THE BALLANTRAE COMPLEX

B.J.Bluck

Introduction

The Ballantrae Complex in SW Ayrshire has attracted a good deal of attention since the last century, when Murchison, Geikie and Bonney discussed its problematical origin. It was clear to most of these and subsequent workers that the association of **serpentinite, chert** and pillow lavas was repeatedly found in major fracture zones, and for this reason they regarded the association as significant. However, the true importance of the rocks at Ballantrae became more apparent when work in Newfoundland and Cyprus demonstrated that rocks, similar to those found at Ballantrae, were fragments of oceanic crust, and that these oceanic crustal slices had been thrust onto the continents. As the knowledge of destructive and passive margins increased it became clear that the slices of oceanic crust which had been thrust onto continental margins were a signature of typical destructive margins: and so, with the Ballantrae complex in mind, Dewey (1969) recognised for the first time the Caledonides as a destructive margin. The whole rock assemblage at Ballantrae has become increasingly important, not only because it is oceanic crust, but also because its presence here raises a number of important questions, two of which are applicable generally to rocks of this oceanic type (ophiolites).

Firstly, in what kind of oceanic setting did these rocks form ? This entails the establishment of criteria by which various types of oceanic crust can be distinguished.

Secondly, how does oceanic crust appear on land areas, when it appears that most of it is being consumed at trenches?

The excursions which follow this introduction all have a bearing on answering both of these questions.

Origin of ocean crust

Ocean crust is now known to form at two main situations: ocean ridges and marginal basins. However, crust of oceanic type may also form in hot-spots (sea mounts) or in oceanic island arcs. Most of the ocean crust now produced on the earth's surface forms at mid-ocean ridges, and this crust has a characteristic structure (Fig. 25.3) which may largely depend upon the rate at which the crust is generated.

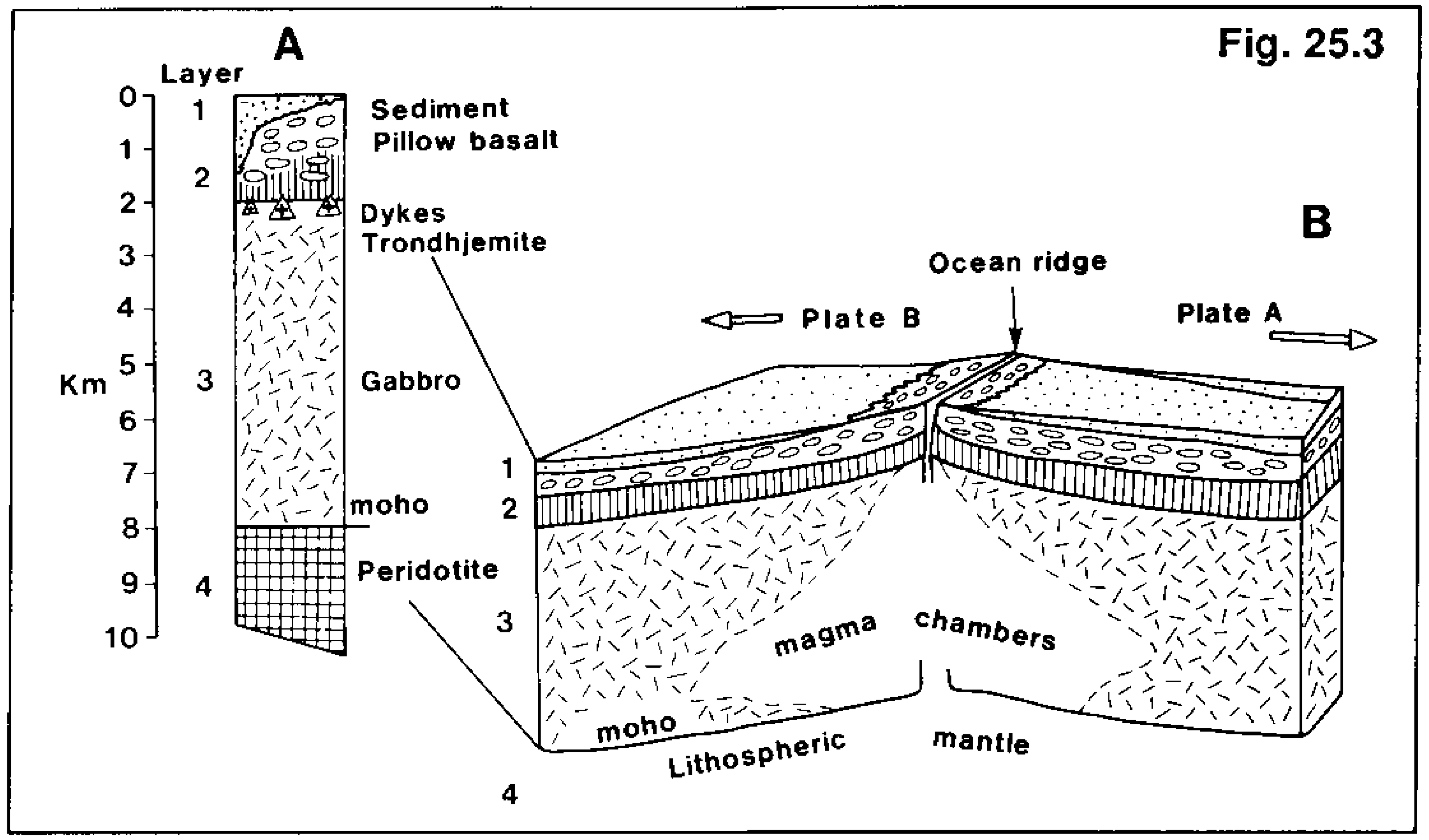


FIGURE 25.3. A. Section through typical oceanic crust. **B.** Diagram showing how oceanic crust is created in an instance of a rapidly growing plate. The peridotite of **A** belongs to the lithospheric mantle of **B**.

When this structure is compared with the rocks at Ballantrae it is clear that, with the possible exception of the sheeted dyke complex, all the rocks which are thought to be characteristic of ocean crust are present; for this reason alone it is fairly safe to assume that the Ballantrae rocks are of oceanic type (Fig. 25.4). But there are many types of oceanic crust to consider, and there has been much debate about which of these types of crust is represented by the Ballantrae Complex. The origin of the various types of crust is shown in Figure 25.5, together with their main characteristics. The most diagnostic characteristic of the various kinds of crust are seen in the layers 1 and 2: differences which might arise in the other layers of oceanic crust are not well known. Where there are faults crossing the hot ridges, deformation and metamorphism may occur whilst the oceanic crust is being formed.

Oceanic ridges are usually found in quite deep water and are characterised by fine grained sediments in layer 1: these are often produced by organisms such as **radiolarians** living within the water column and falling onto the plate surface after death. As there is little explosive activity at these depths this fine grained sediment is usually devoid of much tuff. Black shale-type deposits may comprise layer 1,

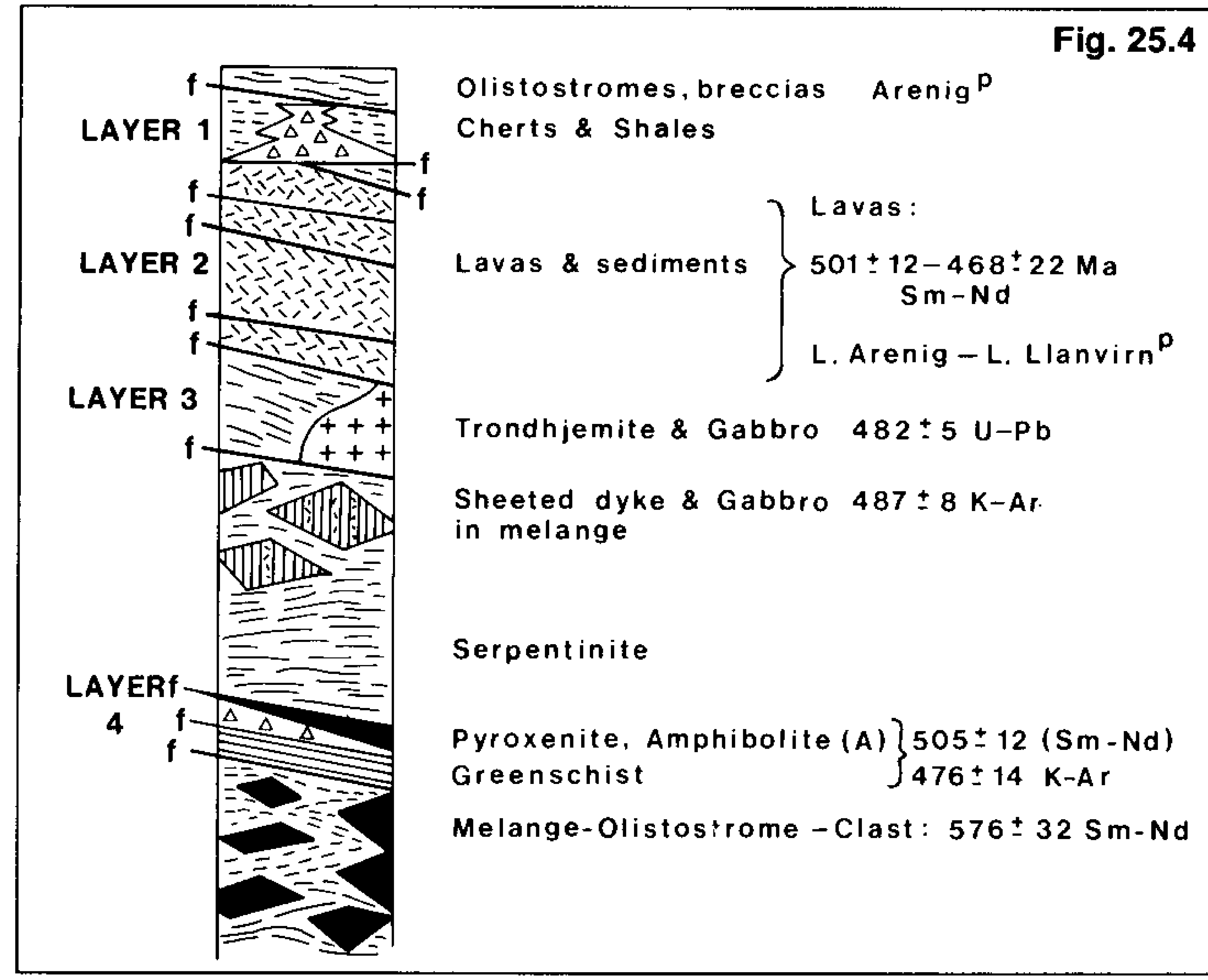


FIGURE 25.4. Compound section through the Ballantrae Complex showing some of the absolute age determinations (the various methods used are shown by the conventional symbols K-Ar etc) and fossil ages (p). On the left of the section the various elements of the complex are interpreted in terms of a conventional ophiolite.

where the ocean floor in near a source of terrestrial sediment, as for example, where ocean crust reaches a subduction zone.

Layer 2, usually comprises basalts with a minor amount of breccia and very little evidence of explosive activity during the formation of the lavas - the water being so deep that the water pressure is too high to permit much explosive release of gas. These lavas are often pillowed but do not extend for great distances since there are low slopes at most of the positions of extrusion on the ridge and the lavas tend to chill quickly. This results in mounds of pillows locally building over the points of extrusion.

With the development of hot-spots and seamounts on the ocean plate the nature and thickness of the lava pile is changed. In this instance the lava pile grows from deep water to shallow, so that

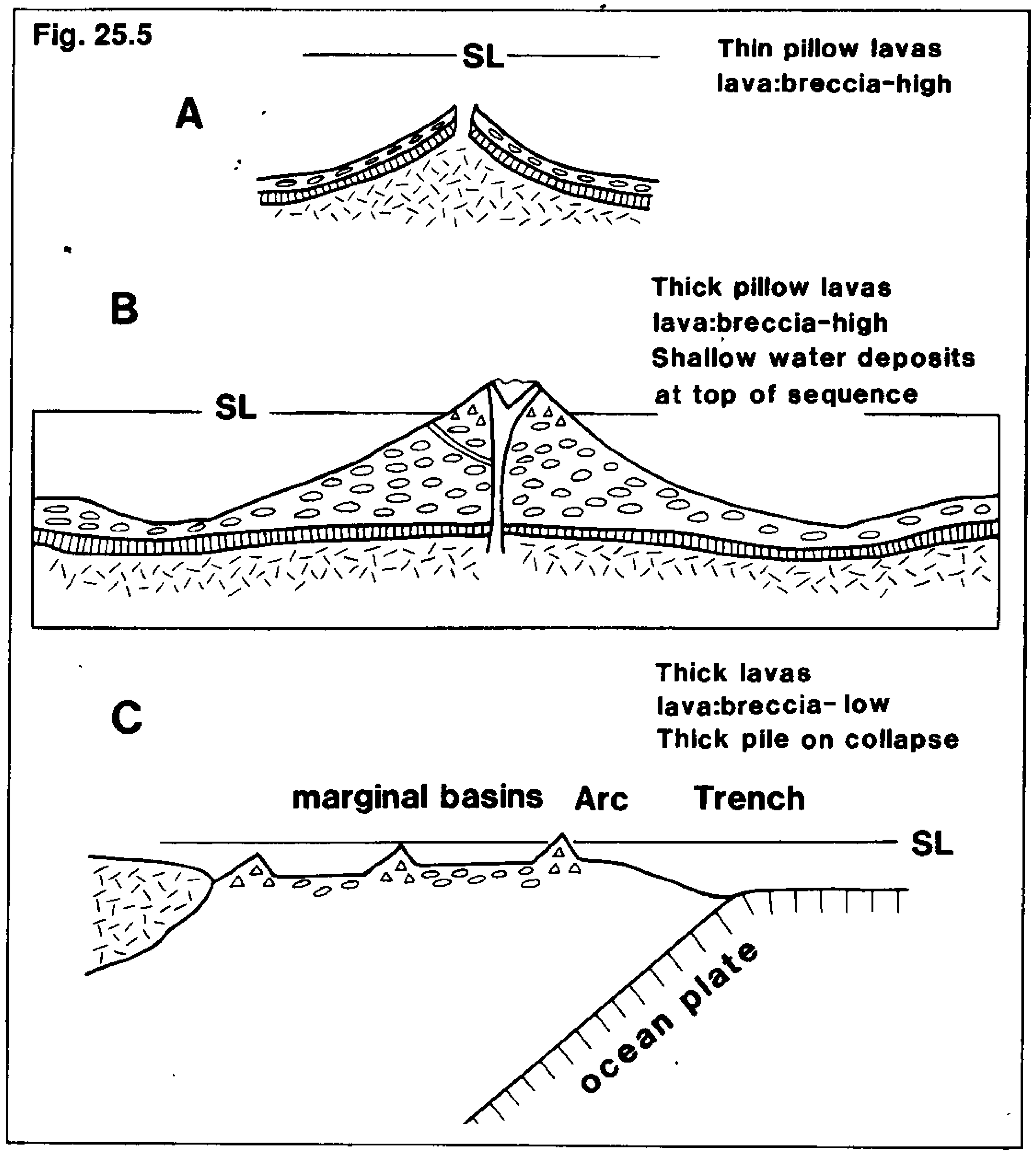


FIGURE 25.5. Diagram showing the various ways in which ophiolites form, together with some of the main characteristics which typifies each one. **A.** Formation at a spreading ridge; **B.** At an ocean seamount, **C.** At a marginal basin-arc. SL = sea level. The symbols are as for Figures 25.3 and 25.4.

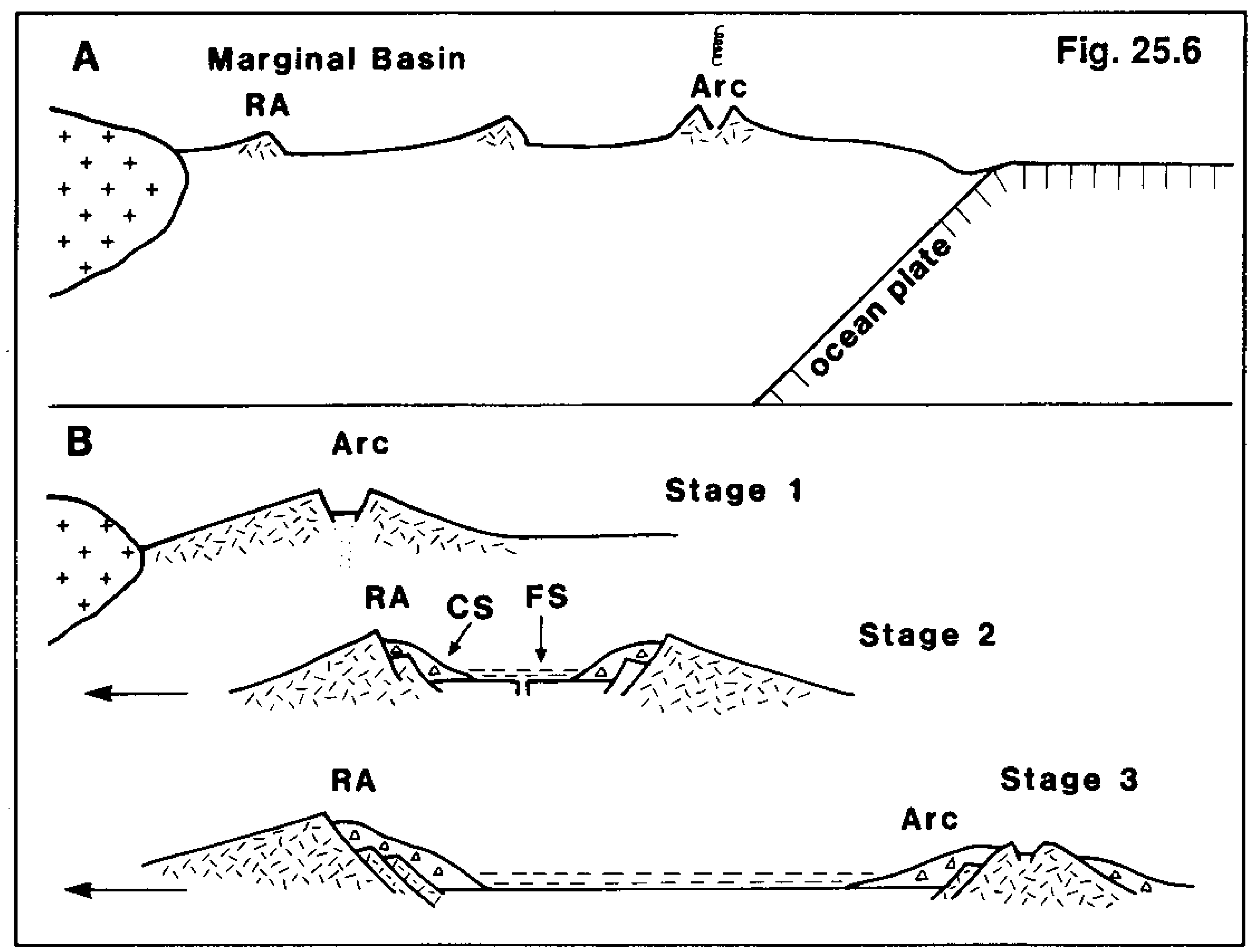
initially the pile is dominated by pillows which lack evidence of explosive activity, to be followed at the top of the pile by a relatively thin interval of shallow water and intertidal flows and finally the subaerial flows as the cone emerges to form an island (Fig.25.5B). As these lavas are built on ocean crust which was generated at a ridge, the age of the lavas will be much younger than the age of the ocean crust,

provided they formed at some distance from the ridge; and, conversely, the ages of the two will be similar if the hot spot is close to a ridge.

Marginal basins are far more complex, and whilst there are apparently many ways in which they are produced, a common way is to rift an arc to produce new oceanic crust within the rift zone (Fig. 25.6). The new oceanic crust may then grow in the usual way to create a new marginal basin terminated on the continent side by a remnant arc (RA, Fig. 25.6), and on the ocean side by the now rejuvenated active arc (Fig. 25.6). In this way there can be stretches of oceanic crust belonging to marginal basins which are divided by remnant half-arcs.

