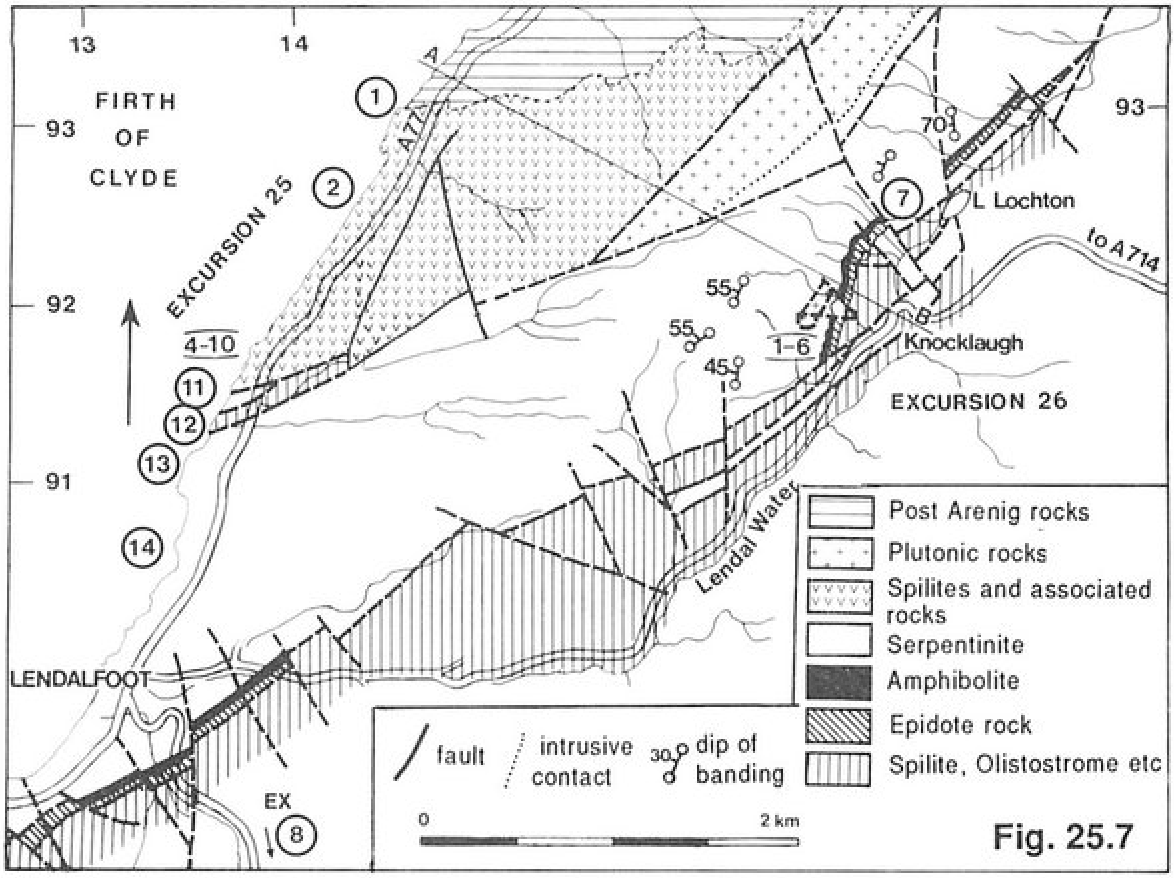
Locality 14. Albite diabase sheets and rodingite

The serpentinite to the south of Bonney's Dyke is intruded by many diabase sheets which can be examined at most tides along the shoreline to Lendalfoot (see Balsillie 1932). These sheets are mostly altered but detailed work on their chemistry and mineralogy has now demonstrated that there are at least two types of sheets present (Holub et al 1984); group 1 have mainly amphibole and plagioclase and a provenance in a depleted mantle source; group 2 have clinopyroxene and plagioclase present, and a source in a less depleted mantle. Group 1 sheets were intruded during elevated temperature, possibly when the serpentinite-peridotite was at greenschist facies; group 2 were intruded into colder rocks. Both have been, in places, altered to a white massive hydrogarnet rock (rodingite) and both therefore were intruded during the period of serpentinization.