The sequences produced by this means are quite different from those produced elsewhere. Layer 1 is not now a sequence of fine grained sediments produced some distance away from source, but coarse detritus derived locally from the splitting arc (CS Fig.25.6B); however, as the marginal basin opens up so the sources become more

FIGURE 25.6. Origin and development of a marginal basin. A. Section through a developed marginal basin. B. Stages in the growth of a marginal basin. Stage 1, the splittting of an arc, Stage 2, the development of new ocean crust in the rifted arc, Stage 3, the development of a wide ocean basin. RA, remnant arc. CS = coarse sediment; FS = fine sediment.



distant and the sediment finer grained (FS Fig. 25.6B). During the splitting of the arc, the foundation upon which some of the sediment has already accumulated becomes unstable and subject to extensional faulting. There are now sharp boundaries between sources and basins. The volcanogenic sediment, which accumulates as an apron around the arc, is made up of angular and (intertidally) well rounded clasts, and is displaced into deeper water as mass flows. Sediments, some of which come from the shallow-water zones are displaced towards the basin axis by slump and mass flow action. Finer grained sediment may be organically produced (e.g.cherts) in association with wind blown ash generated by explosive activity on the arc.

The arc itself comprises a great thickness of volcanogenic sediment and, to a lesser extent, lava. If the arc has been allowed to develop for a long time it becomes mature and produces acidic lavas. If, however, it is continuously rejuvenated, as when it splits to form a new marginal basin, then the arc may remain immature and produce basic lavas only. In this arc regime the volcanogenic sediments range from being subaerial to shallow water intertidal to deeper water.

There are a few critical features of the Ballantrae complex which have some bearing on the type of ocean process which might have formed it. These are as follows:

1. Cherts and black shales occur in a number of associations at Ballantrae and in terms of oceanic layering can be ascribed to layer 1 (Fig.25.3). As discussed above, of special importance are:

(a). The presence or absence of coarse grained clastic sediments; deep ocean basins are dominated by fine sediments; parts of seamounts and all of island arcs are dominated by coarse sediment (Fig.25.5). The sediments of layer 1 can be seen at Bennane Head where they are found in association with boulder-bearing conglomerates and breccias.

(b). The presence of acidic or intermediate rocks, clasts or rock fragments associated with these fine grained sediments. Ocean ridges tend to be dominated by basic rocks only, whilst hot spots may have, in addition to basic, intermediate rocks present as well. Arcs may be largely basic when youthful, but mature to produce calc-alkaline and acidic volcanic rocks. Acidic rocks fragments are associated with the cherts at Bennane Head.

(c). Whether the sediments show any signs of tectonic activity as

might occur when an arc splits to form a new marginal basin i.e. aprons of mass flow deposits with much coarse sediment associated with slumped beds. These sediments can be seen at Locality 12 on Excursion 25 at Pinbain and at Locality 3 on Excursion 27 at Bennane Head. The stratigraphical associations at Pinbain are not clear as the exposure is fault bounded. Bennane Head is the critical exposure, for here is a sequence from lavas and conglomerates up into cherts, and that is the sequence one would expect where layer 2 (basalt layer of Fig.25.3) is overlain by sediments (layer 1 of Fig.25.3). So at Bennane Head it would appear that we have a clear example of layer 1 in its stratigraphical context.

2. The lava sequence is also quite critical for the identification of the origin of the ophiolite at Ballantrae. Lavas occur in at least three quite extensive blocks, the most northerly of which is the Pinbain block. The lavas and associated sediments of this block are terminated to the SE by a major fault (the Pinbain Fault seen on Excursion 25) and to the NE the unconformably overlying Girvan clastic sequence. South of Pinbain lies the Bennane Head block, which has a sequence of cherts and shale at its top (see above) and by a major fault at its base near Games Loup. The most southerly block is found in the Mains Hill-Knockdolian region: it is terminated to the south by cherts and black shales and to the north by a major fracture. A fourth block, the Aldons block is not well known. There is no contact between the lavas and the sheeted dykes and as already discussed, there is an upper contact with strata which are superficially similar to ocean layer 1 at Bennane Head.

For the lavas which can seen on Excursions 25 and 27, there are several critical lines of evidence which allow an evaluation of their origin:

1. The abundance of breccias and conglomerates characterizes shallow water volcanic processes.
2. Massive lavas can be extruded in very deep water, but where massive flows have red tops, then they are almost certainly subaerial flows.
3. Where lavas enter the sea they may produce hyalotuff deltas, the presence of which in the volcanic pile would be a certain indication of volcanicity at or above sea level.
4. Accretionary lapilli require the volcanic ejectamenta to have been through the air column: they do not form in water alone.

5. All the above points above refer to water depth, which is obviously very significant. If a thick sequence of lava is built up from deep to shallow water, then it may have formed in an ocean island environment, possibly at the early stages in the growth of an island arc or (very unusually) a mid oceanic ridge. However if there is a thick sequence of lavas which are constantly extruded into shallow water, then we have to invoke subsidence at the same time as lava accumulation. This feature is common in island arcs, possible in unusual mid-ocean ridges and unlikely in oceanic islands.

Lavas are evidently important indicators in the evaluation of the evolution of an ophiolite, and they will be examined particularly in the light of the points made above.

The age of the Ballantrae Complex

With the Ballantrae Complex comprising both igneous and sedimentary rocks, dating has been carried out using both radiometric and palaeontological techniques. This has the advantage of being able to fit the radiometric into the palaeontological time-scale. The black shales which occur amongst the lavas, and those which are part of the **olistostrome** sequence, have been known for some time to contain fossils of mainly inarticulate brachiopods and graptolites. The latter are particularly useful in relative age determination and they indicate that these rocks are representative of most of the Arenig Series. Radiometric dating has been conducted on a variety of rock types using a range of methods, and with two exceptions yield ages which on the basis of world-wide data are considered to be Arenig (Fig.25.2). The exceptions are within the olistrostrome-**mélange** unit at Knockormal, where a garnet meta-pyroxenite has yielded an age of 576+-32Ma which is Cambrian; and the pillow lavas at Downan Point which have yielded younger ages of 468+-22Ma, which is roughly Llanvirn. The errors on either side of the mean in each of these determinations are large and, in the latter instance the age is not statistically different from age determinations from other pillow lavas to the north of Ballantrae (north of the Stinchar Valley). There is, however, a essential diffence between the lavas to the north and south of the Stinchar Valley; those to the north have a higher proportion of volcanogenic sediment. Downan Point is typified by massive pillow lavas with a minimum of volcanogenic sediment and this probably

reflects the different regime. Indeed many workers would now place the Southern Uplands Fault along the Stinchar Valley to separate the pillows of Downan Point from the rest of the ophiolite.

From these radiometric ages and from the ages given by the faunas it is clear that the main part of the ophiolite was formed within the Arenig, between *c.*501 and *c.*476 Ma. a time-span of *c.*25 my. The age of obduction is anytime between 501-476 Ma , so it was also obducted within Arenig times. These ages imply that the oceanic crust which comprises the Ballantrae Complex was young and near to the site of its generation: wide ocean basins have oceanic crust which is often >100 Ma, since it has travelled a great distance from the ridge which created it. Thus, the diversity of the complex cannot then be explained by the great differences in its age: it has to be explained by differences within the region of its formation.

There are several papers which review the nature and origin of the Ballantrae Complex in terms of its ocean crust setting. The earliest of these are by Church and Gayer (1973), Dewey (1974), Bluck *et al* (1980) and Stone and Smellie (1988). The last is also a comprehensive guide to the complex with much new and significant information.

The significance of the Ballantrae Complex

The geological significance of the Ballantrae complex extends far beyond the region of Ballantrae. The presence of oceanic crust leads to a number of important conclusions, some of which have helped to unify a geological history over a considerable part of Scotland. The prime conclusion is that during the Arenig this part of Scotland was a destructive margin, where oceanic crust was being consumed. This further suggested that to the continent side of this margin there would have lain a volcanic arc; this, on the basis of information from the overlying Ordovician rocks (see Excursion 28 Locality 3) is thought to have lain to the NW. To the south there would have been an ocean.

The nature of the Ballantrae Complex is also highly significant. If it was produced in a marginal basin, as suggested here, then there would have been a major subduction zone to the south where dense, and probably old oceanic crust would have been consumed (marginal basins are at present seen to form where old oceanic crust is being consumed, as in the western Pacific). This in turn would suggest that there had been quite a long history of subduction in the Ballantrae region.

The North Atlantic region has a number of ophiolitic masses which are of this general age. They occur in Newfoundland, Scotland and Scandinavia (Dunning and Krogh 1985).

References

References are given after Excursion 31. However, attention should be drawn at this stage to the valuable systematic account of the Ballantrae area published by the British Geological Survey (Stone and Smellie 1988). It contains some helpful maps and photographs and should be used in conjunction with the ensuing excursion accounts.

Excursion 25 PINBAIN BLOCK

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**)

Locality 1. South end of Kennedy's Pass NX (146 928): **Lavas and breccias** (Figs. 25.7, 25.8). Park cars south of Kennedy's Pass, walk 50m 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.

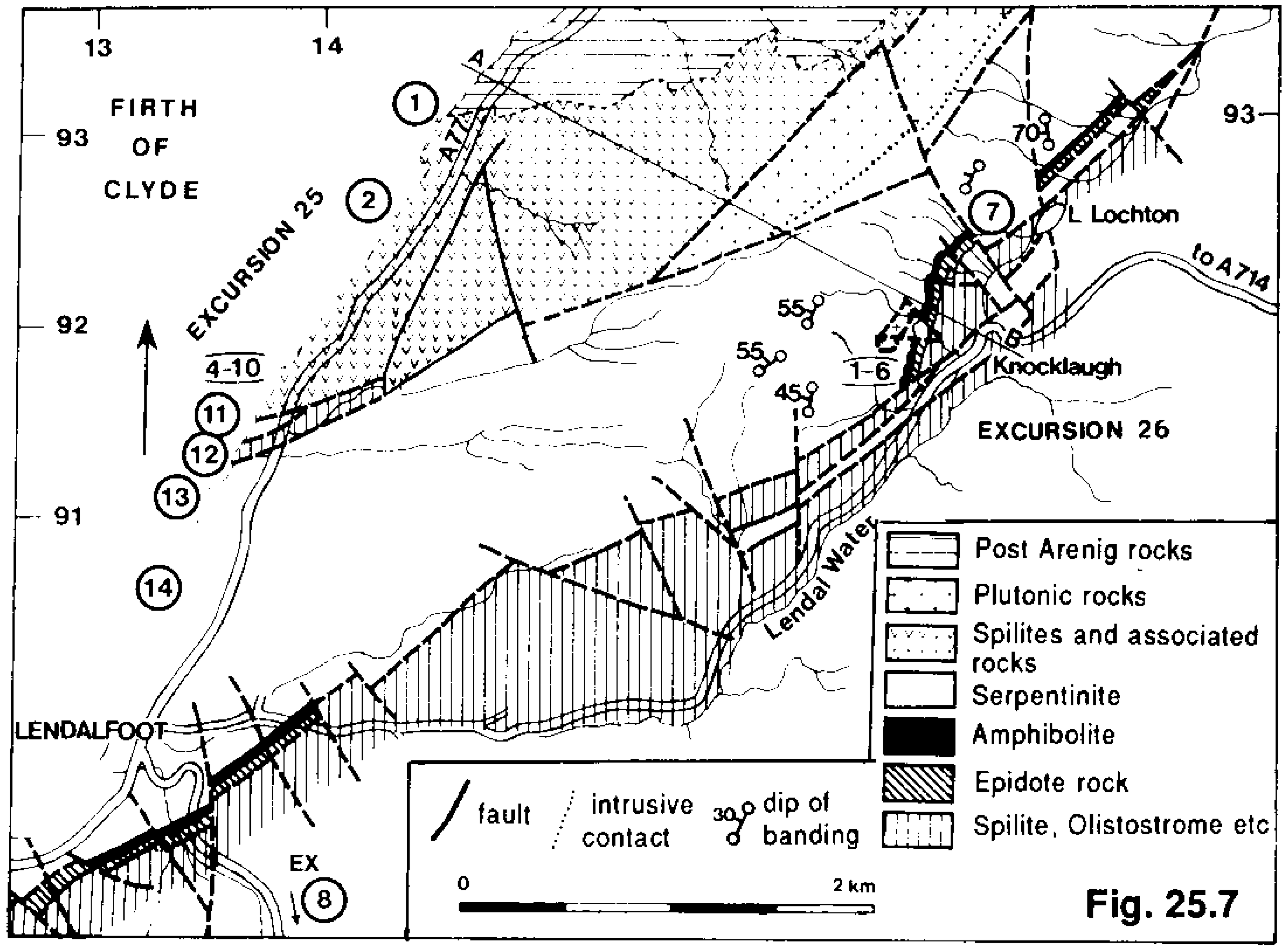
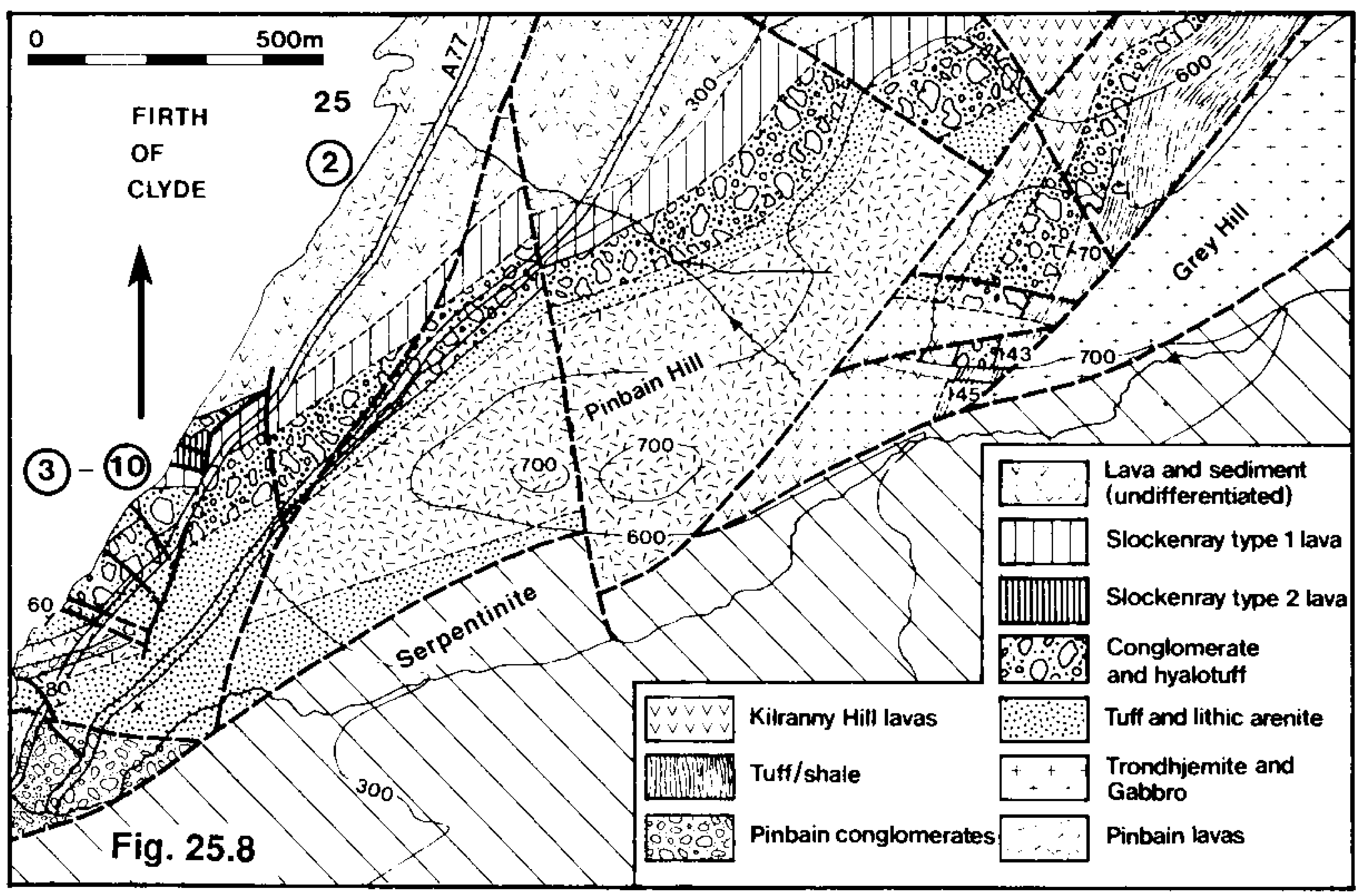


FIGURE 25.7. Simplified map of the northern part of the Ballantrae Complex, with positions of localities mentioned in Excursions 25 and 26.



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' (Fig.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 Fig.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

FIGURE 25.8. Map of the Pinbain Block showng lateral extent of some of the lavas.

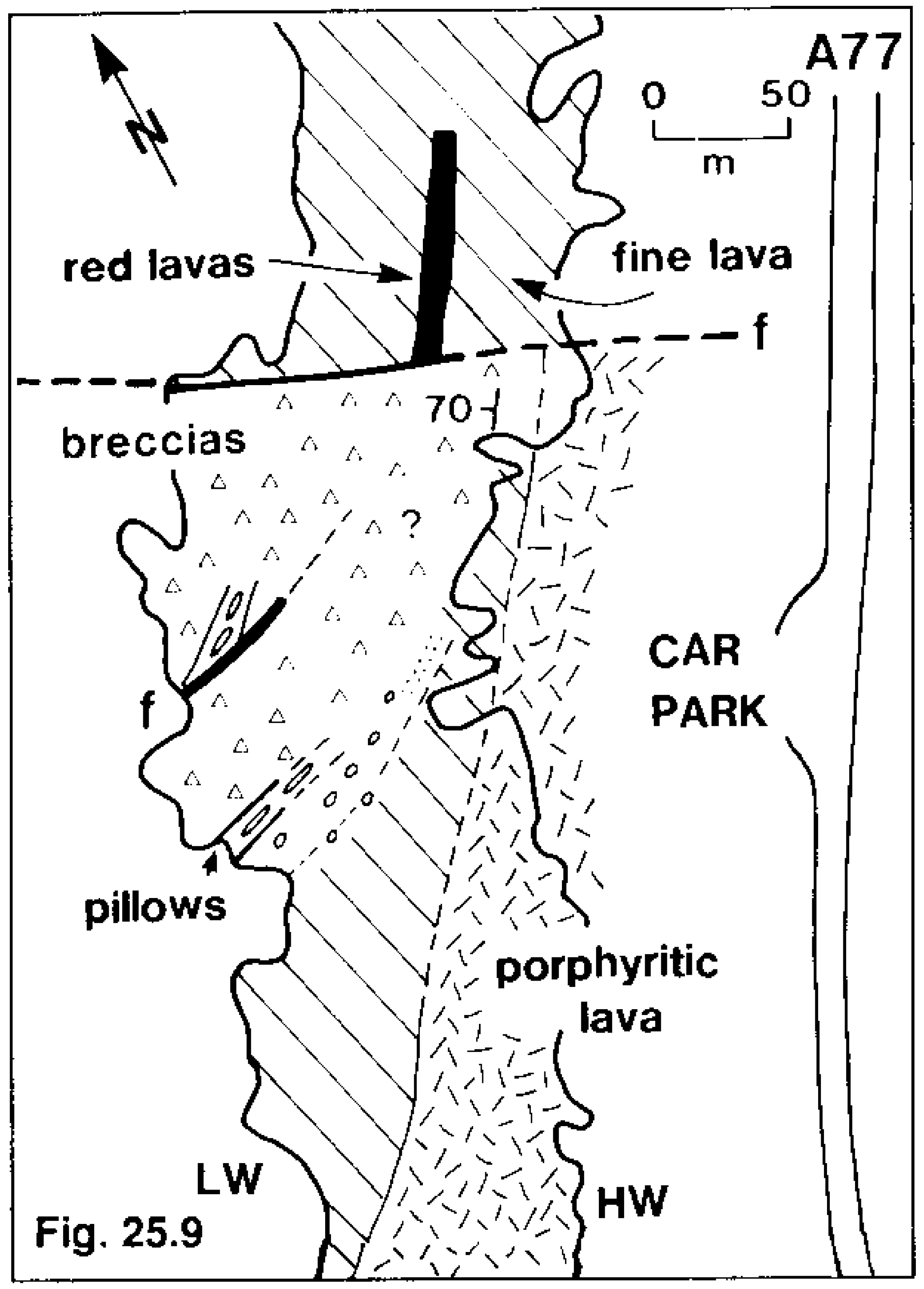


FIGURE 25.9. Sketch map to illustrate the geology at Locality 2, Pinbain Block. LW and HW refer to low and high water marks.

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** (Fig.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; Fig.25.8).

A detailed map of this headland is given (Fig. 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 (Fig.25.11) hyalotuff deposits overlie the porphyritic lavas, and interfinger with a dark aphyric lava-type.

Locality 4. Lavas (best seen at low tide). (Fig.25.10). There are two lava types in this sequence, each of which comprises multiple flows. The main one is porphyritic with abundant **phenocrysts** >1cm long of plagioclase (now mainly albite) arranged in a swirling fashion and

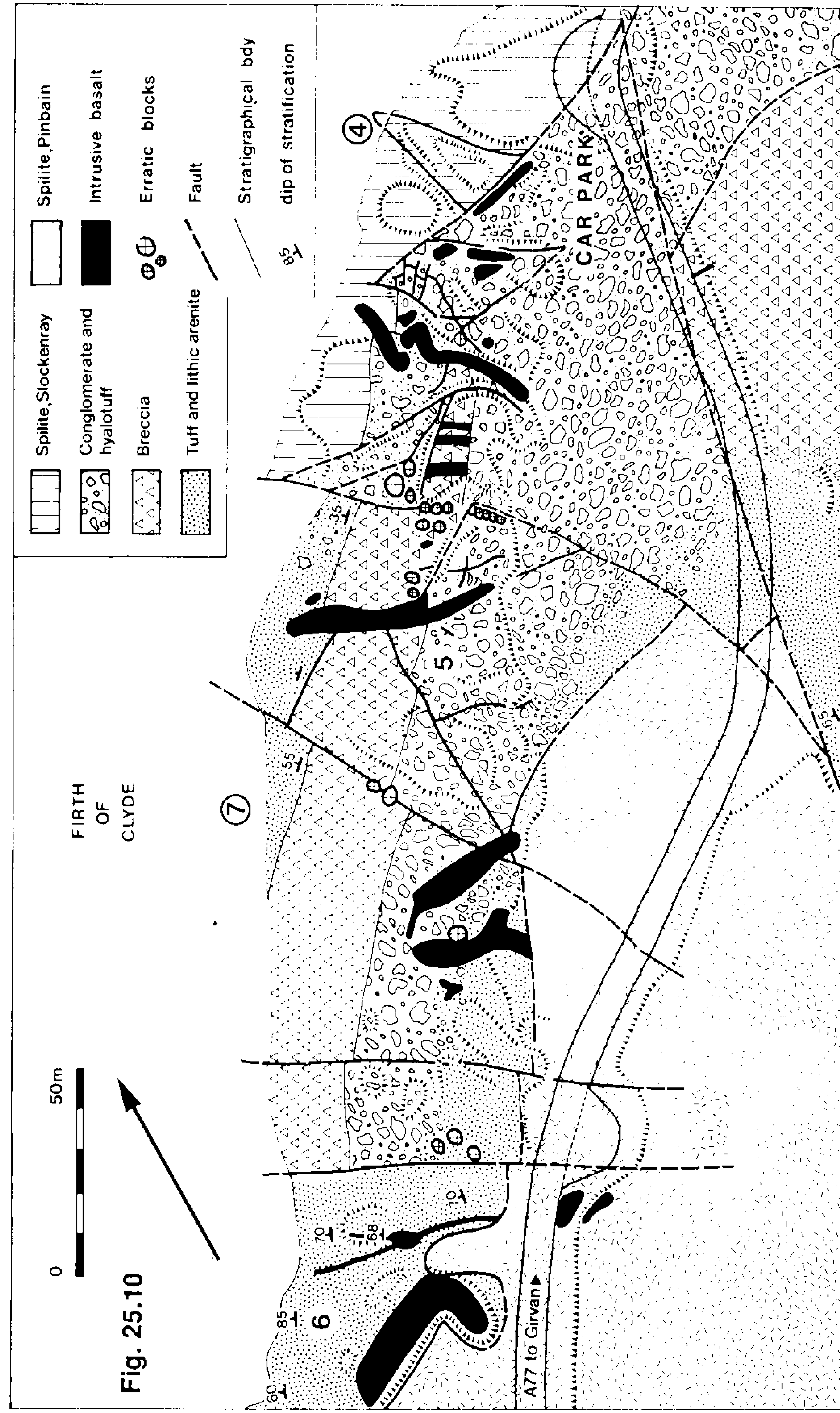


FIGURE 25.10. Map of the southern end of Slockenray. The north margin of map begins at the headland which divides this map from Figure 25.12, and the two lava types - porphyritic and dark aphyric (not subdivided in the map). They both belong to the Slockenray spilite of the caption.

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 (Fig.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 >80cm 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 bytownite. 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 Figs 25.11, 25.13). The sediment

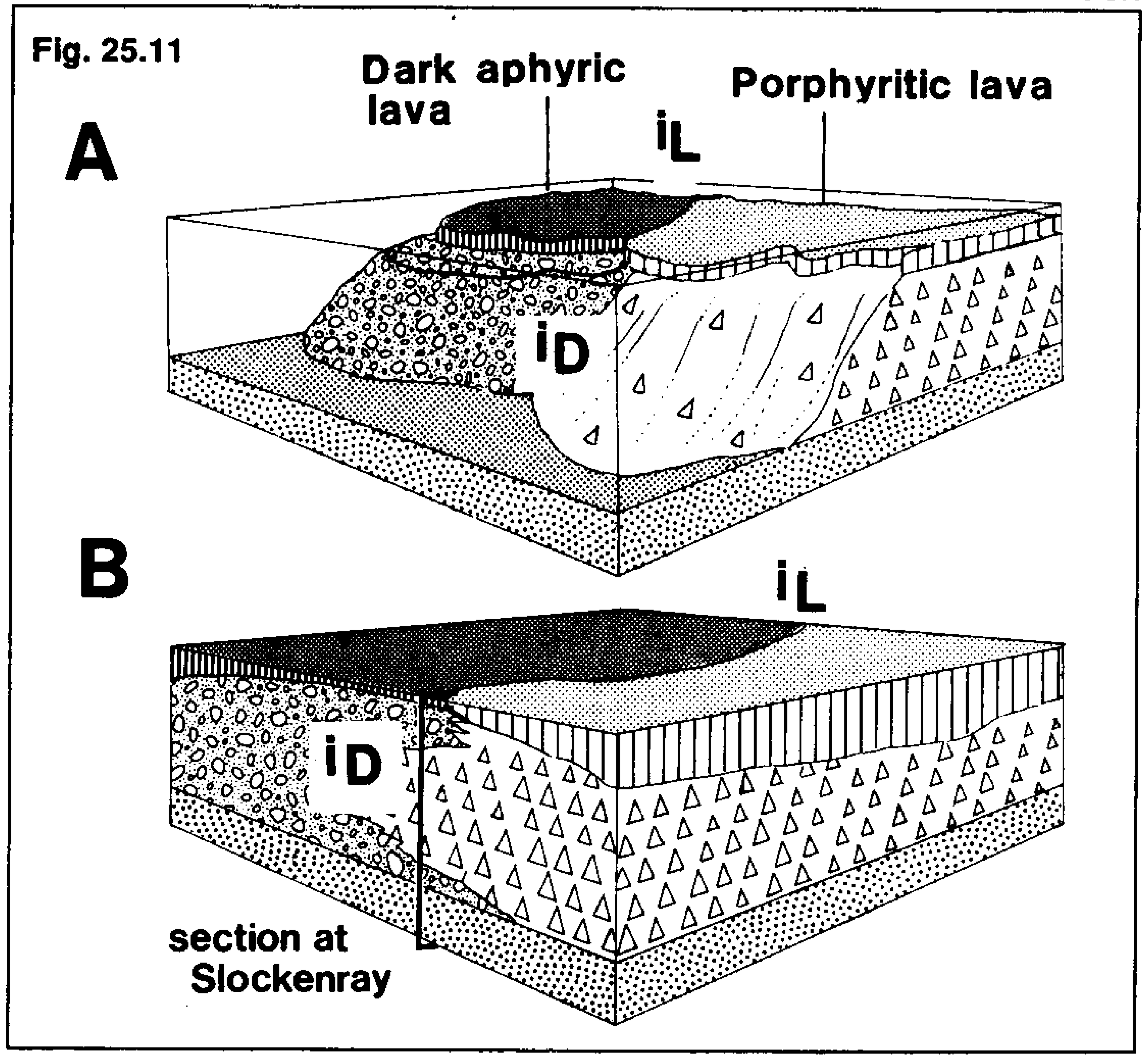


FIGURE 25.11. Explanation of the Slockenray sequence. **A.** Two lava flows, one porphyritic, the other dark aphyric, are both simultaneously extruded and flow together towards the coastline where they both build out delta cones adjacent to each other, each cone being sourced by their individual lava 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.

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 progradations 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 (Fig. 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 (Fig.25.10). **Graded** and ungraded beds 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; Fig. 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 (Fig.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 (Fig.25.10) and in the bays to the north (Fig.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 Fig.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 (Fig.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 (Fig.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

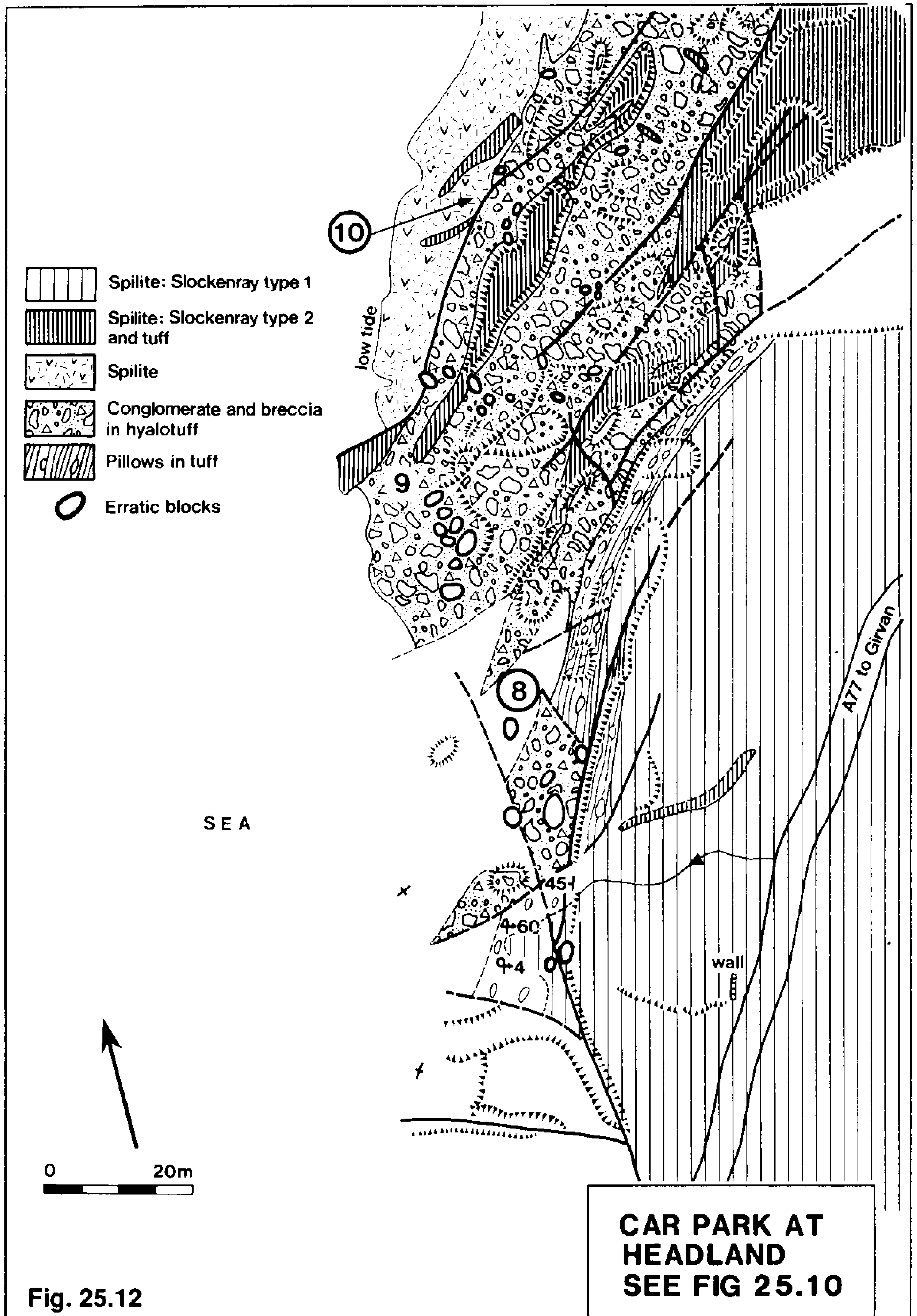


Fig. 25.12

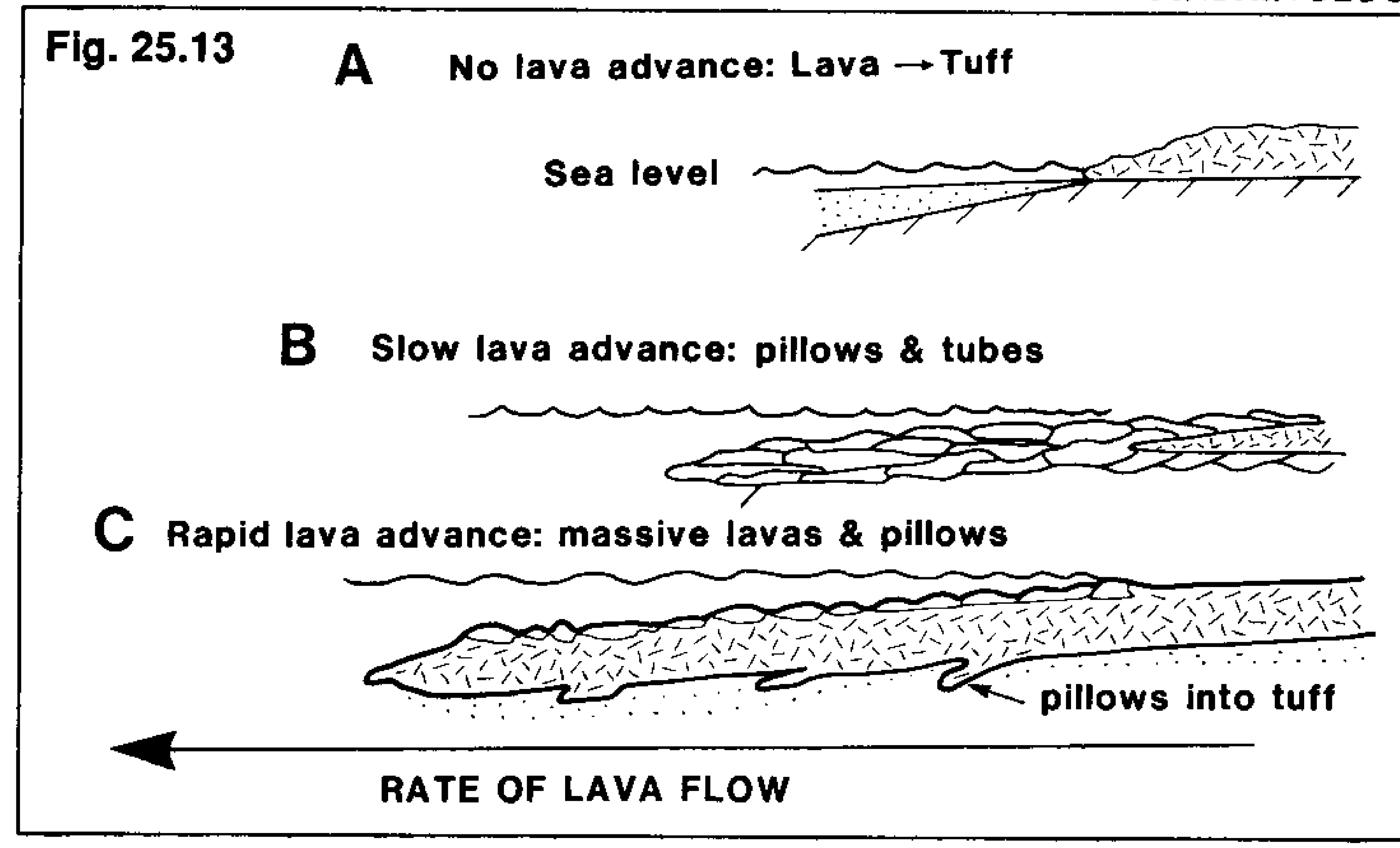


FIGURE 25.13. Explanation of the development of various lava structures and tuffs in lavas which enter the sea from the land. **A**, the lava front is moving slowly and as it enters the sea where it is rapidly chilled, all of it is converted to tuff at the shoreline. Waves and currents move the tuff 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.

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 (Fig. 25.12). Thin finger-like units of aphyric

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.

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 hyaloclastic deltas seen forming at the present-day.

Locality 10. Contact with the overlying porphyritic lavas (Fig.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 (Fig.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

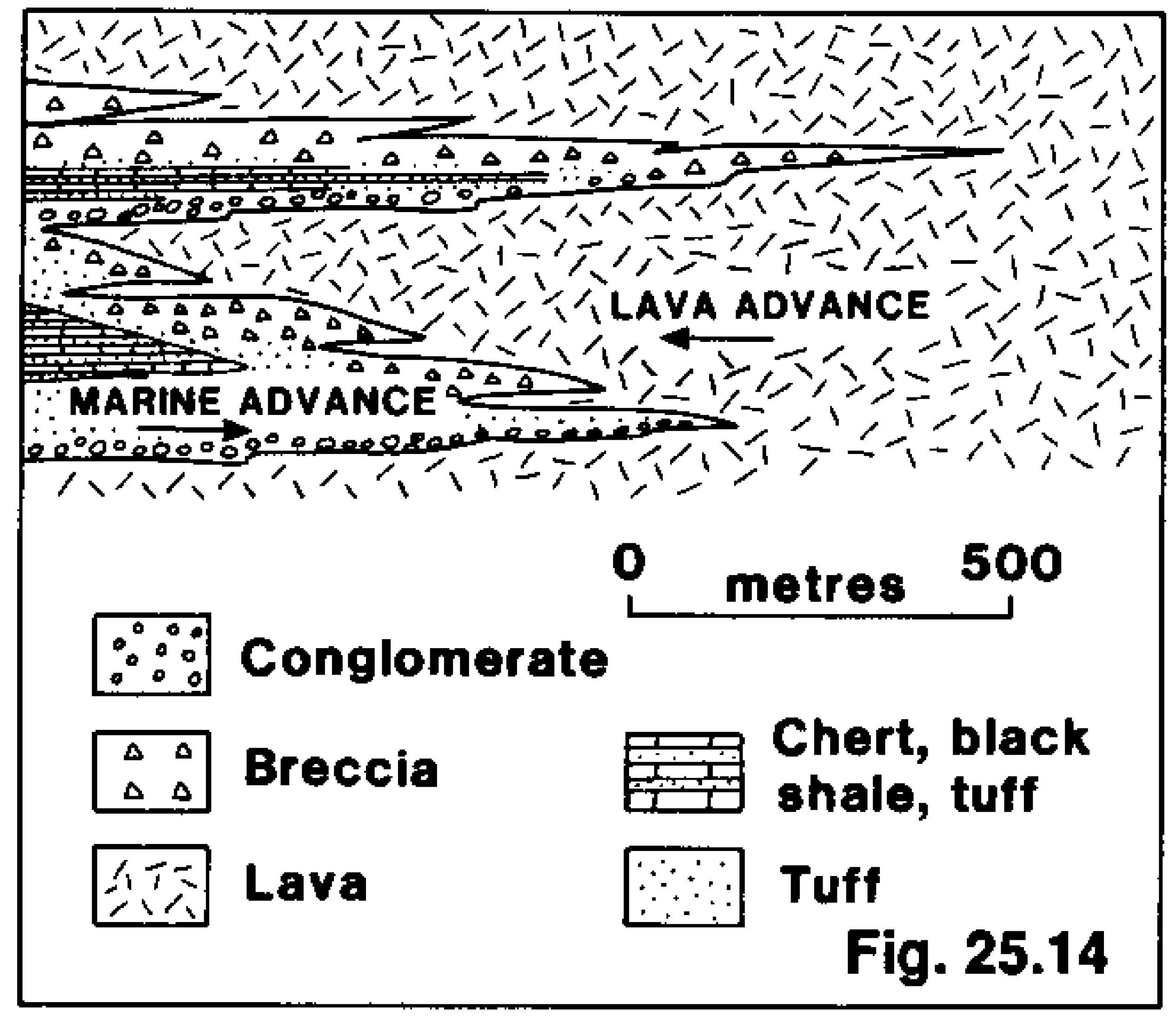


FIGURE 25.14. Explanation of the sediment-lava cycles in the Pinbain Block and elsewhere. When the rate of lava extrusion is rapid or the rate of sea-level change is slow, the lavas advance into the sea. Because of seawater-lava interactions, where the lavas break down by explosive or erosional activity, lavas are always associated with abundant breccias, often 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).