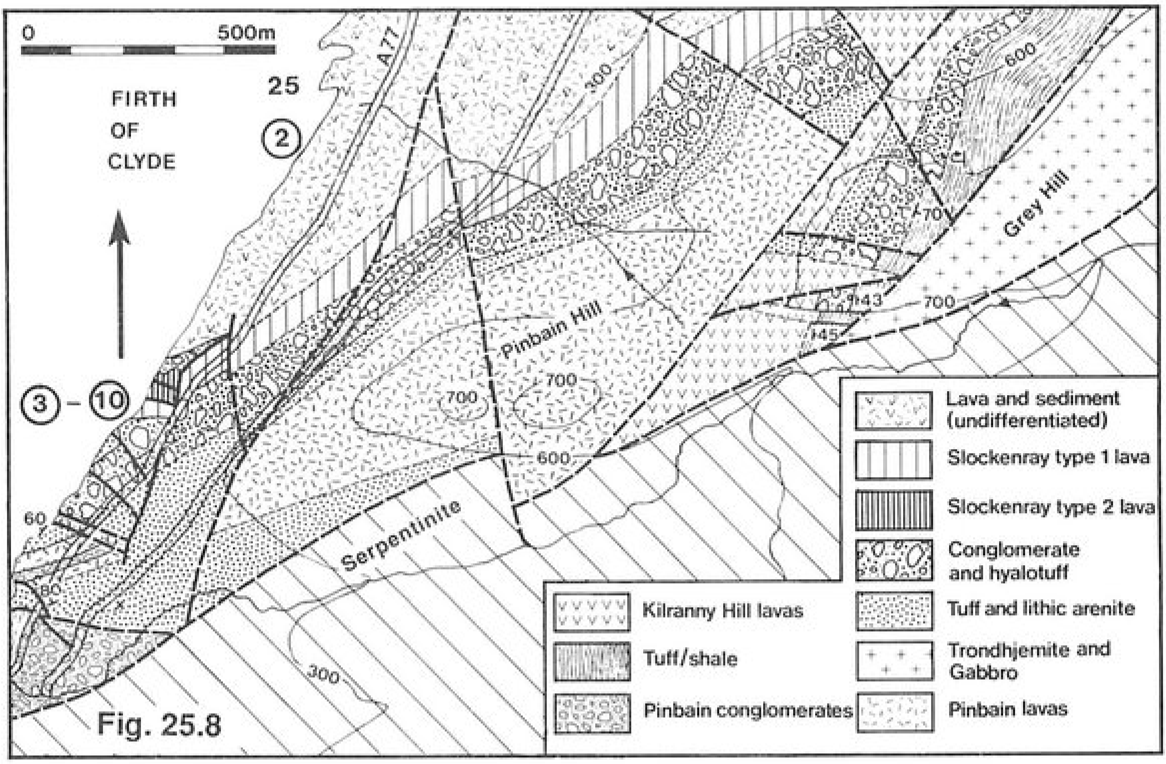
Many of these intrusions are pod-like. It is often possible to trace chilled margins over entire exposed surfaces. The sheets have large xenoliths of serpentinite within them and may have a very irregular, flame-like contact with the serpentinite which encloses them.

References

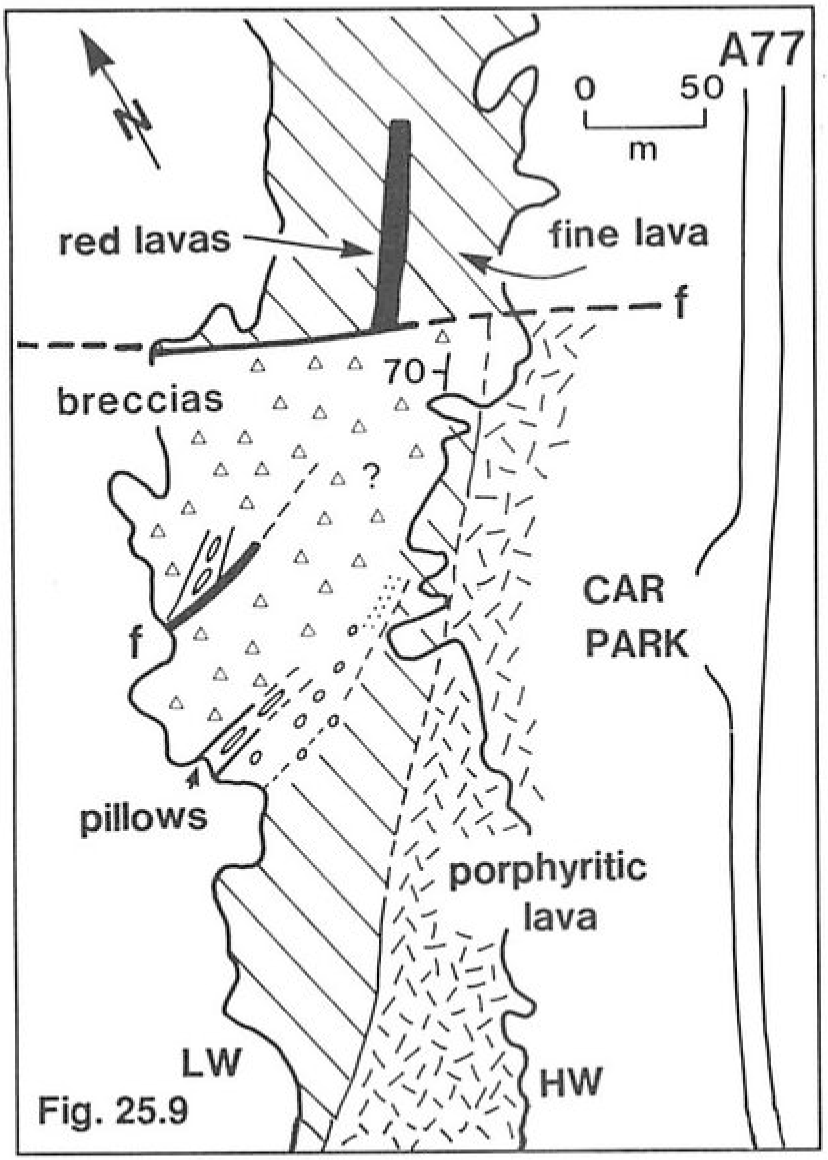
[References for excursions 25–31]