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 (Fig. 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 (Fig.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 137916): **Pinbain Fault and associated features** (Fig. 25.15 a,b,c,d,e,). The section begins at the north end of the beach to the north of Pinbain Burn (Fig.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

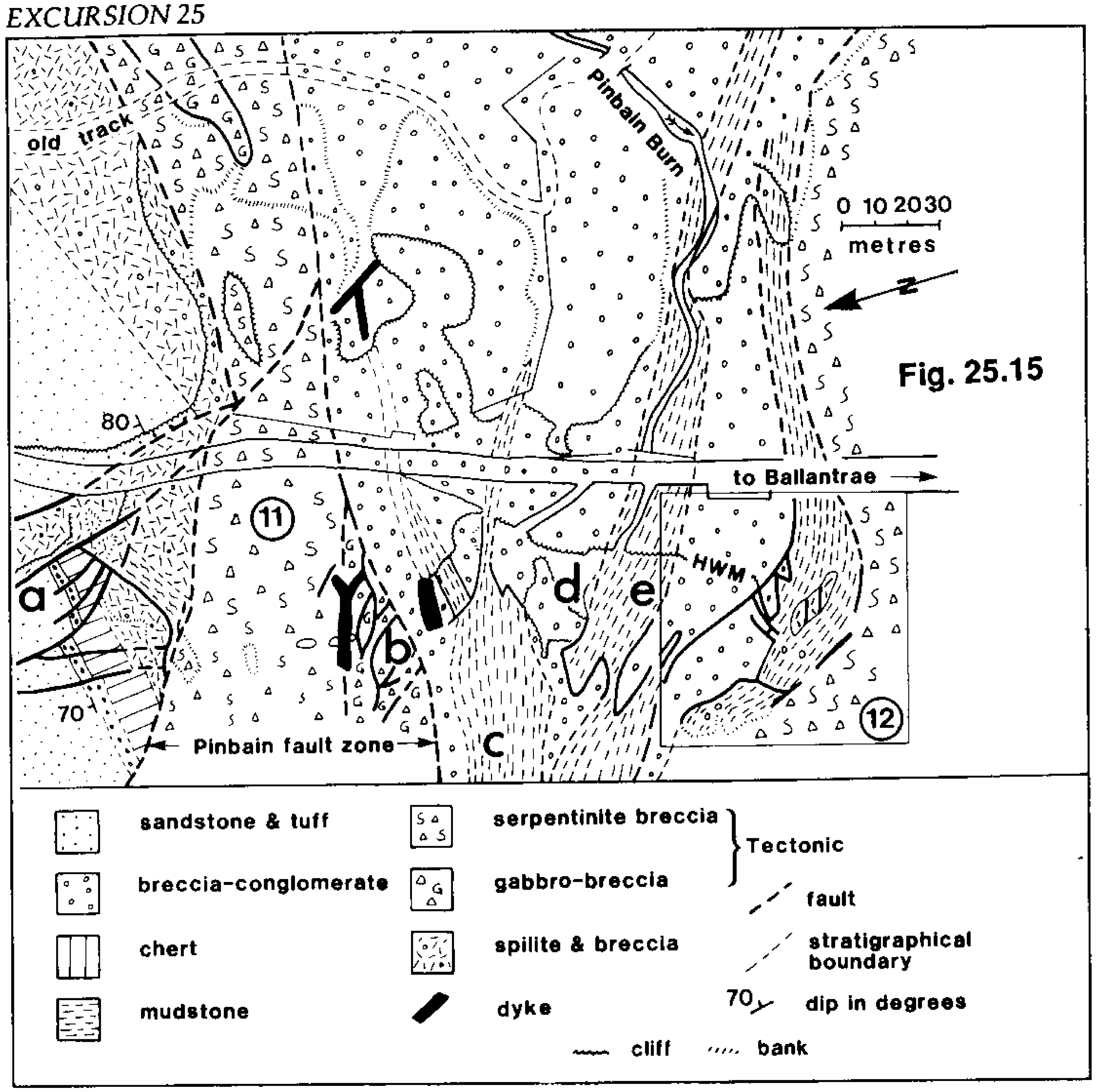


FIGURE 25.15. Plane-table map of the region near Pinbain Burn. Letters **a**, **b** etc refer to localities discussed in the text; inset is the approximate position of Figure 25.16.

belt lie the olistostromes and mass flow deposits of Pinbain.

(a). The sequence to the immediate north of the Pinbain Fault (**a** on Fig. 25.15) contains tuffs, lithic 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 breccia-conglomerate, 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**) (Fig. 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 cherty 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 20m in diameter (Locality 12; a of Fig. 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** (Fig.25.16). The olistostromes are exposed on the rock platform at **(b)** Fig.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.

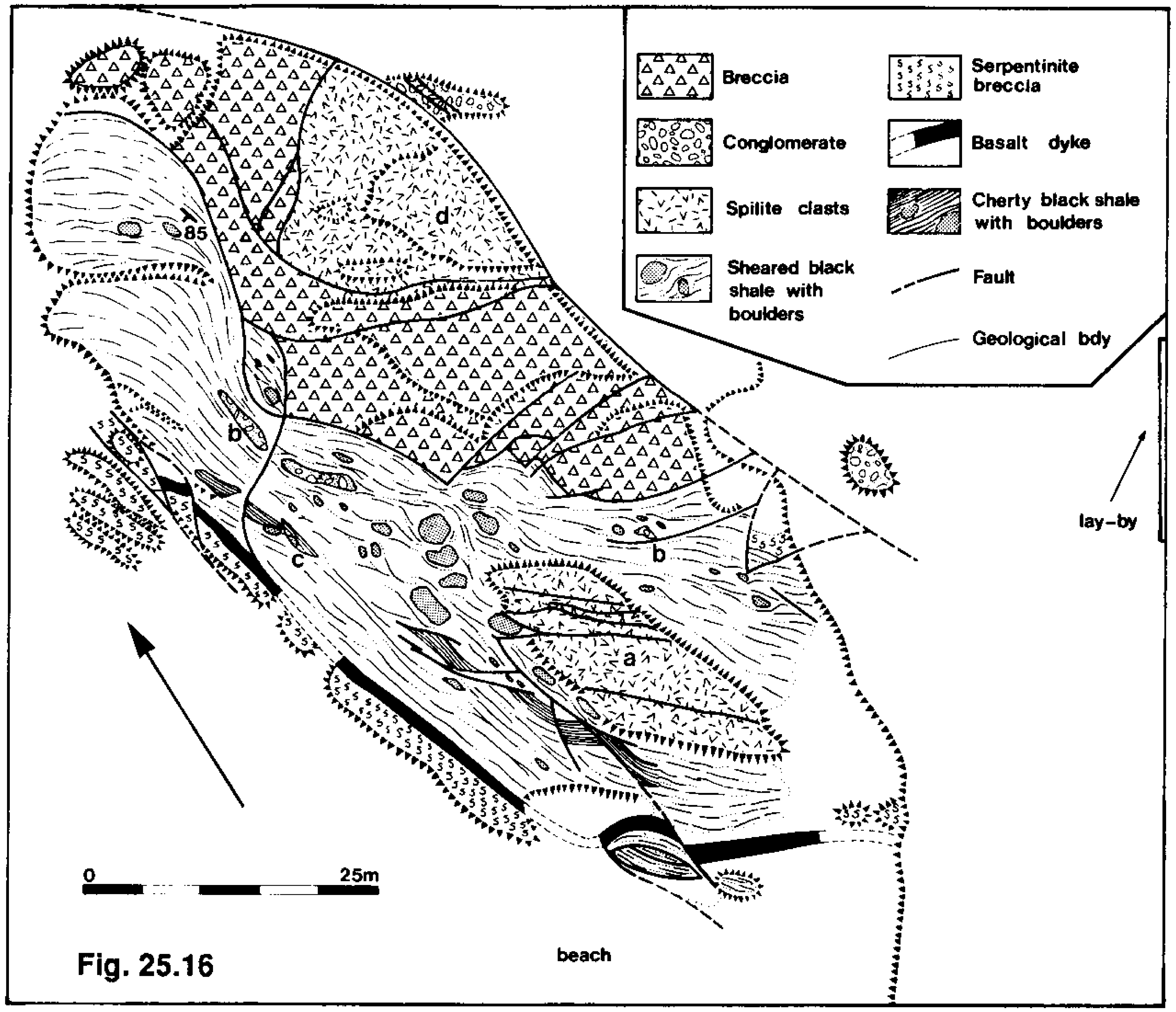


FIGURE 25.16. Plane table map of the ground south of Pinbain Burn (see Fig. 25.15 for position).

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 (Fig.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 (Fig.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 clastic sediment. For this reason they are not thought to be associated with an ocean ridge, although it

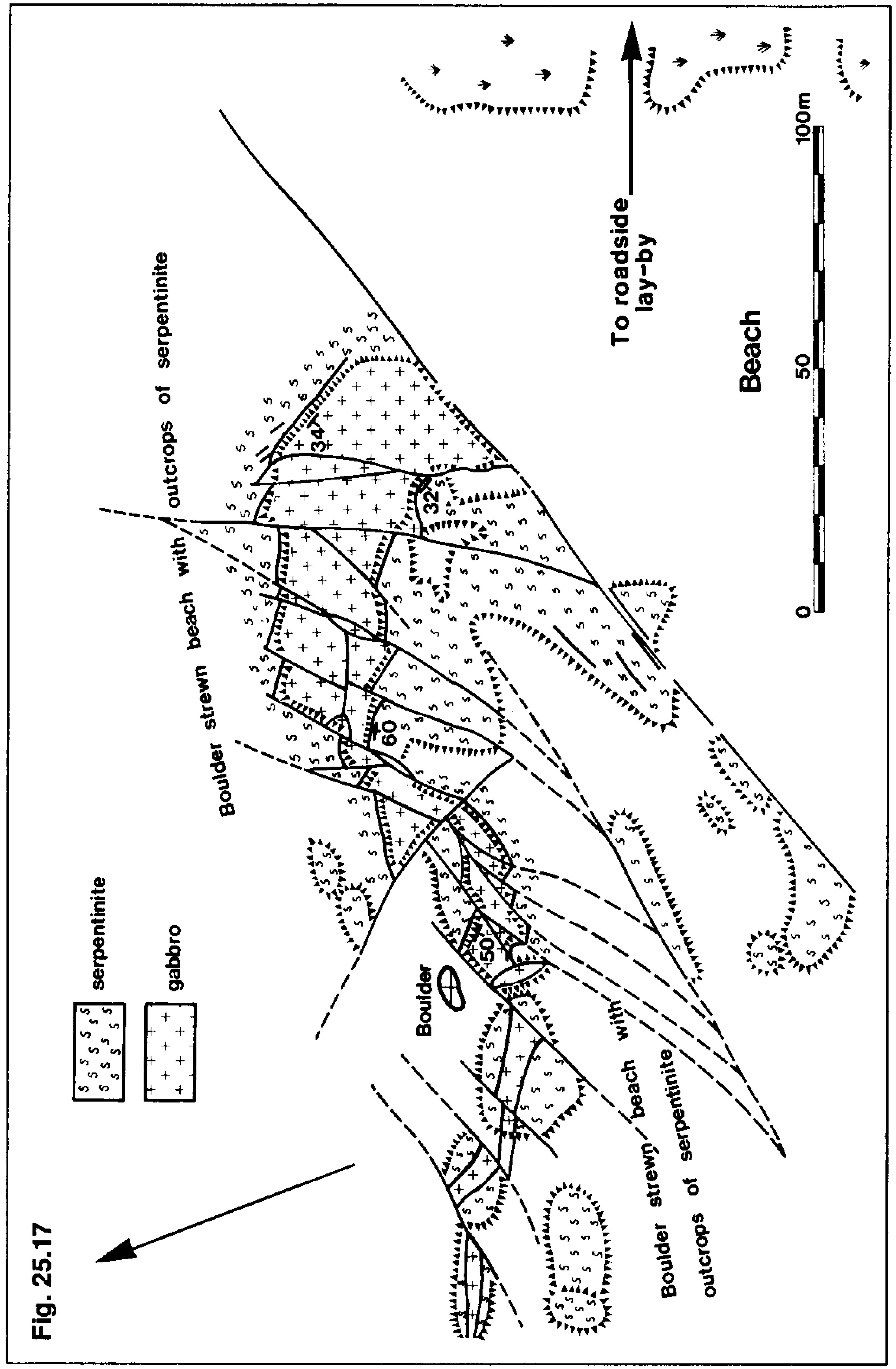


FIGURE 25.17. Plane-table map of Bonney's Dyke.

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** (Fig.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 are given after Excursion 31.

Excursion 26 KNOCKLAUGH

B.J.Bluck

Theme:	The metamorphic sole produced beneath the ophiolite during its emplacement (obduction); some of the ultramafic rocks of the overlying ophiolite; the dykes which cut these metamorphic rocks.
Features:	Metamorphism, mylonites, greenschists, amphibolites, garnet-metapyroxenite, olistostromes, dykes, pyroxenites, serpentinites.
Maps:	O.S. 1: 50 000 Sheet 76 Girvan O.S. 1: 25 000 Sheet NX 19 B.G.S. 1: 50 000 Sheet 7 Girvan 1: 25 000 Sheets NX 08, 18 and part of 19 Ballantrae
Terrain:	Moderately rough open ground, excursion begins with a steep walk.
Distance and Time:	2 km : 4 hours walking.
Access:	Permission should be requested from Knocklaugh Farm.
Short Itinerary:	stops 1-6

Introduction

The base of the northerly **serpentinite** belt runs from near the coast at Carleton Port, south of Lendalfoot to roughly the base of Cairn Hill, NE of Loch Lochton (Fig 25.7). In places this contact is marked by a metamorphic **aureole** comprising a zone of structurally bounded slices of rocks with highly contrasting metamorphic grade. The zone is not always well exposed and has often been totally cut out by a fracture which brings serpentinite in contact with **spilite** and **olistostrome** (Figs 25.7,26.1). This can be seen in the vicinity of Knocklaugh, on the NE and SW margins of Figure 26.1.

The component rocks which make up the aureole vary in thickness along strike. Here the amphibolites are well exposed and quite thick but elsewhere, as near the coast, the epidote mylonite rock is thick. Garnet metapyroxenite is particularly well exposed in this locality but is so thin everywhere that **it should not be collected.**

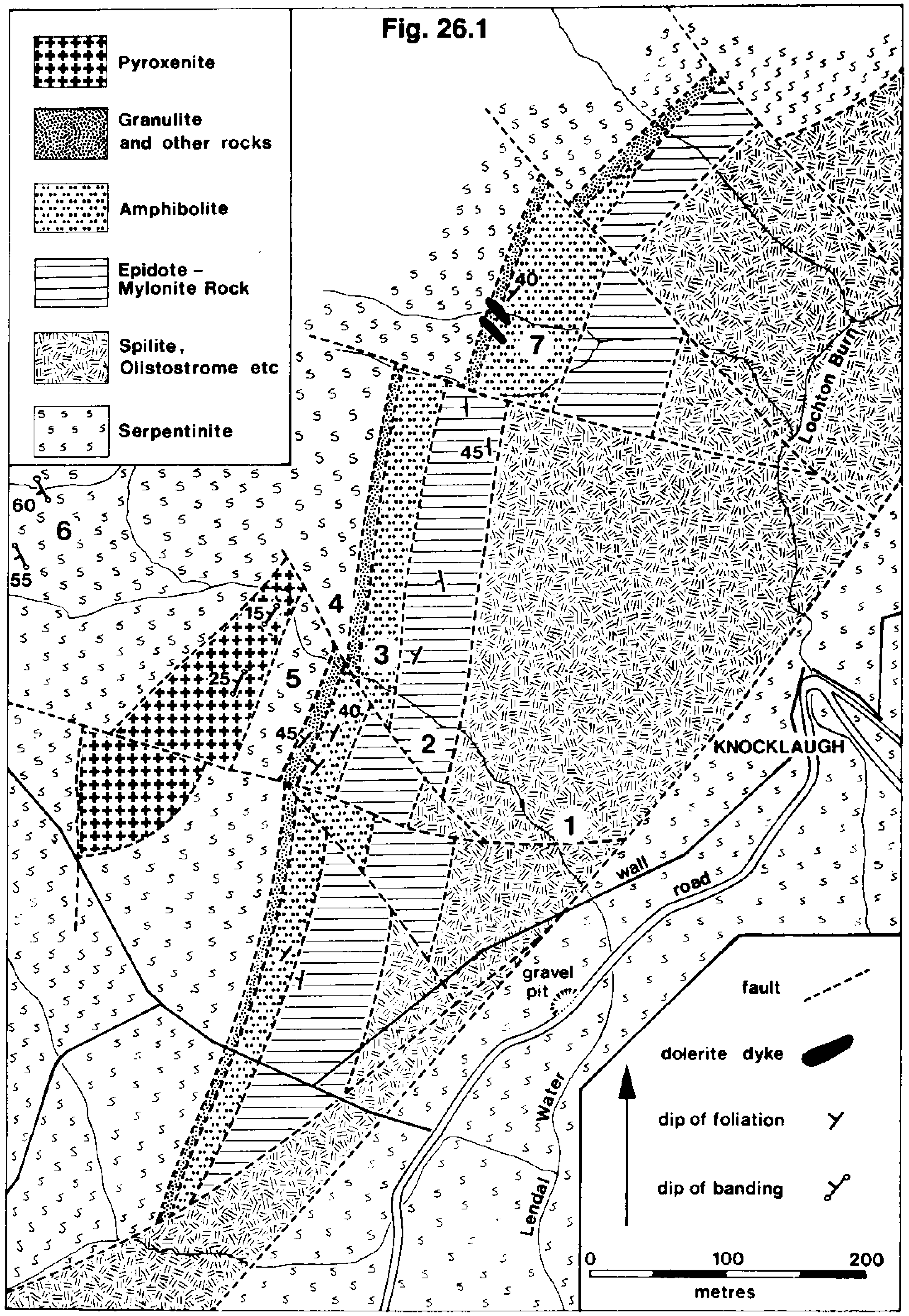


FIGURE 26.1. Map of the metamorphic sole at Knocklaugh, Excursion 26.

Locality 1. (NX 1692 9183): **Olistostromes** (Fig. 26.1). At the base of the waterfall sheared black shale is faulted against sheared serpentinite. The black shales have yielded graptolites to Peach and Horne (1899), and form part of a thick unit which comprises most of the immediate hillside. These black shales have thin graded beds of lithic-arenite and large clasts of basic and ultra-basic rock, carbonates and other sedimentary rocks. The graded beds comprise mainly grains of basic volcanic rock implying a source in a basic volcanic terrane. This unit, which is widespread in the ground to the SE of the serpentinite contact zone, is an olistostrome formed when large clasts roll or are otherwise displaced into an area where finer-grained sediments are accumulating. This type of deposit is typically generated near areas of submarine tectonic activity.