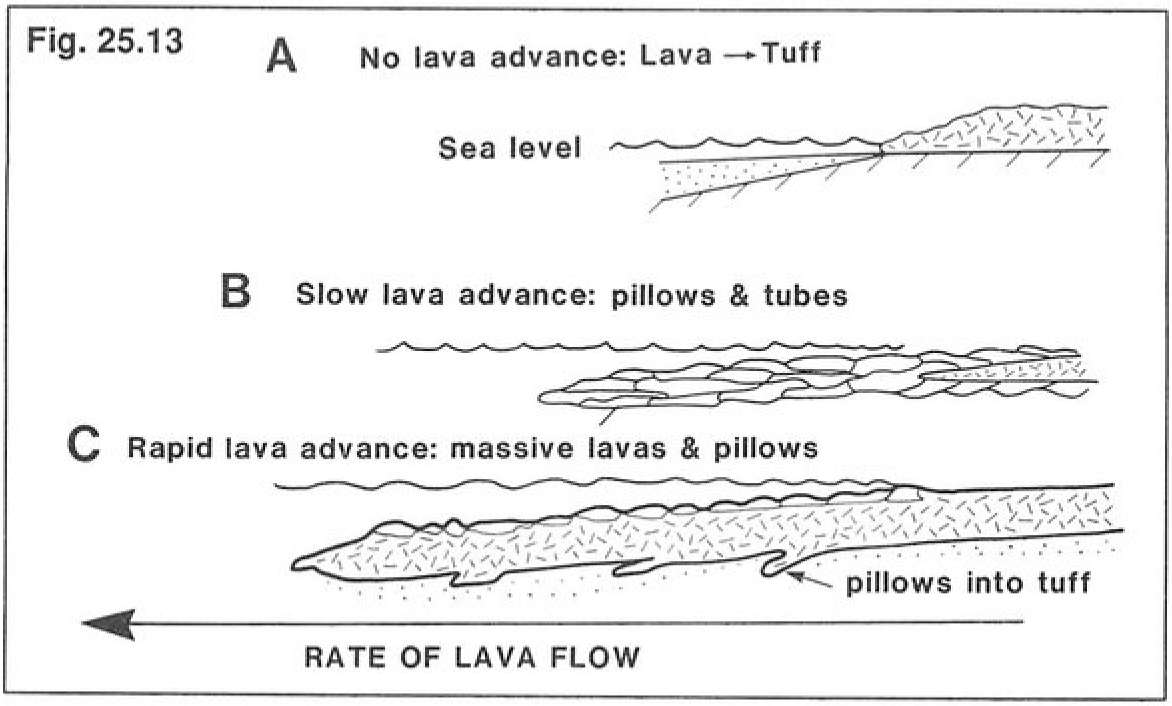
(Figure 25.7) Simplified map of the northern part of the Ballantrae Complex, with positions of localities mentioned in Excursions 25 and 26.



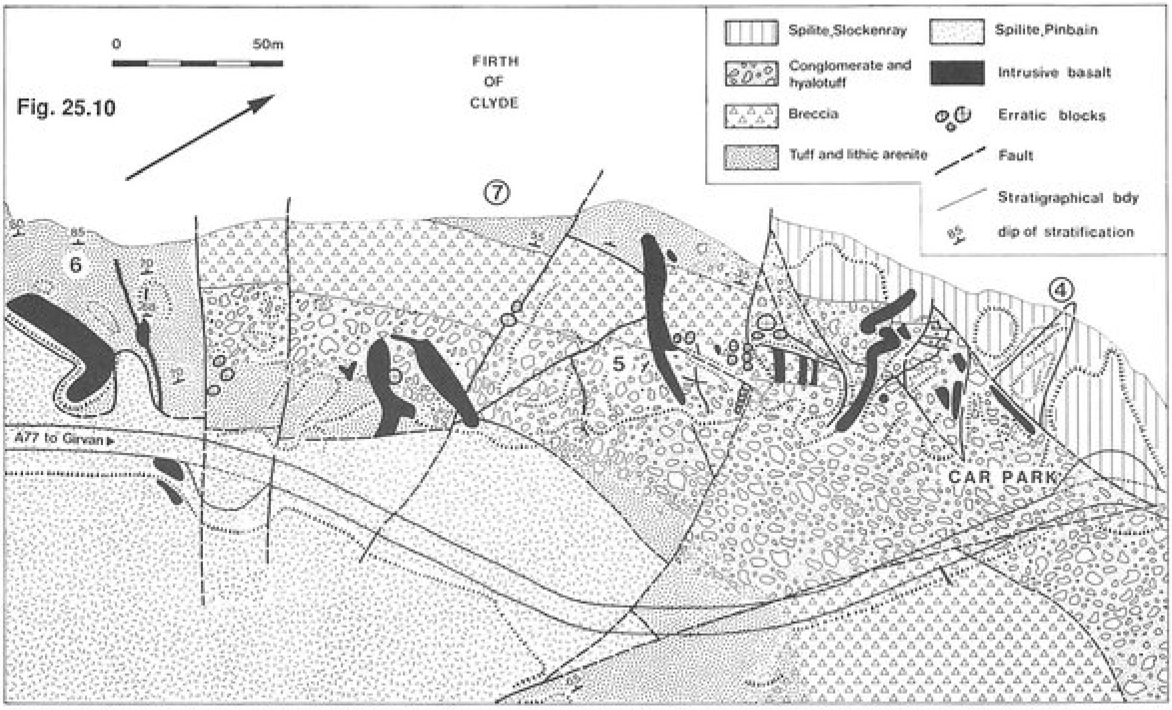
(Figure 25.8) Map of the Pinbain Block showng lateral extent of some of the lavas.