Commonly, at the top of the olistostrome unit where it is in contact with the metamorphic zone there are spilitic lavas which are often greater than 30 m thick. These have not been mapped as a separate unit although one interpretation of the metamorphic sole would require that they should be (see below). Within the ground covered by the map (Fig. 26.1) there are places where the metamorphic aureole is in contact with shale.

Locality 2. Banded epidote-mylonite rock (Fig.26.1). Banded epidote **mylonitic** rock, grey-green in colour is exposed in a small waterfall and adjacent banks. The rock has in many places a classical mylonitic texture with coarse augen of epidote, albite and quartz. Small-scale folds are associated with the augen, and larger scale folds are up 30 cm in amplitude. The epidote rock is composed of actinolitic-hornblende, chlorite, albite **porphyroblasts**, epidote (sometimes as porphyroblasts) and mica. They are seen elsewhere to interleave with sheared shales and **phyllites** and no doubt are partly derived from them; but some of the epidote rocks have the chemistry of basalts, suggesting that they had a basaltic **protolith**.

Locality 3. Amphibolite (Fig.26.1). Amphibolite, sometimes with excellent foliation, forms small exposures in the river banks and scattered over the hillside. The foliation generally dips to the NW but there is sufficient regional and local variation to suggest that there have been episodes of subsequent refolding which may be both ductile and brittle (see Spray and Williams (1980). Folded amphibolite is also fairly well seen at Locality 7. Garnets up to 20 mm in diameter occur in this amphibolite. The amphibolite is seen outside this area to interleave with the epidote rock and Spray and Williams (1980) have further subdivided the amphibolite into a lower, and an upper, the latter distinguished by having fairly abundant garnets. The amphibolites have a chemistry suggesting a protolith in basaltic rocks. The temperature-pressure regime under which the amphibolites formed is thought to be a minimum of 7kb and 850 ^{o}C. This pressure is equivalent to a depth of burial of 21 km.

Locality 4. Serpentinite and garnet metapyroxenite (Fig.26.1). Sheared, tough, platy tremolite-bearing serpentinite outcrops in this and many other sections along strike. These rocks are replaced upstream by unsheared, banded serpentinite which represents the sheared base of that segment of the oceanic lithosphere which was involved in the obduction event. A few metres upstream from this point is the outcrop of the garnet metapyroxenite, although there are only a few scattered small blocks now to be seen. This rock is thought to have formed under pressures greater than 10 kb (= >30km of burial depth) and temperatures of about 900 ^{o}C (Treloar *et al* 1980).

Altogether the rocks exposed at Localities 2-4 represent the sheared slices of the metamorphic sole to the ophiolite.

Locality 5. Pyroxenite (Fig.26.1). This bold feature on the hillside to the NW comprises altered pyroxenite. It has a pod-like outcrop, is bounded by shears and is probably a structurally detached block caught up during the obduction of the serpentinite and incorporated into it. The rock is an olivine websterite with evidence of banding produced by tectonic granulation.

Locality 6. Serpentinite (Fig.26.1). Serpentinite, which forms much of the high ground in this area, is exposed at a variety of places. It is a fairly tough foliated rock with foliations being almost N-S and striking into the zone of metamorphism at the base of the serpentinite. This discordant foliation is probably the result of tectonic overthrusting and rotation of an original foliation which may have been produced in the mantle during the movement of the oceanic plate away from the ridge where it developed. The original ultra-mafic rock was mainly a lherzolite comprising olivine, enstatite, diopside and picotite. However, plagioclase-bearing ultramafic rocks associated with these lherzolites would certainly not have formed in the same pressure-temperature regime. This evidence together with the granulation of

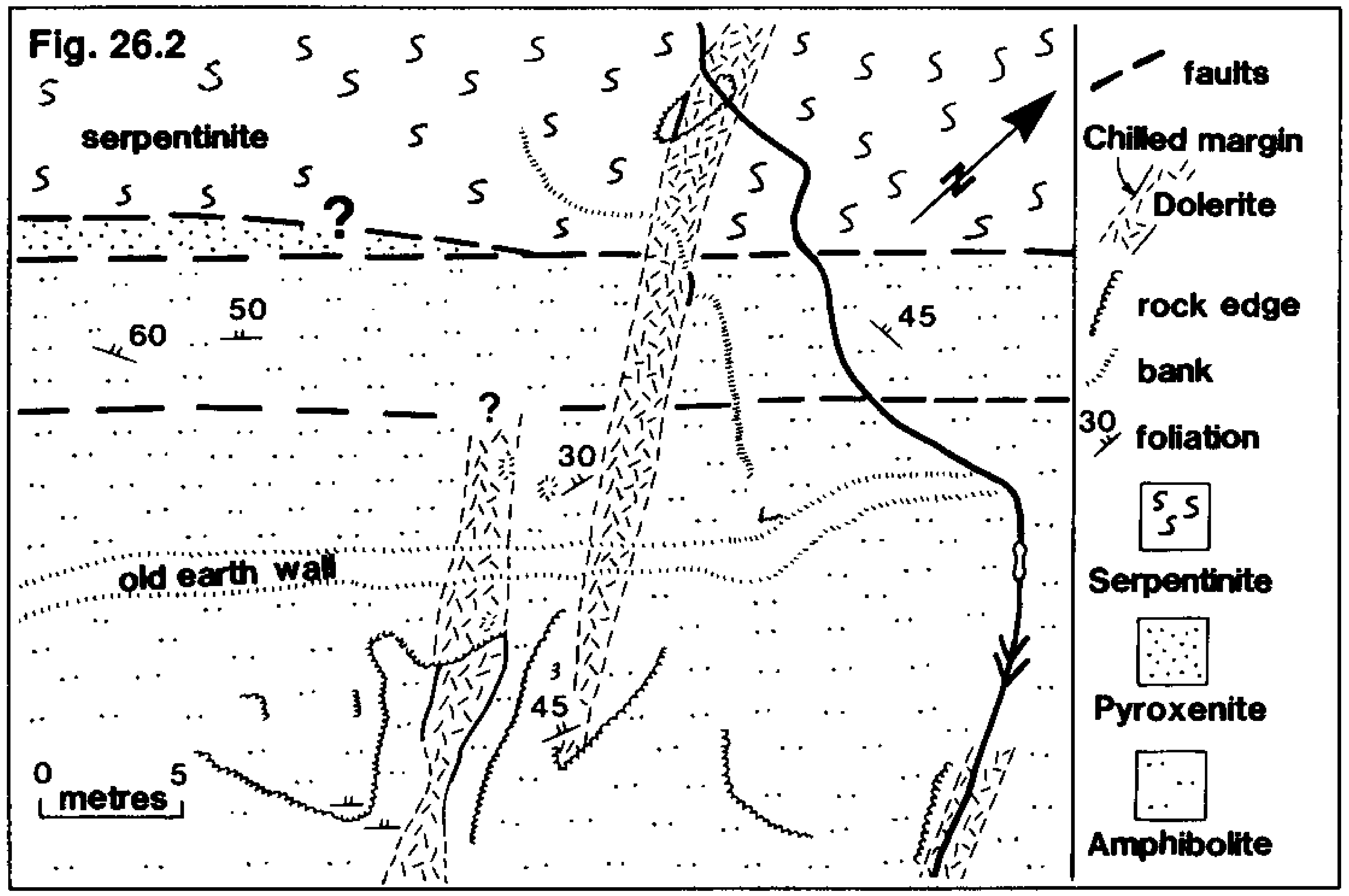


FIGURE 26.2. Detailed map showing the relationship between the dykes and the metamorphic sole at Knocklaugh (see Locality 7).

the ultramafic rocks suggest structural interleaving within the serpentinite.

Locality 7. Dykes cutting aureole (Figs 26. 1 & 2). Returning now to the metamorphic sole it is possible at this locality (Fig. 26.2) to study the rocks of the thermal aureole cut by dykes of the type seen along the coast (Excursion 25 Locality 14) and also at Carleton Bridge. The dykes are porphyritic and have distinct chilled margins against the amphibolite and serpentinite into which they intrude. In outcrop they clearly cut across the metamorphic zones with no displacement, thus indicating that both the development of the zones and their reduction in width were accomplished before dyke intrusion. It is also clear that the serpentinite was fairly cold at the time of dyke intrusion. Chemical and mineralogical analyses of these dykes indicate that they belong to the second phase of intrusion as identified by Holub *et al* (1984). At this locality it is possible to see some well developed banding and preferred mineral growth in the amphibolite and also some metre-scaled folds. The amphiboles from this locality yielded ages of 476+-14 Ma which represent the cooling time of the amphibolite during its obduction.

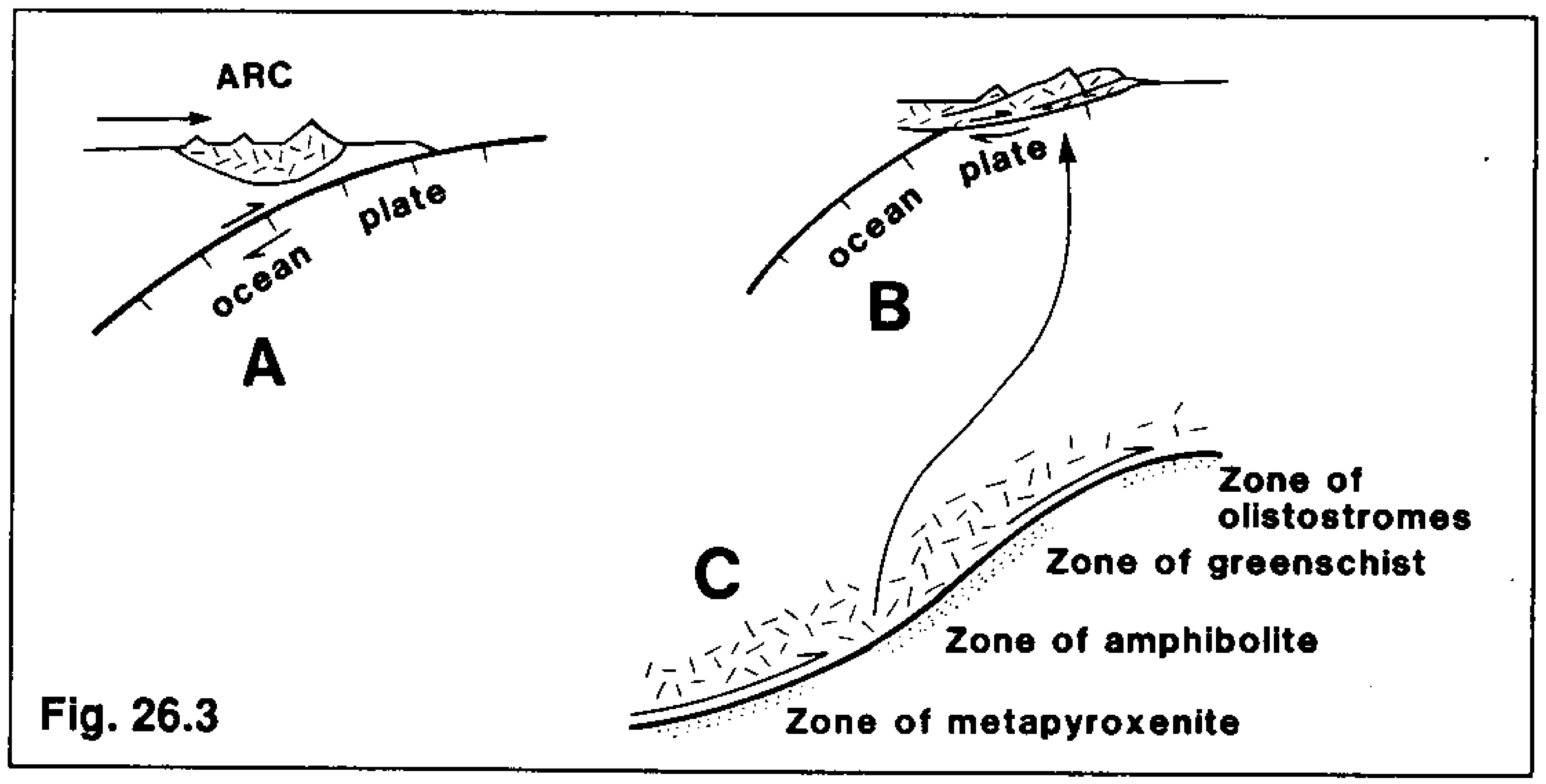
Here, as elsewhere along the outcrop, there is a zone of sheared tremolitic serpentinite between the outcrop of amphibolite and the meta-pyroxenite. The serpentinite and metapyroxenite may belong to the late stage history of the thermal aureole, being emplaced during the time when the originally much thicker aureole was being sheared down to the thin representative now present. The metapyroxenite at this locality has a cooling age of 505+-11 Ma.

From this position it is possible to see Grey Hill and the tops of Byne and Mains Hill to the NW. These are gabbros, diorites and trondhjemites which have intruded into the serpentinite. To the SE lie the lavas of Aldons, the southern serpentinite and pods of sheeted dykes of Millenderdale and Fell Hill and beyond that the higher hills of the Southern Uplands.

There are some paradoxes concerning this thermal aureole which need discussion.

1. In most metamorphic complexes, where there has been no tectonic inversion, the pressure-temperature path increases downwards in the sequence. This example records the reverse of this: the rocks indicate a pressure-temperature decrease down section from the serpentinite to the black shales and olistostromes.

FIGURE 26.3. Possible explanation of the metamorphic sole to the ophiolite. **A** an arc, because of changes in the location and sense of subduction, is driven towards the source of the plate which created it. In colliding with the underriding plate it underplates onto it the high pressure rocks belonging to this oceanic plate (**B**). But only fragments of this plate are accreted to the sole of the arc (**C**).

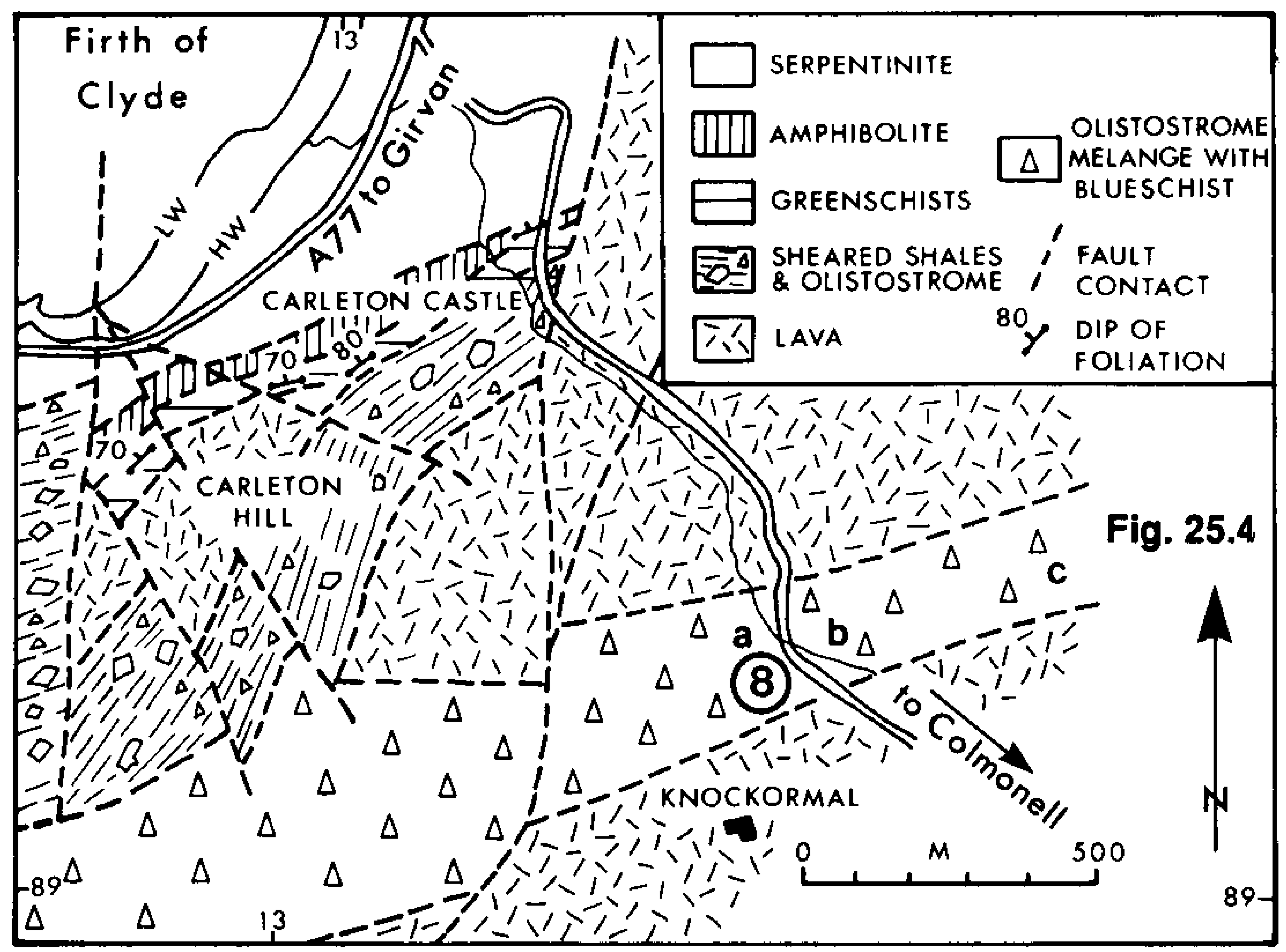


2. In a distance of <100m the metamorphic grade changes from low grade, graptolitic shales to high grade metapyroxenite. This change in metamorphic grade, would normally have taken place over vertical distances of >30km, so that the boundaries between the various lithological units here must be greatly compressed. The faults which bound each of the metamorphic slices are therefore considered to have substantial throws.

3. One interpretation suggests that as the oceanic pile was being thrust onto the continental edge, shear planes developed within the ocean lithosphere and detached slices of lithosphere from differing levels along the subducting plate brought them together as a condensed sequence within the aureole (Fig. 26.3 B). These slices were then attached to the upward moving hot plate and became the metamorphic sole to the ophiolite.

4. The protolith to the amphibolite and garnet meta-pyroxenite is mafic rock similar to that found in layer 2 of the oceanic crust. If the P-T conditions of formation of these rocks require that they come from > 20-30 km depth, then how can layer 2 occur that deep in the

FIGURE 26.4. General map of the Carleton Hill - Knockormal area, Locality 8.



lithosphere? The obvious solution is to have it descend down a subduction zone and then become detached as outlined in Figure 26.3 B. There are, however, many other ways of effecting the obduction, and these can be followed up in the literature.

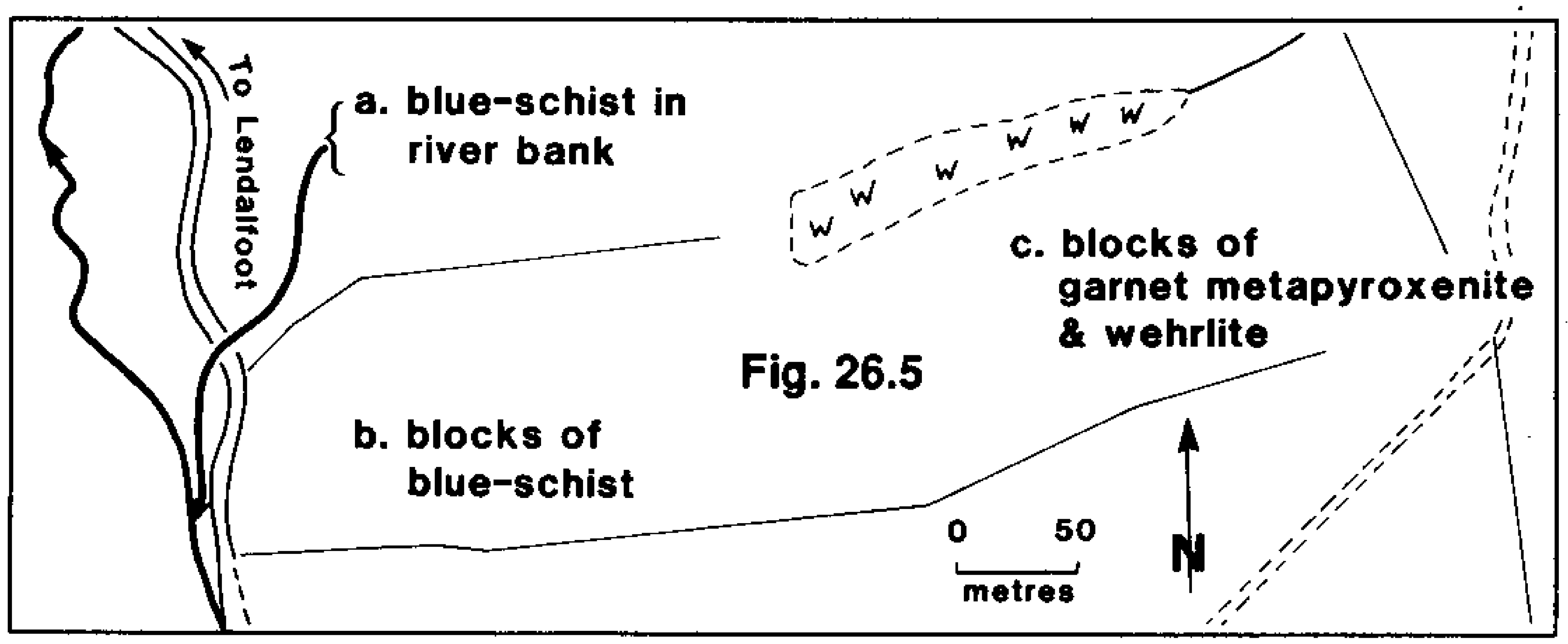
Return to the road and travel west to Lendalfoot. At Lendalfoot turn left at the A77 and then immediate left again to take the narrow road south towards Garnaburn, which is NW of Colmonell. After passing Carleton Castle and Little Carleton stop at the first cattle grid where there is a stretch of broad open marshy ground. Park south of the cattle grid. The exposures form the high northern edge to the low ground.

Locality 8. (NX1375 8890). **Blue-schist, pyroxenite and wehrlite** (Figs. 26.4, 26.5). The whole of this poorly exposed ground is characterized by blocks of granulites, lavas, limestones, cherts, granites, blue-schist, pyroxenites, serpentinites and wehrlites. In some instances these blocks are clearly in a highly deformed and mylonitic matrix; in others they are associated with less deformed black shales. It therefore seems reasonable to interpret the unit in which the blue-schist occurs as a mélange into which a wide variety of different rock-types have been emplaced.

. Blue-schist is exposed in the river section to the west (**a**) and in the field to the east are crags of blue-schist (**b**), and some 200m to the east of these are further blocks of pyroxenite and wehrlite (**c**) Fig. 26.5).

References are given after Excursion 31

FIGURE 26.5. Detailed map showing the locations of the outcrops discussed (Locality 8).



Excursion 27 BENNANE HEAD TO DOWNAN POINT

B.J.Bluck

Themes : Accretion of a thick sequence of lavas and sediments; types of lava and associated sedimentation; sedimentation of cherts and conglomerates and their significance; the structural significance of the contacts between the lavas and the serpentinite which bound both the north and south margins, but are examined only in the south.

Features: Sedimentation of cherts, slumping, tuff-beds and their significance; recognition of various volcanogenic features; evaluation of the tectonic regimes of sediment accretion and the origin of the ophiolite.

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 walking along foreshore, walking on cliff edge past headlands; **slippery and sometimes difficult rocks to cover.**

Distance and Time: Distance 2 km. The less agile should visit localities 1-4. 4-6 hours.

Access: Although all the localities are on the foreshore and are part of a coastal **SSSI** access is mostly via private roads or over private land.

It is vital therefore, that the rights of the owners should be respected.

These include Melville (Bennane Lea and Meikle Bennane), Shanklin (Little Bennane) and Melville (Troax). Otherwise access may be refused to these most significant and instructive geological localities. See details under particular localities.

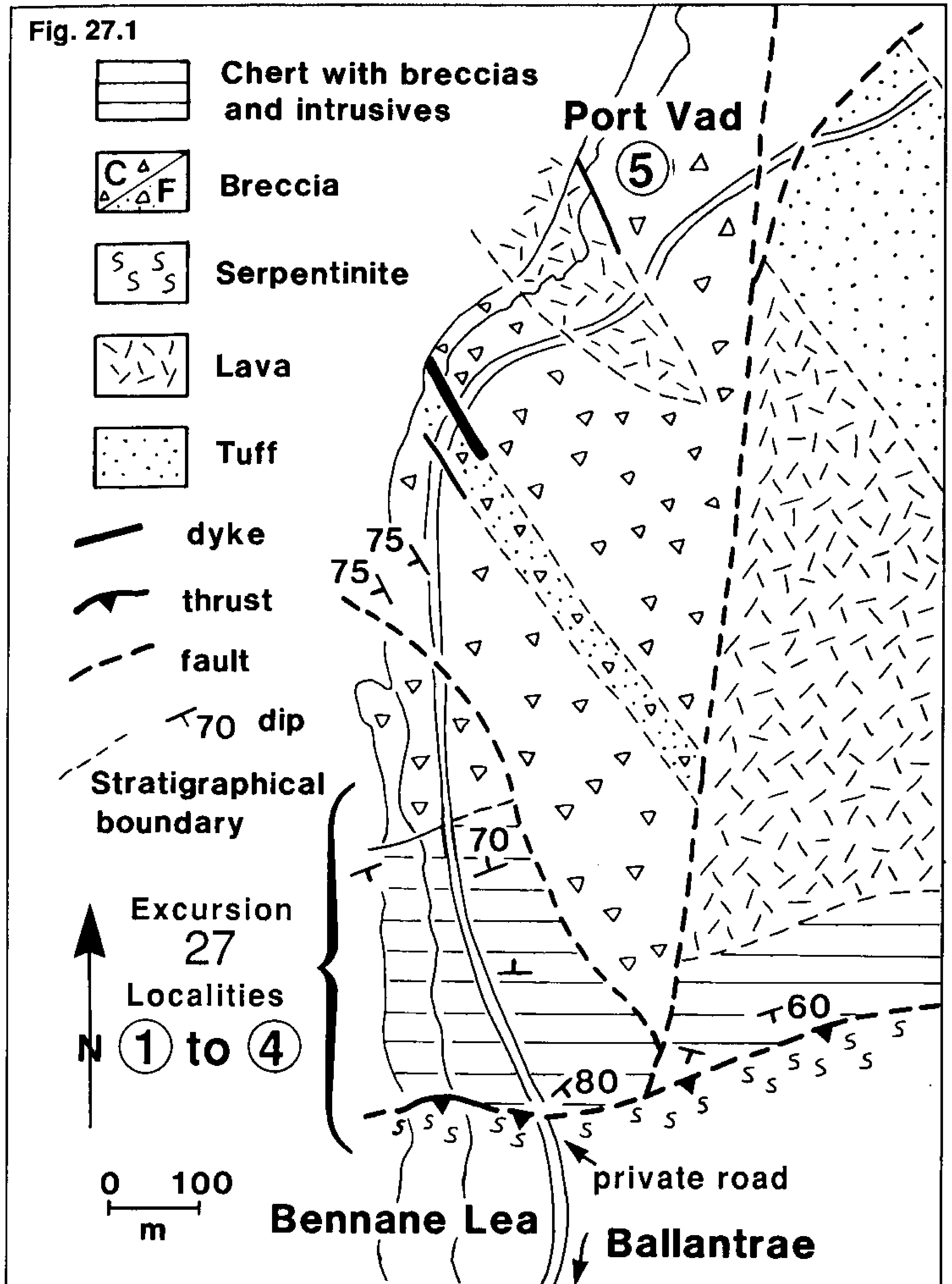


FIGURE 27.1. Simplified map of Bennane Head. In the breccia symbol, C refers to coarse, F refers to fine grain size. (Note that the A77 road has been rerouted to the east of Bennane Hill and reaches the coast at the south of this map.)

Some important problems raised by the Bennane Head section

There are four important points raised by this section which have considerable bearing on the origin of the Ballantrae Complex as a whole.

1. The origin of the sedimentary pile which sits above the lava sequence. In particular, was it of shallow or deep water origin? Was it deep but close to a source in shallow water? Was it deposited in a tectonically stable or unstable regime? Is it conventional layer 1 of the ocean crust - if not, what is it?

2. The nature of the volcanogenic pile; did it form in deep or shallow water and what is the importance of water depth anyway? Are the lavas and the sediments they produce all basic in composition? Is this sequence conventional layer 2 of oceanic crust?

3. If these aspects of the section (1,2 above) are different from conventional oceanic crust, what is the significance of that difference?

4. What is the nature of the contact beween high level **spilitic** lavas and mantle-depth **serpentinite**; and when did that juxtapositioning occur?

The exposures on this excursion are very good, confused in places by faulting, but in general contacts are clear even if their interpretation is not. There is evidence for structural repetition of parts of the sequence (Stone and Rushton 1983) so it may not be as thick as previously supposed (Bluck and Halliday 1981). Some of the lavas from this block have been analysed for their chemistry by Wilkinson and Cann (1974), Lewis and Bloxam (1977) and Thirlwall and Bluck (1984) and all agree that chemically they resemble ocean island type basalts.

Locality 1. (NX 0921 8599) **Contacts beweeen Triassic sandstones and Arenig serpentinite; serpentinite and doleritic and tuff rocks belonging to the Bennane Head sequence** (Figs. 27.1, 27.3). The coastal road shown on Figure 27.1 is now private although it is assumed that well-behaved groups will be allowed to **walk** along it to Locality 5. The A77 road has been rerouted east of Bennane Hill and joins the old road at Bennane Lea where a lay-by is to be constructed. **Vehicles must be parked south of the cattle grid and well away from the cottages and their access routes.** Go onto the beach via a track which leads to a sand pit just south of Bennane Burn. Localities 1, 2, 3 and 4 can be studied by keeping entirely west of the road. Note the small monument to 'Snib', a bank clerk from Dundee, who finally settled for a life free of income tax and other boring attributes of modern living. He adopted a cave in the cherts as his home. Snib, a man suspicious of what he assumed to be 'authority', had many very articulate discussions with this writer about what he could do with the subject of geology and where he could put the plane table set-up used to draw Figure 27.3.

The southern margin of the Bennane Head Block is bounded by a fault dipping at a fairly low angle to the south which brings volcanic rocks and cherts in contact with the southern serpentinite (Fig.27.3). Much of the foreshore and the raised beach platform farther to the south is underlain by Triassic red beds which along the raised beach cliff and at Ballantrae comprise breccias of serpentinite and other local rocks. The raised beach cliff is therefore a little more complicated than it seems at first sight: it is at or near to an old boundary of a Triassic basin. When there is little sand on the beach, as in winter, it is possible to trace out the contact between the Triassic rocks and the serpentinite. The Triassic sediments are red sandstones and shales; the sandstones often showing rippling which is particularly clear in section where they show-up as small scaled cross strata.

The serpentinite at this point is red stained and forms the southernmost of the dark rocks. It is thrust over the volcanic rocks of the Bennane Head sequence, which youngs towards the south. The trace of the reverse fault is marked by a band of carbonate breccias and mineralisation and it is a thrust of some magnitude: it places mantle lithologies over the superficial sediment of the ocean basin - a depth difference of 10's of km. However the serpentinite may have been emplaced at high levels in several stages; indeed by comparison with present-day marginal basins the serpentinite could have been brought up structurally even at the time of **chert** sedimentation; there are clasts of carbonate rock which contain chrome spinel (Bailey and McCallien 1957).

The dolerite-tuff may have been intruded into the chert at quite high levels as this would account for the development of associated tuffs and breccias. However, the contact between dolerite and chert on the coast does not have the kind of pepperite mixtures of both lithologies which characterises intrusions into wet sediment. This may be due to the nature of the chert or may indicate that the chert was indurated at the time of intrusion.

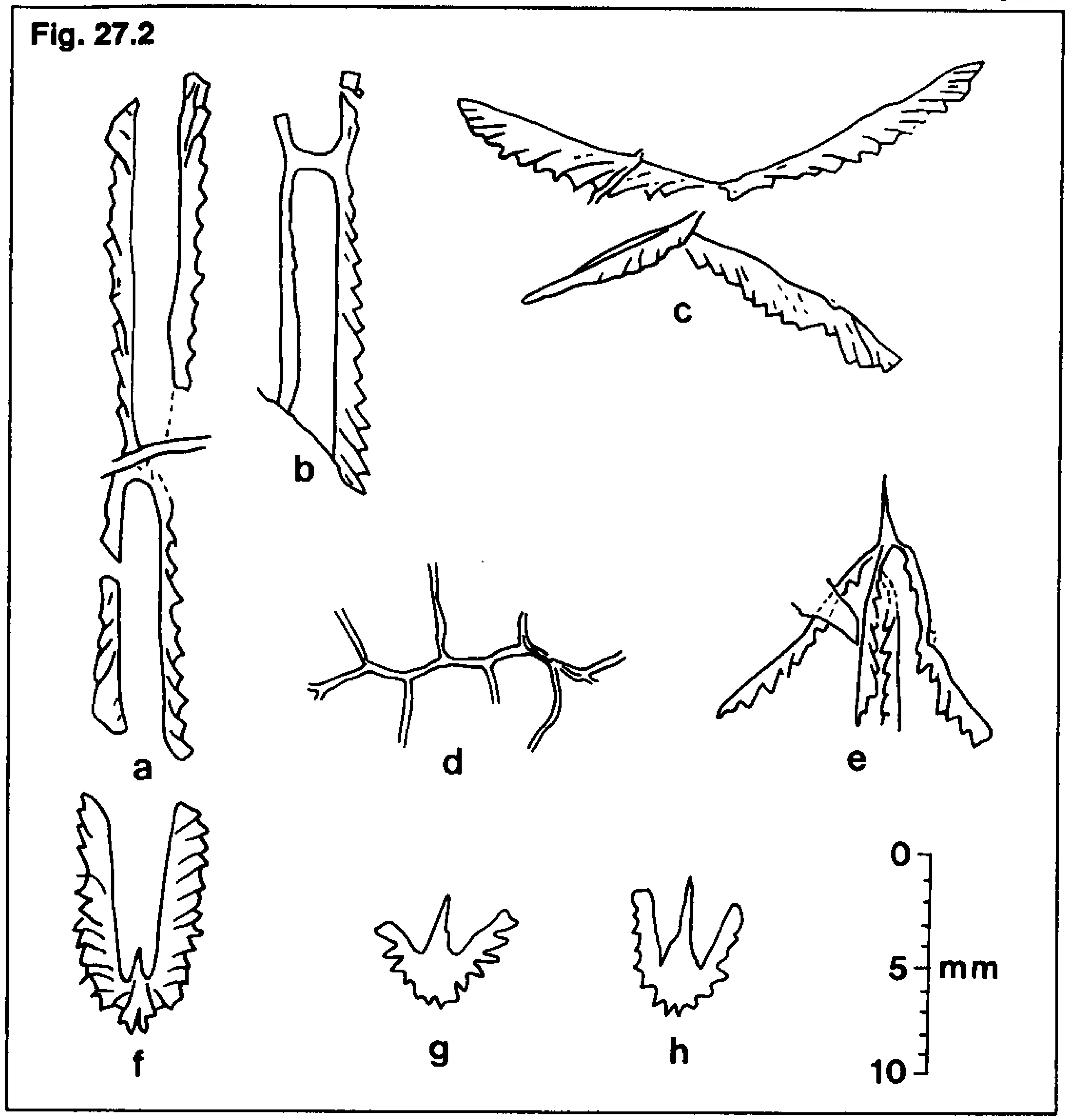


FIGURE 27.2. Graptolites recovered from the Bennane Head sequence by Stone and Rushton (1983). **a,b.** *Tetragraptus approximatus*, **c.** *Tetragraptus reclinatus reclinatus*. **d.** *Sigmagraptus praecursor*, **e.** *Tetragraptus fruticosus*, **f.** *Isograptus caduceus*, **g, h.** *Pseudoisograptus dumosus*

Locality 2. Folded cherts (Fig. 27.3). Below the dolerite-tuff there are red bedded cherts interstratified with coarse-grained, buff-coloured tuffs. Both these lithologies are deformed by a series of slump folds in which the limbs of the folds are thinned and the axes considerably thickened. The tuff bands comprise angular sand sized grains of a variety of volcanic rock fragments including acidic, intermediate and basic, together with some angular clear quartz. The cherts have small circular holes, often filled with fibrous silica, which are sections of

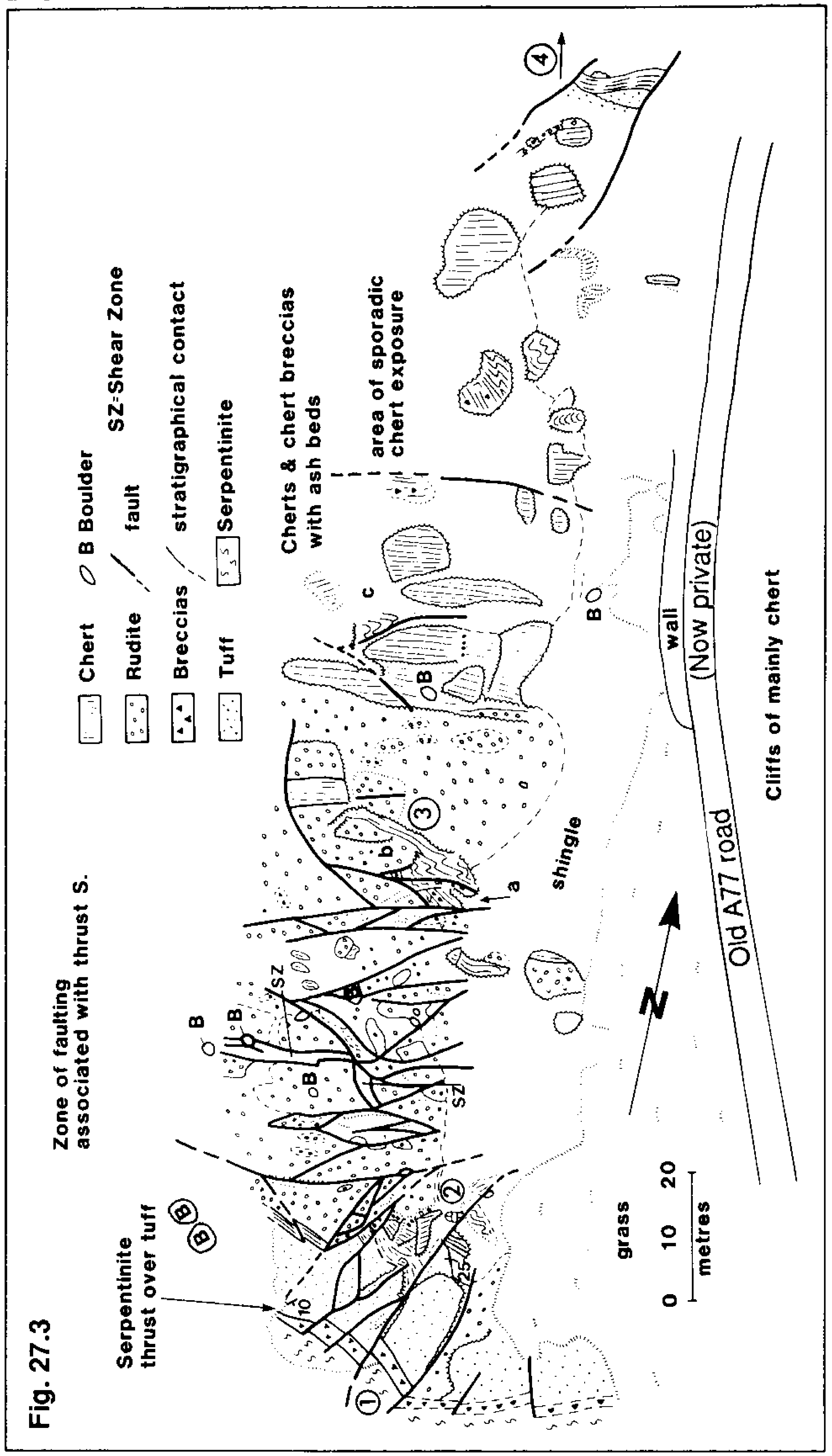


FIGURE 27.3. Detailed plane-table map of the contact beween the serpentinite and the dolerite-tuff and chert; the cherts and conglomerates and associated rocks, N of Bennane Lea.

radiolaria. Some of the coarser cherts have numerous, partly **chloritized** glass shards and angular grains of clear quartz.