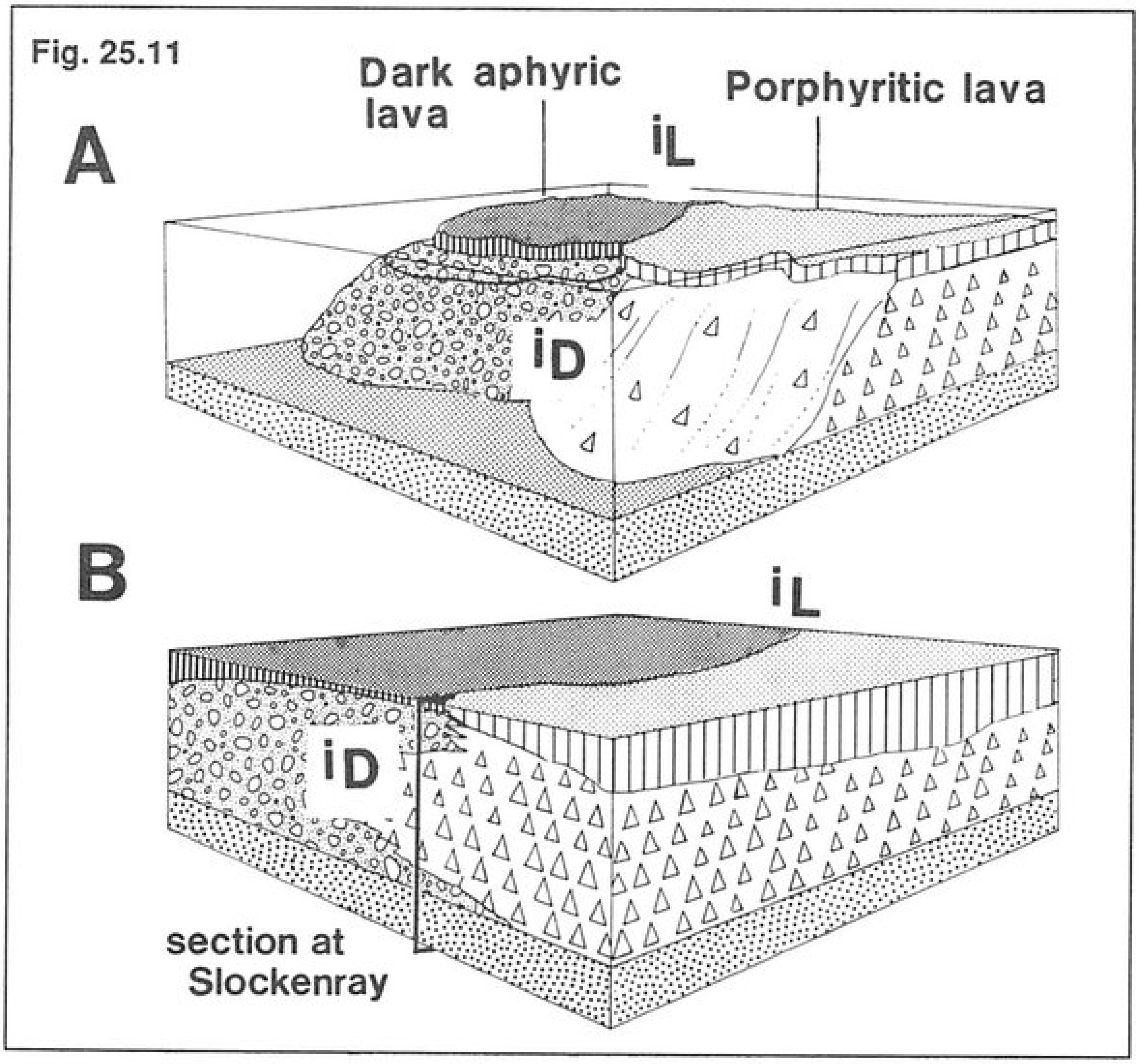
(Figure 25.9) Sketch map to illustrate the geology at Locality 2, Pinbain Block. LW and HW refer to low and high water marks.