The tuff bands show flame-structures, some of which inject the axial planes of the folds; other bands in differing parts of the outcrop here and at Locality 3 show extensional faults cutting them but with the tuffs thinned along the fault planes (Fig. 27.3). All these features attest to the syn-sedimentary origin of the extensional and compressional features and have probably formed at different positions in the slumped sheet (Fig 27.3).

The cherts are followed by breccias and conglomerates which have clasts of volcanic (including acidic) rocks which form the main component, together with a pink coloured, coarse grained igneous rock, limestones of uncertain origin, clasts of dark coloured and red coloured chert. These clasts are often angular but some are very well rounded. They have clear, sharp erosive contacts with the underlying cherts; but the conglomerate-chert contact is often marked by contortion in the cherts and the incorporation of breccias and conglomerates into the slumps.

Locality 3. Deformed cherts and interstratified, interfingering rudites (Fig.27.3). There are a number of important features to examine at this locality. The contact between chert and overlying conglomerate-breccia is marked by abundant deformation (**a**). The conglomerate has no visible bedding and has large rafted blocks of chert within it. The cherts within the blocks are folded, suggesting either that the conglomerate was deposited by a mass flow which displaced some of the soft cherts, or that both the emplacement of the conglomerate and the slumping of the chert were produced by the same mechanism, such as a seismic shock or a grossly oversteepened surface of sedimentation. In any event the tectonic regime in which this sediment accumulated is unstable and close to an abundant supply of coarse grained sediment.

Massive chert is seen to interfinger with the conglomerate (**b**), with once again the disruption of the cherts at the contact. As at locality (**a**) the cherts are overfolded with a vergence towards the east and in this direction the conglomerates of the shore are replaced by the cherts in the raised-beach cliff. Mapping at low-tide and in the shallow waters to the west has shown that there is an almost entire boulder-bearing conglomerate to the west which is replaced by cherts to the east, and the shore section is along the interfingering zone between the two.

At (c) is a beautifully exposed fold, although there may be divergent opinions as to its origin. There are beds in the fold which appear to have been plastically deformed; but the main fold itself may be a late brittle feature.

Locality 4. (NX 0912 8618), **Black and red cherts, black shales and breccias with clasts having a range of compositions** (Fig.27. 2,3). On the rocky foreshore platform, near the most northerly of the high chert stacks, the red chert sequence rests on breccias and green-gray tuffs with abundant volcanic clasts. At this locality (Fig.27.3) there are excellent extensional features, such as **boudinage** in the tuffs and some fine glass shards in some of the cherts. In addition, there are very fine breccias with basic and acidic volcanic fragments; tuffs (**lithic-arenites**) with abundant quartz, feldspar, and basic-acid rock fragments; black cherts and interstratified black shales.

The black shales have yielded graptolites which indicated to Stone and Rushton (1983) a mid-Arenig age. These graptolites include *Tetragraptus fruticosus, Sigmagraptus praecursor, Didymograptus extensus and D.* cf. *protomurchisoni*. The first two graptolites are illustrated in Figure 27.2.

The breccias and lavas are some 300 m thick and can be traced as far as Port Vad, Locality 5, where they rest on spilitic pillow lavas. They represent a mixture of breccias and conglomerates which contain both reddened and dark grey coloured clasts: some are poorly sorted with boulder-sized clasts randomly scattered through them; others are well stratified with rounded clasts in a distinctive white calcite cement. Some beds are dominated by accretionary lapilli enclosed in hyalotuff, and these must represent very shallow-water accumulations. It is clear that this section represents the sediment accumulating on the periphery of a volcanic centre. When the rate of lava eruption was high, basic lavas reached well out into the adjacent sea bed and developed extensive pillow basalts. These may have been accompanied by mass flow deposits comprising volcanogenic breccias; and during periods of intermediate and acidic volcanic activity volcanogenic sediment would have been produced in the marine realm whatever the rate of lava extrusion. At times when the rate of volcanic activity was low, the sediment underwent recycling during a transgression of the sea with the result that well rounded clasts accumulated in well stratified conglomerates (Fig.25.14). The reddened colouration of some of the clasts may well be the result of subaerial weathering (as

discussed in Excursion 25).

Locality 5. Port Vad (NX 0927 8695) **Spilitic pillow lavas and associated tuffs** (Fig. 27.1). Port Vad may be reached (on foot only) from the south after studying localities 1 and 4 by rejoining the private road north of the northernmost gate and walking for less than a kilometre to a point where the road cuts through solid rock. It is then possible to descend, **with great care,** into Port Vad. Coming from the north, park vehicles in the prominent car park (NX 100 875) above Balcreuchan Port and follow the old road southwards to the point above Port Vad. At this locality some excellent exposures of pillows and associated sediments are to be seen. The pillows are relatively undeformed, and show the interpillow carbonate growth typical of Downan Point, Locality 6 and some have tuff within the inter-pillow spaces. At the south side of Port Vad it is possible to trace these pillowed beds into the overlying tuffs and breccias, although there is much faulting here and some splendid examples of sheared pillows.

To the north of this locality there is a thick sequence of interstratified pillows, conglomerate-breccias and tuffs. Stone and Rushton (1983) demonstrated repetition within the lava within this sequence and cautioned against unconditionally assuming the lava sequences to be thick.

Significance of the Bennane Head sequence

The Bennane Head sequence illustrates a number of important features which have relevance to the interpretation of the whole ophiolite. The section from the red cherts through the breccias to the pillow lavas at Port Vad is comparable with a section through the layers 1 and 2 of normal oceanic crust, where sediments cover the lavas produced at the spreading ridge. But there are significant differences:

1. The cherts interfinger with boulder bearing breccias and conglomerates, and the cherts are themselves often highly contorted. This suggests that the cherts formed in or close to a zone of much tectonic instability where the slope of the floor on or near which they accumulated changed radically to allow mass flows to enter the basin and the cherts to slide.

2. There is also much significance in the presence of tuff bands with acid-basic lithologies, layers of crystal tuff and the presence of glass

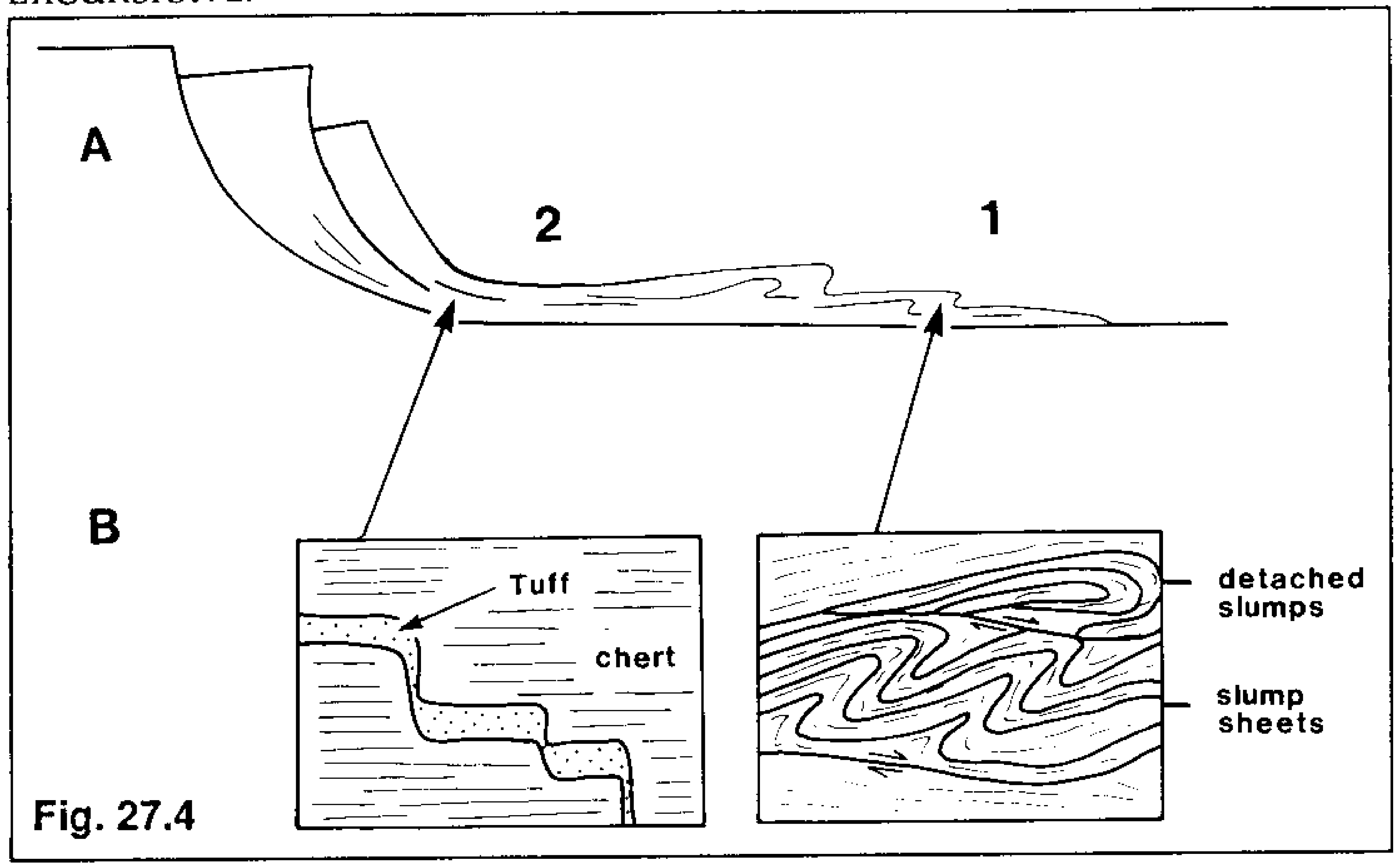


FIGURE 27.4. Diagram showing in **A** the structures typical of the cherts south of Bennane Head. The folds, with associated injection features, were probably formed at position 1; the extensional structures probably formed in position 2. **B** shows the structures which may occur when a large scale slump takes place.

shards within the cherts. These features, together with clasts within the interstratified conglomerates and breccias which are acid to basic in composition, all point to much contemporaneous volcanic activity. Volcanic activity of the kind which produces large volumes of tuff is not a characteristic of most mid-ocean ridges, neither is acidic volcanic activity normally associated with ocean ridges, which are dominantly totally basic in character. This volcanic activity took place during the periods of instability which produced the slumping: the instability was probably caused by faulting and the faulting probably took place in an extensional regime.

3. The great thickness of breccia and conglomerate occurring below the cherts implies that volcanic activity persisted over a considerable period of time. The record here is dominated by rudaceous rocks; lavas form <10% of the record. The presence of well rounded clasts in this sequence also implies shallow water depths and the presence of mass flows, a fairly copious supply of volcanogenic sediment. Indeed the whole sequence through Bennane Head, from Bennane Bridge to

Games Loup, comprises an alternation of breccias, conglomerates, tuffs and lavas. Although faulting and the repetition of strata make estimates of the thickness of this sequence difficult, there is at least 2 km of such lithologies there.

Any hypothesis for the origin of these sequences must therefore account for a thickness of fairly shallow water lava related activity, the presence of contemporaneous acidic volcanic activity towards the top of the pile, general tectonic instability, particularly during the sedimentation of the cherts, and the fairly rapid change in facies from coarse volcanic detritus to fine grained chert.

Of the three possible ways of producing oceanic crust, the sequence at Bennane Head does not resemble the crust of a mid-oceanic ridge, although a possible exception may be the crust found in Iceland which sits astride the Mid-Atlantic ridge. Oceanic islands show a trend from deep water at the base to shallow at the top, and this trend is not seen at Bennane Head. Island arcs, and in particular the basins which lie within or alongside them, have abundant acidic volcanic activity, an abundance of volcanogenic sediments, rapid facies changes from the ground bordering the volcanoes to the deeper basins, and much instability produced not only by the volcanic eruptions but also by the structural fragmentation of the arc during periods of marginal basin formation (see Figure 25.5). It is therefore concluded that the Bennane Head sections represent the basement (Port Vad and the ground to the north) and the rift facies (breccias and cherts of the upper part of the sequence) of a splitting arc. The doleritic intrusive-extrusive unit at the top of the sequence is thought to represent a shallow sill intruded into this extensional regime.

Locality 6. North of Downan Point (NX 0748 8100): **Pillow lavas, showing clear examples of how they are formed, lava caves and the relationship between massive and pillowed lavas** (Fig 27.5). Take the A77 road south from Ballantrae. Shortly after crossing the bridge over the River Stinchar branch off to the right along the road through Garleffin. Cars and minibuses can take the next turn right, near a bungalow, and drive through Kinniegar. Turn right down a rough track c. 300m SW of Kinniegar and drive towards the storm beach. Although the route is private, geologists are at present allowed to open (and close) the gate and park near to the old cottage. Coaches, however, should discharge passengers at the T- junction beyond Garleffin and then proceed south-westwards to the cemetery where

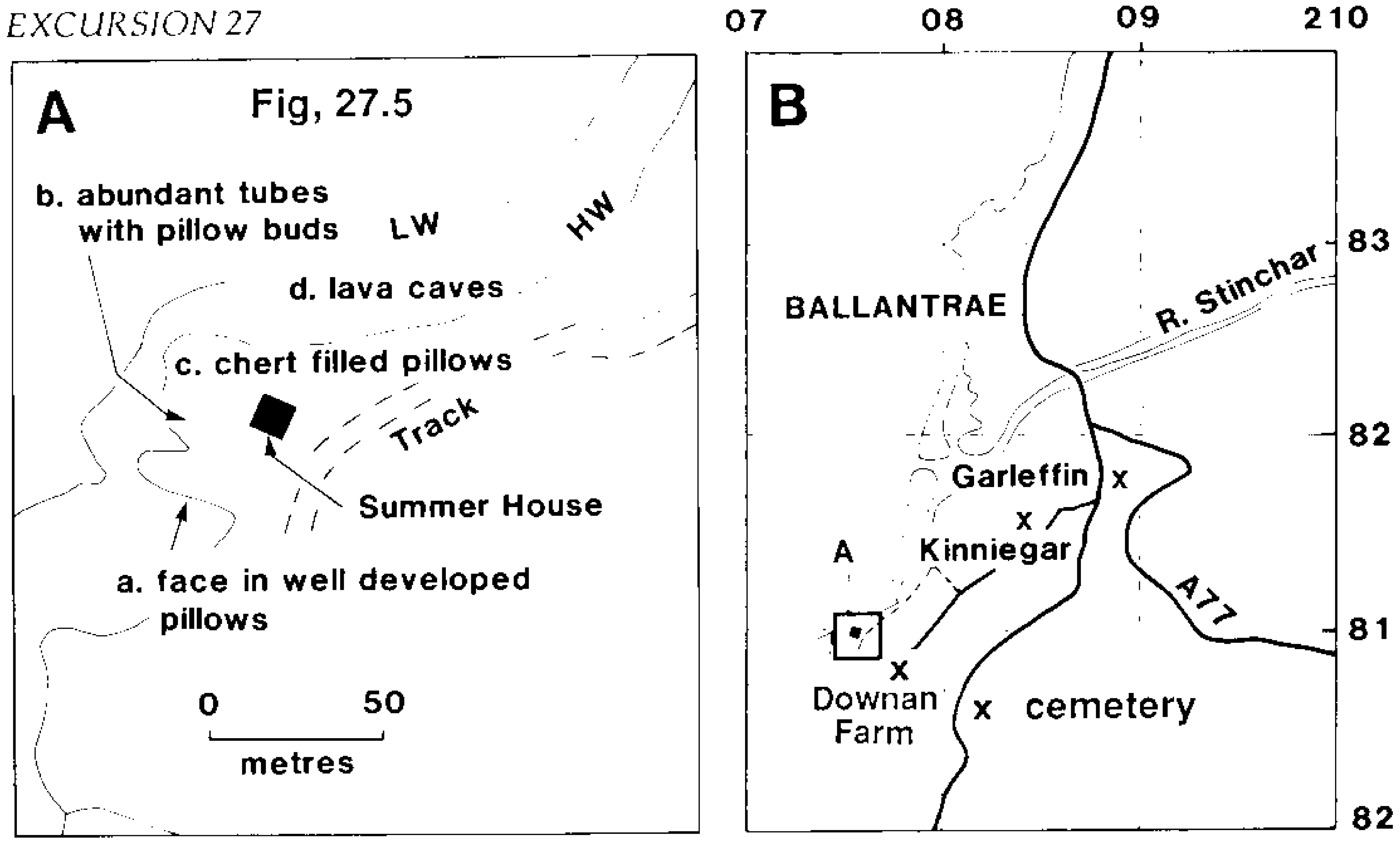


FIGURE 27.5. Maps showing the location of the pillows north of Downan Point, Locality 6. **A**, detailed map, LW, low water and HW high water marks; **B**, general map of the region.

there is ample parking space. From here, the driver can watch the party returning from the shore and meet them at the drop-off point. **Vehicles must not be parked at or near Downan Farm** unless prior written permission has been obtained from Mr. E. MacIntyre, Downan Farm, Ballantrae, Ayrshire KA26 OPB.

Follow the path on foot to the headland c.100 m to the south and stop at the gravel beach south of the stream (Fig.27.5). The north face of the headland is a magnificent exposures of pillow lavas and their geometry can be seen in some detail.

Pillow lavas are magnificently exposed along this coastline to as far south as Dove Cove. The age of these pillows has been the subject of some controversy. Lewis and Bloxam (1977) thought they were Caradoc in age. Thirlwall and Bluck (1984) obtained an age of 468 Ma with a rather large error of 22Ma, which taken at face value would be roughly Caradoc. The chemistry of these lavas has been studied by Wilkinson and Cann (1974), Lewis and Bloxam (1977) and Thirlwall and Bluck (1984) all of whom found that they resembled the chemistry of ocean islands.