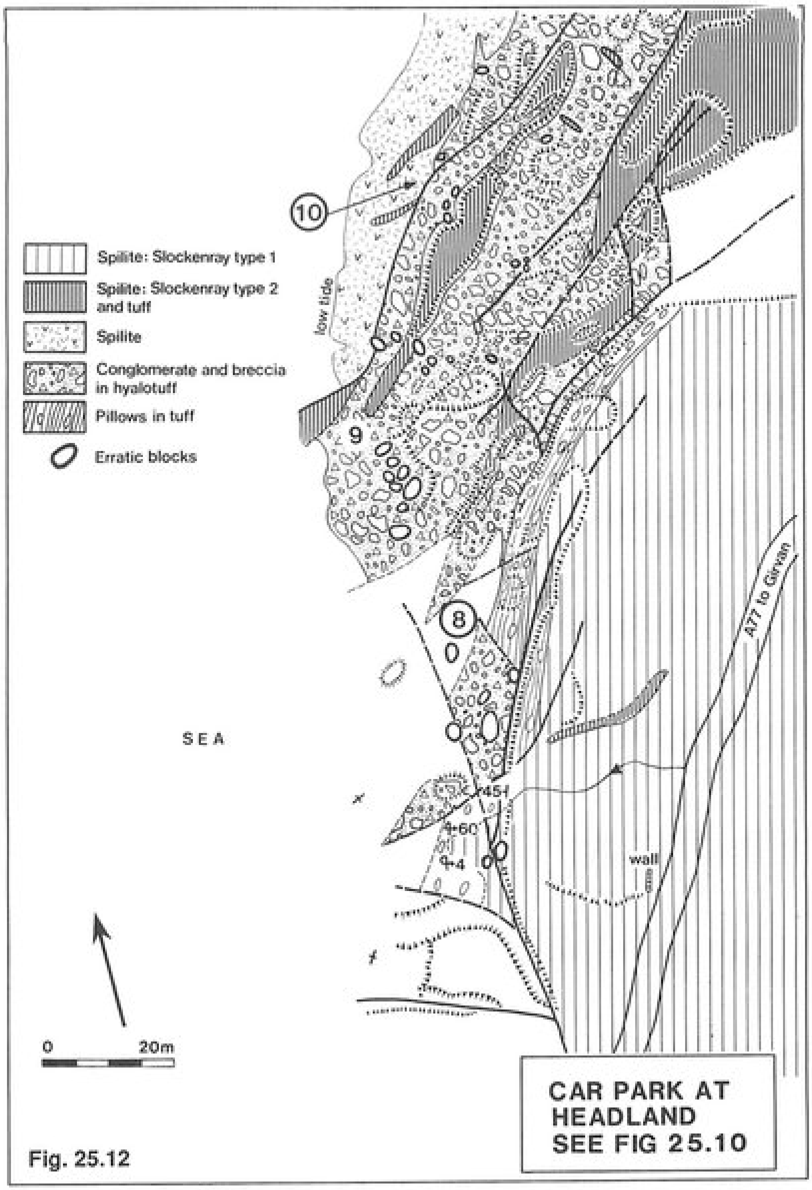
(Figure 25.13) Explanation of the development of various lava structures and tuffs in lavas which enter the sea from the land. A, the lava front is moving slowly and as it enters the sea where it is rapidly chilled, all of it is converted to tuff at the shoreline. Waves and currents move the tuff offshore. If the tuffs are generated in sufficient abundance then the lavas will flow over them to build up a hyalotuff delta, as seen at Slockenray. B, Lava is moving sufficiently rapidly to enter into the sea, but much of its outer skin is chilled by contact with the sea water. The chilled skin is inflated by magma which is under pressure and many pillows are produced. C, the lava advance is rapid, so that the outer skin chills and forms pillows, either by contact with water at its top surface or by tuff at the base. However the rapidly moving interior is insulated by this pillow growth and cools to form a massive lava which cannot be chilled by contact with the sea water. The porphyritic lava at Slockenray is of this type: it is pillowed at the top and sometimes at the base, but has a thick, massive interior.



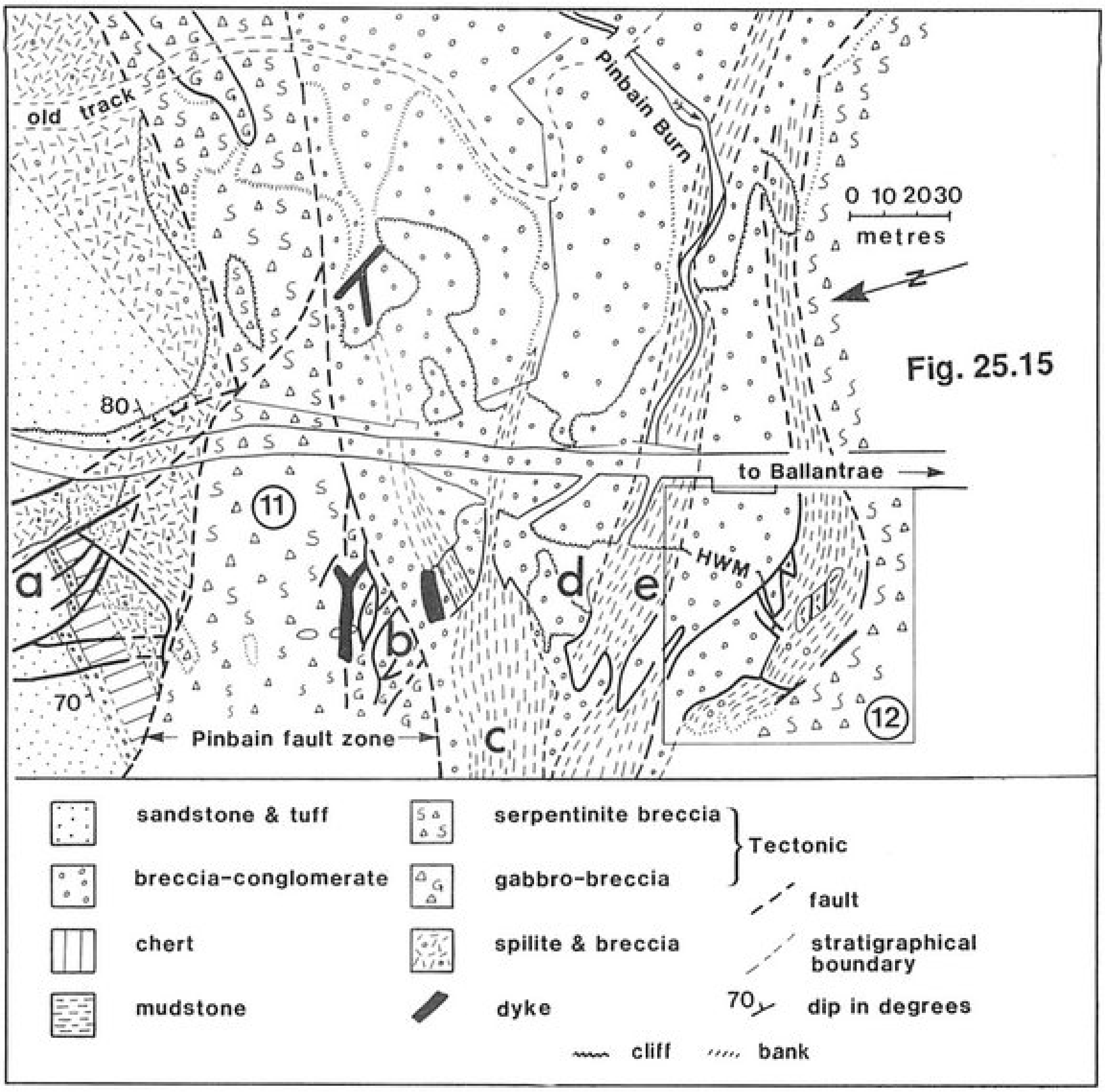
(Figure 25.10) Map of the southern end of Slockenray. The north margin of map begins at the headland which divides this map from (Figure 25.12), and the two lava types–porphyritic and dark aphyric (not subdivided in the map) They both belong to the Slockenray spilite of the caption.



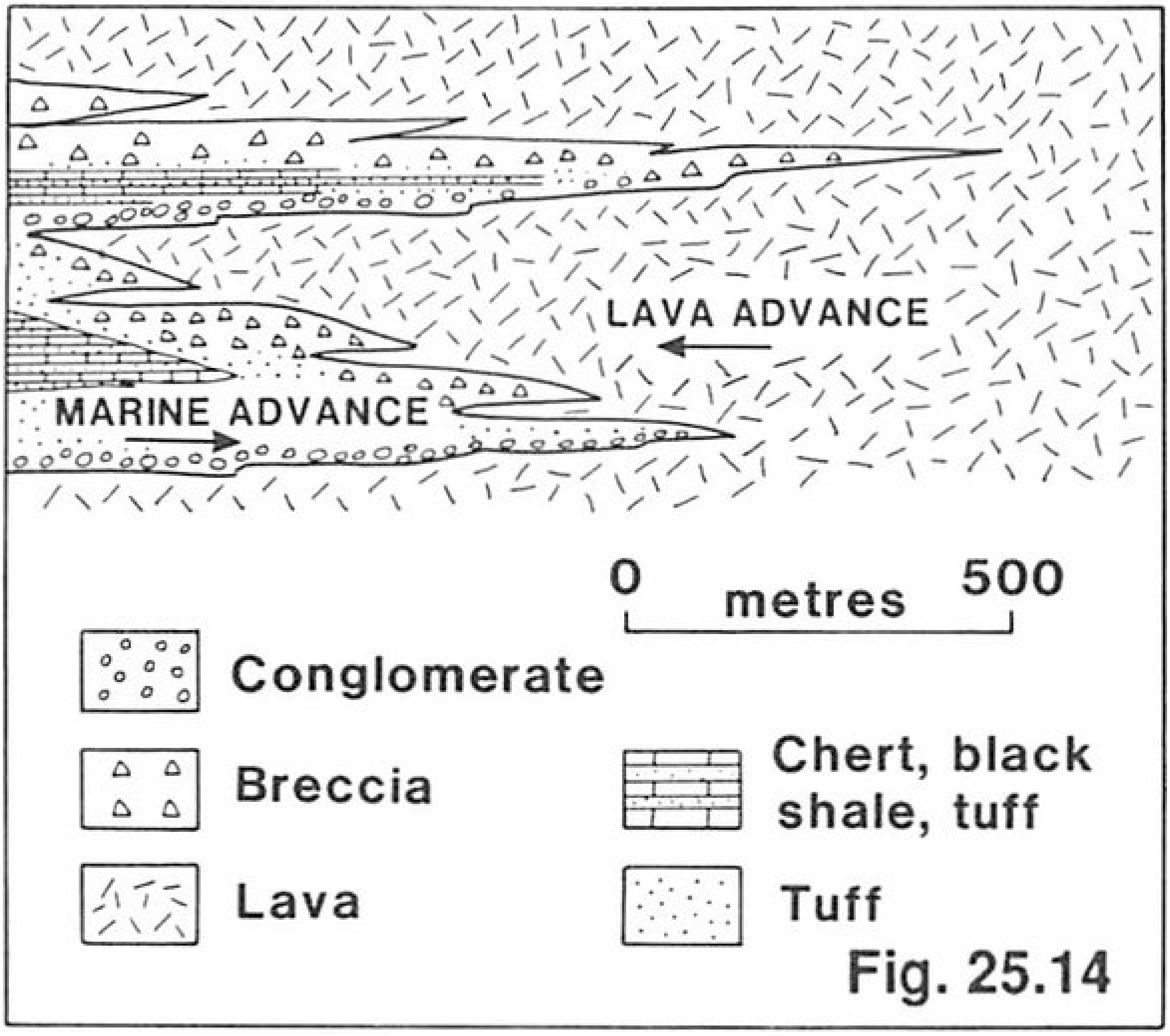
(Figure 25.11) Explanation of the Slockenray sequence. A. Two lava flows, one porphyritic, the other dark aphyric, are both simultaneously extruded and flow together towards the coastline where they both build out delta cones adjacent to each other, each cone being sourced by their individual lava types. When one lava type becomes dominant the boundary between them i_L) changes position to extend the area of the dominant flow. At the same time the delta produced by the dominant lava expands at the expense of the delta produced by the less dominant lava and the boundary between them (i_D) is affected. B, shows the location of the Slockenray section and the interfingering of the hyalotuff delta deposits which may have been caused by the growth of one delta at the expense of the other.



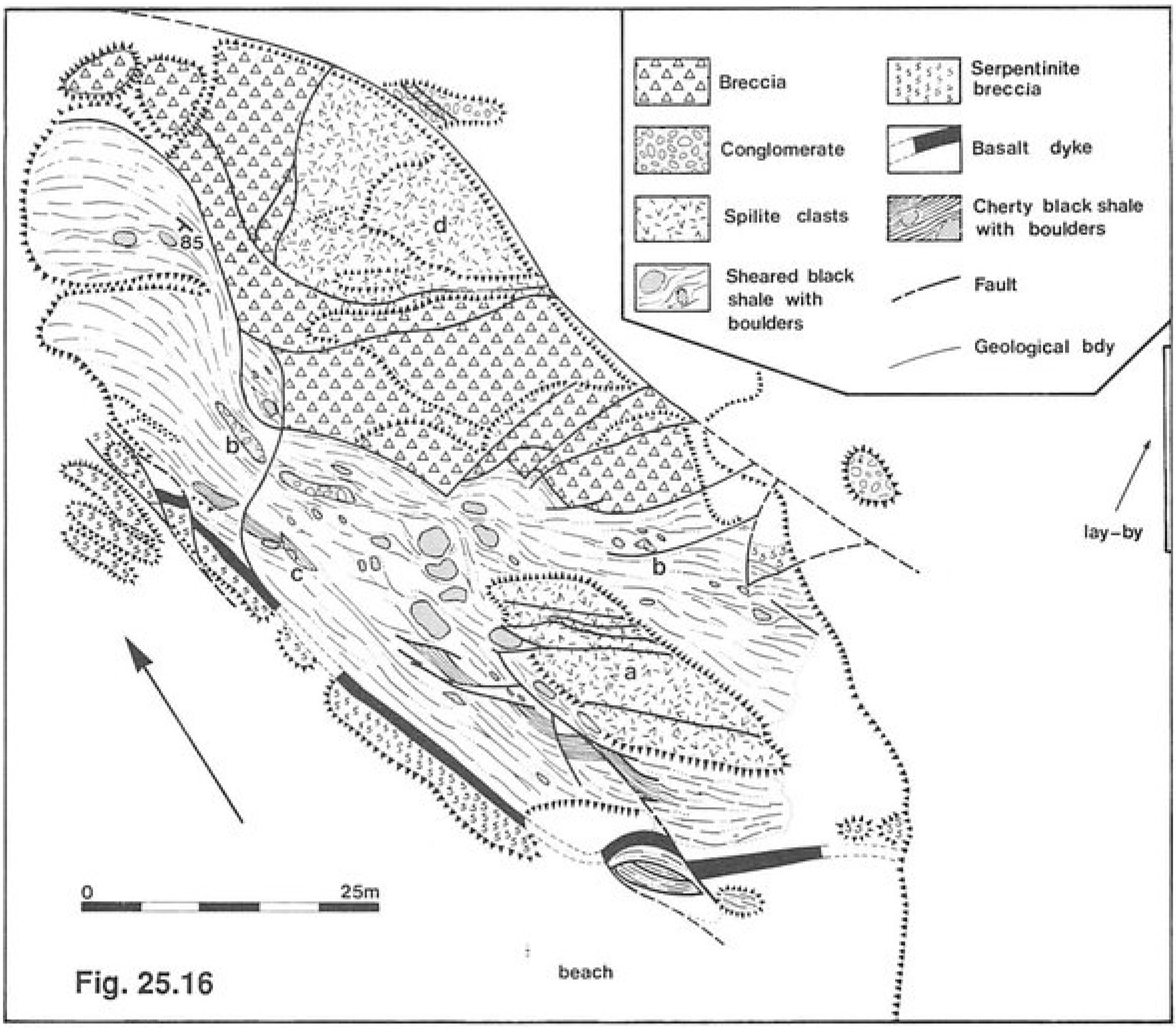
(Figure 25.12) Map of the north of Slockenray headland, showing the sequence above the lavas. The south margin of the map is north of the car park and a key point in locating the exposures with reference to the map is the small wall on the edge of the road as marked on the map. Lava type 1 refers to the porphyritic spilite; type 2 to the aphyric.



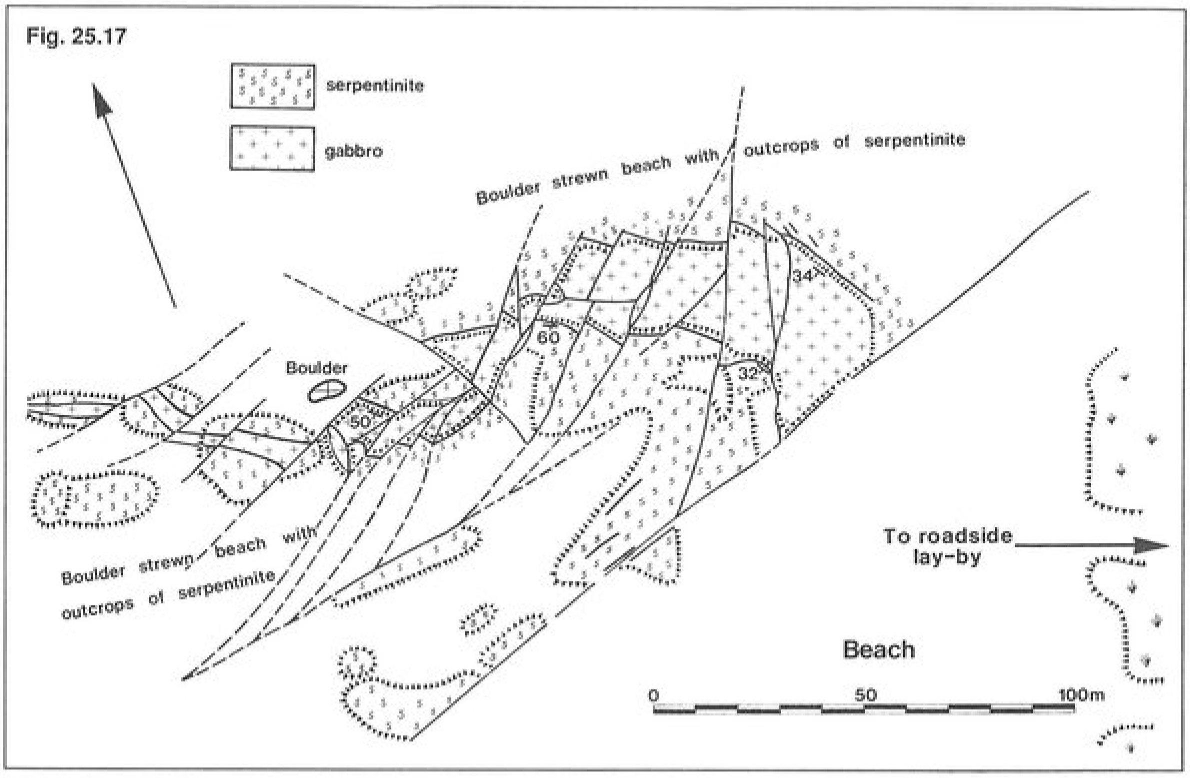
(Figure 25.15) Plane-table map of the region near Pinbain Burn. Letters a, b etc refer to localities discussed in the text; inset is the approximate position of (Figure 25.16).



(Figure 25.14) Explanation of the sediment-lava cycles in the Pinbain Block and elsewhere. When the rate of lava extrusion is rapid or the rate of sea-level change is slow, the lavas advance into the sea. Because of seawater-lava interactions, where the lavas break down by explosive or erosional activity, lavas are always associated with tongues of lava entirely enclosed in breccia. However further towards the source of the lavas there are fewer breccia deposits. Shales and cherts on the other hand accumulate in deeper water and associated with them are tuffs which were deposited there either by air-fall (from explosive activity), storm deposition or turbidites. When volcanic activity has ceased or is waning the sea transgresses over the lavas to yield well rounded conglomerates, sometimes with reddened clasts if the lavas have been subject to subaerial exposure. This association of lava and breccia is common in nearly all the major lava sequences at Ballantrae, and in this Pinbain section massive lavas characterize Pinbain Hill; interfingers of breccias and lavas are seen on the coastal section (Localities 1–10) Transgressive conglomerates are seen at Localities 2 and 7).



(Figure 25.16) Plane table map of the ground south of Pinbain Burn (see (Figure 25.15) for position).



(Figure 25.17) Plane-table map of Bonney's Dyke.