On the intertidal, well washed exposures it is possible to see some fine details of the pillows. They have green chloritic rims which are the result of the alteration of glassy chilled margins (Fig.27.6 A **a**). Young

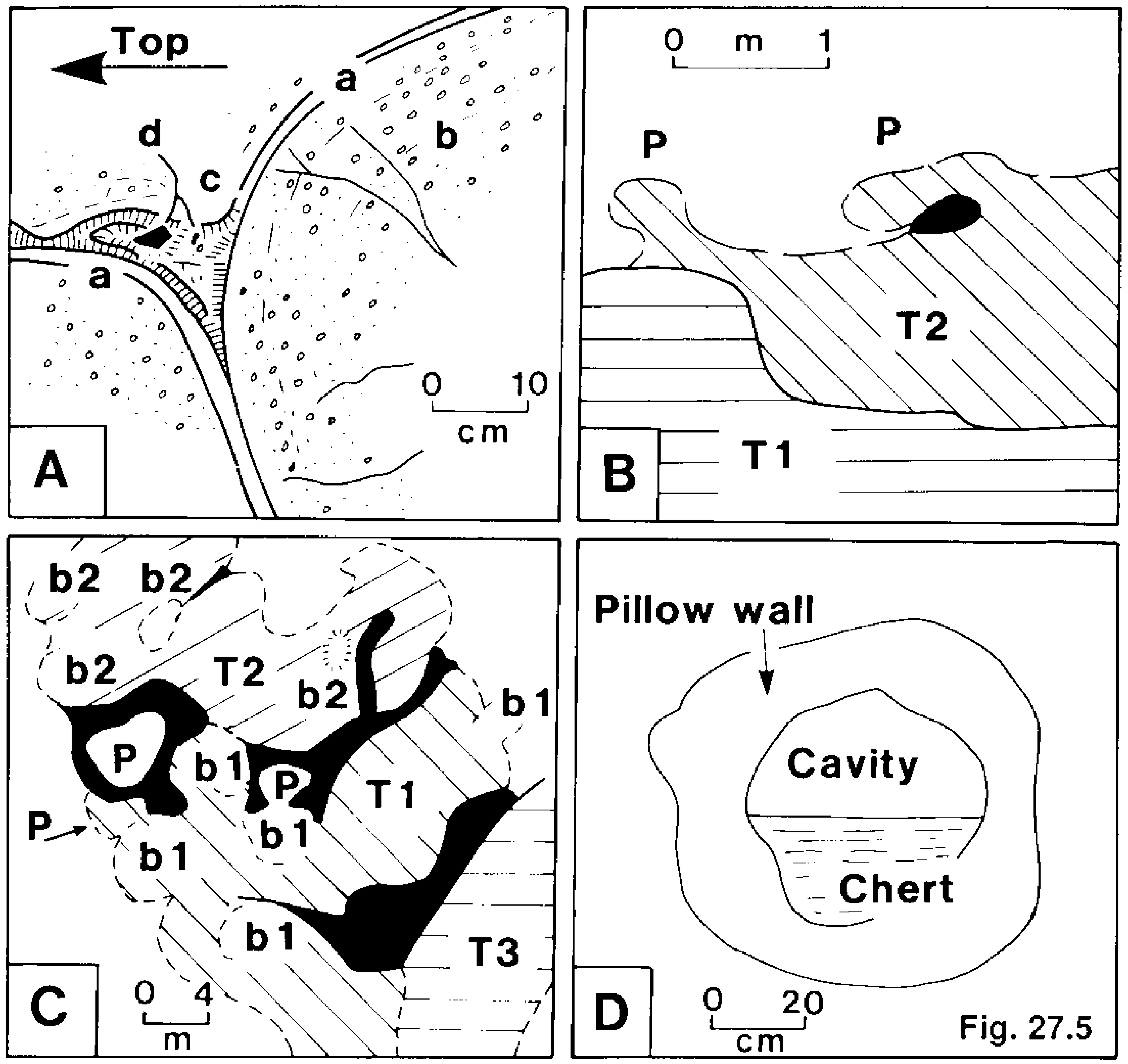


FIGURE 27.6 **A**, sketch showing the details of some of the pillows. The outer surface (**a**) is chilled, and can rarely be seen as a glass; mostly it is glass which has been almost totally replaced by chlorite, thus giving many of the pillows a green external colour. The interior of the pillow often has vesicles (**b**) which may be in radiating trails or in concentric lines which are preferentially developed towards the top of the pillows. The spaces beween the pillows are often filled with a fibrous calcite (**c**) and the centre of the pores may be filled with chert (**d**). **B**, illustrates two tubes T1 and T2 in section, with the lavas flowing from the right. T1 has provided the relief over which T2 has flowed; P are the two pillow-buds (incipient pillows) developed on the top of T2, and one pillow probably flopped over after it was inflated. The dark area is unfilled space. **C** is a plan sketch of the base of a sequence of pillow lavas. T1, T2 and T3 are the bases of three tubes. T1, 2 in particular have grown many pillow-buds b1 and b2, and some pillows (P). The dark areas are pore-space fillings which occur between the tubes and pillows. **D,** a section through a pillow which, when inflated did not fill with lava. It was subsequently partly filled with chert.

pillows wrap around older giving a clear indication of younging to the west, and the abundant vesicles, sometimes arranged concentrically with the pillow outline, are concentrated near their the top margins (Fig. 27.6 A **b**). Interpillow spaces are filled with calcite, and only rarely do these show a geopetal infilling (Fig. 27.6 A **c**).

Moving to the north, and on the north side of a small indentation of the sea, (**b**), Figure 27.5 A, there are thin bands of dark chert and tuff amongst the pillows showing a clear NE-SW strike. At this locality it is possible to see a whole range of important attributes of this sequence. Although in section the pillows have a roughly circular outline, examination of faces which are in the strike of the beds reveal them to range in shape from long tubes with many pillow buds on them (Fig.27.6 B **T1,T2**) to irregular flat bottomed masses with many tubes and buds coming off them (Fig.27.6 B **P**;27.6 C). It is readily apparent that the first face visited showing the classical pillow shapes gives a false impression of their real shape.

Some pillows at this locality have hollow centres now partly filled with stratified chert (Fig. 27.6 D). These must have been inflated with lava and then drained to leave a space subsequently to be filled with the chert.

From these localities it is possible to speculate on the origin of the pillow pile. The almost spherical pillows are produced from the buds which appear on the lava tubes. These buds grow by the lava pressure forcing out the tube wall at points of weakness or points where there is a greater pressure in the tube. They inflate to a size where they are unstable and then roll off. The lava tubes grow along the floor, sometimes climbing over pre-existing tubes and sometimes filling the inter-tube spaces, all the time budding and breaking up into separate tubes. In this way the pillow lava pile grows upwards.

A little to the north, at (**d**), (Fig. 27.5), there are larger lava caves, where up to 2m wide holes in the lava, now partly filled with stratified lava and tuff, probably represent the routeways for lava through horizontal conduits in the lava pile. There is much alteration along these small lava caves and it is probable that much gas escaped along here during the later degassing phases of the lava pile.

There are numerous exposures of pillow lavas and massive spilite to the south of this locality, and many of these are worth examination for the insights they give into the growth of lava piles of this type. However some 200m to the south of locality 6 there are excellent

exposures showing the development of pillow tubes off a massive lava (see Fig. 25.13). In this instance the rate of lava extrusion was so great that it did not break down into tubes except at the edges where it was easily chilled. The relationships between the various types of pillows, tubes and massive lavas are illustrated in Figure 25.13.

The lava flows of Downan Point differ from those studied at Pinbain (Excursion 25) and at Bennane Head (Excursion 26) in having comparatively little volcanogenic clastic rock present. This implies that the environment of extrusion was different and an obvious factor in this respect is the water depth. The Downan point lavas were probably extruded in deeper water than most of the lavas in the Ballantrae Complex. They may well be much younger than the Ballantrae Lavas, and many geologists now put the Southern Uplands Fault along the river Stinchar, so making Downan Point part of the Southern Uplands.

References are given after Excursion 31.

THE ORIGIN OF THE BALLANTRAE COMPLEX
Conclusions for excursions 25-27
B.J.Bluck

The origin of the Ballantrae Complex has been debated by several workers (Dewey 1974; Church and Gayer 1973; Bluck *et al*, 1980; Barrett *et al*, 1981; Stone 1984). There are several lines of evidence which are critical to the understanding of the Ballantrae Complex and solving the problem of its origin:

1. The nature of the sediments which make up the top layer of the **ophiolite**, equivalent to layer 1 of normal ocean crust, as produced in the large ocean basins. These are seen at Pinbain (Excursion 25) and Bennane Head (Excursion 26): and the make-up of Bennane Head in particular is significant in that it differs from normal layer 1 of the crust in the following ways:

(a). Although the cherts and black shales are well bedded, both contain the remnants of volcanic activity in the form of glass shards and lithic fragments of acidic and intermediate volcanic rock. Sometimes thin breccias of volcanogenic origin are interstratified with the cherts. All of this implies that the basin in which the chert formed was close to an active acidic volcanic complex and received some of the ash blown out by the eruptions. Ash would be incorporated into layer 1 of the deep oceans at hot spots which produce ocean islands, or where the mid-ocean spreading ridge has grown subaerially, but in these instances most of the ash would be basic in composition.

(b). The cherts and black shales are very commonly slumped i.e. they have been deformed by submarine sliding when they were still unconsolidated. This repeated slumping, seen at Benanne Head and Pinbain, implies deposition on either some reasonably steep slope or a shallow slope which was affected by tectonic activity, setting the sediment pile in motion. This is not a characteristic feature of deep oceanic sediment: slopes are normally very shallow in the deep abyssal plains where chert sequences of this thickness are normally found. However, in the deep ocean steep slopes and instability may be generated near ridges and transform faults.

(c). The cherts and black shales are interbedded with conglomerates and breccias, some of which are boulder-bearing. Some of the clasts

are well rounded and may have spent some time in fluvial transportation or have been abraded by wave activity whilst in shallow coastal regimes. When the whole sequence is mapped out both at Bennane Head and Pinbain, it can be clearly seen that the cherts and black shales pass rapidly into laterally equivalent thick sequences of rudite. This type of relationship is not characteristic of normal oceanic crust, is not usual at spreading ridges and is uncommon at transform faults. It is, in contrast, found in basins produced by arc splitting where new ocean crust is being formed. Here there are many normal faults over which substantial piles of arc-derived (acid-basic) sediments are draped. Steep surfaces upon which sediment accumulates are common, and many are rendered unstable by the continual fault activity which extends the basin. During this time there is extensive deformation of newly laid sediment.

2. The characteristics of the lavas equivalent to layer 2. There are a number of important points here:

(a). Layer 2 of oceanic crust produced at the major ocean ridges is characterized by an abundance of massive and pillowed lava; breccias and conglomerates account for <10% of the rock record. The presence of abundant breccias and tuffs throughout these lava sequences is evidence of explosive eruption of the lavas, and that in turn implies that they were continuously erupted in shallow water.

(b). A shallow water origin is also demonstrated by the presence of well-rounded lava clasts (Excursion 27); hyaloclastic deltas (which are intertidal, Excursion 25) and subaerial lava flows with reddened tops. The repetition of such features through the stratigraphic column suggests subsidence *during lava extrusion* which is not a common feature of ocean ridges or oceanic islands. Lavas created at oceanic ridges generally undergo subsidence after extrusion has stopped and when the lithosphere is cooling and moving away from the ridge. Ocean islands on the other hand build up from deep water to shallow, so the base of the pile should be dominated by massive and pillowed flows and the top by shallow water volcanic activity.

(c). Although there is repetition by faulting which tends apparently to thicken the section, the lavas are nevertheless very thick. Normal oceanic crust is < 1.5 km thick, and oceanic islands can produce extremely thick lavas.

(d). Breccias and lava flows of intermediate-acid type occur at the top of both the Pinbain and Bennane Head lava sequences. This type of lava is not typical of oceanic spreading centres, but is commonly found in magmatic arcs.

(e). Much work has been done on present-day oceanic lavas with a view to correlating their chemistry with the tectonic setting of their extrusion. For present-day basaltic lavas major element discriminant plots are very useful in indicating the tectonic settings in the oceanic realm. When these techniques are applied to the lavas at Ballantrae they fall into a wide variety of geochemical fields typical of oceanic islands (hot-spots), marginal basins, oceanic ridges and island arcs (Wilkinson and Cann 1974; Thirlwall and Bluck 1984). Although the validity of using these geochemical plots for ancient oceanic crust has been questioned, for recent environments of ocean crust formation those basalts formed in island arcs and marginal basins show the greatest degree of diversity in chemistry.

3. Evidence from the age dating and the obduction.

It is clear from the age determinations carried out on the ophiolite that it is of Ordovician age and was generated and obducted within the Arenig epoch. It was therefore not old crust at the time of obduction but young. Young crust tends to be thin and hotter than old, yet the metamorphic rocks at the sole of the ophiolite are required to have been buried by at least 30 km of rock (see Excursion 26). If the ophiolite were to be an arc-marginal basin assemblage which has collided with a subduction zone, then the thickness of the crust, its age and the age of obduction can easily be accounted for as shown in Figure 26.3.

Ophiolites with similar characteristics and roughly similar age to the Ballantrae Complex have been found in Newfoundland and SW along the Appalachians (Dunning and Krogh 1985). Ophiolites of this age have also been found in Scandinavia, and their widespread occurrence at this time suggests a long destructive margin to the northern continent of Laurentia (see Introduction). In most of these instances a marginal basin origin has been proposed for the ophiolites, and we may therefore infer that the oceanic crust being consumed at that time was quite old.

There is evidence within the Ballantrae Complex and elsewhere for a long history of subduction. There is a meta-pyroxenite block within the mélange at Knockormal which has yielded an age of 576 +- 32 Ma. On most time-scales this date is Cambrian, and the composition of the

block suggests its **protolith** to be basalt. If this is so, then it is probable that oceanic crust, older than 576 Ma was somewhere being consumed. Of course this event may not have taken place at Ballantrae; the block could have been tectonically transported from some distance. However, in the dating of island arc type andesitic and rhyolitic fragments from the Southern Uplands, Kelley and Bluck, (1989) found them to have ages of 560 Ma and 630 Ma respectively— again pointing to the presence of Cambrian subduction. Dempster and Bluck (1991) have shown that an ophiolite along the Highland Border zone was thrust onto the continent at about 537 Ma., and they further discuss the implications of these ages.

References for Excursions 25 to 31

BAILEY, E.B., and McCALLIEN, W.J. 1957. The Ballantrae serpentinite, Ayrshire. *Trans. Edinb. Geol. Soc.* **17**, 33-53.

BALSILLIE, D. 1932. The Ballantrae Igneous complex, south Ayrshire. *Geol. Mag.* **69**, 107-131.

BERGSTROM, S. M. 1990. Biostratigraphic and biogeographic significance of Middle and Upper Ordovician conodonts in the Girvan succession, south-west Scotland. *Courier Forsch.-Inst. Senckenberg.*, **118**, 1-43, 4 pls.

BLOXAM T.W and ALLEN, J.B. 1960. Glaucophane schist, eclogite and associated rocks from Knockormal in the Girvan-Ballantrae Complex, south Ayrshire. *Trans. R. Soc. Edinb.* **64**, 1-27.

BLUCK, B.J. 1982. Hyalotuff deltaic deposits in the Ballantrae ophiolite of S.W. Scotland: evidence for the crustal position of the lava sequence. *Trans R. Soc. Edinb. Earth Sci.* **72**, 217-228.

_______ 1983. Rôle of the Midland Valley of Scotland in the Caledonian Orogeny. *Trans. R. Soc. Edinburgh: Earth Sciences*, **74**, 119-136.

_______ 1985. The Scottish paratectonic Caledonides. *Scott .J. Geol.* **21**, 437-464.

_______ and HALLIDAY, A.N. 1981. Comment and reply on age and origin of the Ballantrae ophiolite and its significance to the Caledonian Orogeny and the Ordovician time-scale. *Geology*, **10**, 331-333.

_______ HALLIDAY, A.N., AFTALION, M., and MACINTYRE, R.M. 1980. Age and origin of the Ballantrae Complex and its significance to the Caledonian orogeny and the Ordovician time scale . *Geology* **8**, 492-495.

CHATTERTON, B. D. E and LUDVIGSEN, R. 1976. Silicified Middle Ordovician trilobites from the South Nahanni River area, District of Mackenzie, Canada. *Palaeontogr., Abt. A*, **167**, 77-119, pls 9-24.

CHURCH, W.R. and GAYER, R.A. 1973. The Ballantrae ophiolite. *Geol. Mag.* **110**, 497-510.

COCKS, L. R. M. and TOGHILL, P. 1973. The biostratigraphy of the Silurian rocks of the Girvan district, Scotland. *J. Geol. Soc. London*, **129**, 209-243.

DEMPSTER, T.J. and BLUCK, B.J. 1991. The age and tectonic significance of the Bute amphibolite, Highland Border Complex, Scotland. *Geol. Mag.* **128**, 77-80.

DEWEY, J.F. 1969. Evolution of the Appalachian/Caledonian orogen. *Nature* **222**, 124-129.

_____ 1974. The geology of the southern terminations of the Caledonides, *in* Nairn, A, (ed) *The ocean basins and their margins. Vol. 2 The North Atlantic.* 205-231. New York

DONOVAN, S. K. and CLARK, N. D. L. 1992. Unusual crinoid columnals from the Llandovery of England and Wales. *Palaeontology* **35** (in press).

DUNNING, G.R. and KROGH, T.E. 1985. Geochronology of ophiolites of the Newfoundland Appalachians. *Canadian Journ. Earth. Sci.* **22**, 1659-1670.

HARPER, D. A. T. 1981. The stratigraphy and faunas of the Upper Ordovician

High Mains Formation of the Girvan district. *Scott. J. Geol.*, **17**, 247-255.

_______ 1982. The stratigraphy of the Drummuck Group (Ashgill), Girvan. *Geol. J.*, **17**, 251-277.

_______ 1984. Brachiopods from the upper Ardmillan succession (Ordovician) of the Girvan district, Scotland. Part I. *Palaeontogr. Soc. [Monogr.], London.* 1-78, pls 1-11.

_______ 1989. Brachiopods from the upper Ardmillan succession (Ordovician) of the Girvan district, Scotland. Part 2. *Palaeontogr. Soc. [Monògr.], London.* 79-128, pls 12-22.

HENDERSON, S. M. K. 1935. Ordovician submarine disturbances in the Girvan district. *Trans. R. Soc. Edinburgh*, **58**, 487-509, 4 pls.

HOWELLS, Y. 1982. Scottish Silurian trilobites. *Palaeontogr. Soc. [Monogr.], London.* 1-76, 15 pls.

HOLUB, F.V., KLAPOVA, H. BLUCK, B.J. and BOWES, D.R. 1984. Petrology and geochemistry of post-obduction dykes of the Ballantrae complex, SW Scotland. *Trans R.Soc. Edinb. Earth Sci.* **75**, 211-223.

HUGHES, C. P., INGHAM, J. K. and ADDISON, R. 1975. The morphology, classification and evolution of the Trinucleidae (Trilobita). *Phil. Trans. R. Soc. London, B*, **272**, 537-604.

INCE, D.M. 1984. Sedimentation and tectonism in the middle Ordovician of the Girvan district, SW Scotland. *Trans R.Soc.Edinb. Earth Sci.* **75**, 225-237.

INGHAM J. K. 1968. British and Swedish Ordovician species of *Cybeloides* (Trilobita). *Scott. J. Geol.*, **4**, 300-316, 2 pls.

_______ 1974. The Upper Ordovician trilobites from the Cautley and Dent districts of Westmorland and Yorkshire . Part 2. *Palaeontogr. Soc. [monogr.]* ,59-87, pls 10-18.

_______ 1978. Geology of a continental margin 2: middle and late Ordovician transgression, Girvan. *In* Bowes D. R. and Leake, B. E. (eds) Crustal evolution in northwestern Britain and adjacent regions. *Geol. J. Special Issue* **10**, 163-176.

_______ , CURRY, G. B. and Williams, A. 1986. Early Ordovician Dounans Limestone fauna, Highland Border Complex, Scotland. *Trans. R. Soc. Edinburgh: Earth Sciences*, **76** (4) (for 1985, publ. Feb.1986). 481-513.

_______ and TRIPP, R. P. 1991. The trilobite fauna of the Middle Ordovician Doularg Formation of the Girvan district, Scotland, and its palaeoenvironmental significance. *Trans. R. Soc. Edinburgh: Earth Sciences*, **82**, 27-54.

KELLEY, S. and BLUCK, B.J. 1989. Detrital mineral ages from the Southern Uplands using ^{40}A - ^{39}A laser probe. *J. Geol. Soc. Lond.* **146**, 401-404.

KIELAN - JAWOROWSKA, Z., BERGSTROM, J. and AHLBERG, P. 1991. Cheirurina (Trilobita) from the Upper Ordovician of Vastergotland and other regions of Sweden. *Geol. Foren. ; Stockholm Forhandl.*, **113**, 219-244.

LANE, P. D. 1971. British Cheiruridae (Trilobita). *Palaeontogr. Soc. [Monogr.], London.* 1-95, 16 pls.

LEWIS, A.D and BLOXAM, T.W. 1977. Petrotectonic environments of the Girvan-Ballantrae lavas from rare-earth element distributions. *Scott. J. Geol.* **13**,

211-222.
LONGMAN, C.D., BLUCK, B.J. and VAN BREEMEN, O. 1979. Ordovician conglomerates and the evolution of the Midland Valley. *Nature*, **280**, 578-581.
LUDVIGSEN, R. 1978. Middle Ordovician trilobite biofacies, southern Mackenzie Mountains. *In* Stelck, C. R. and Chatterton, B. D. E. [eds] Western and Arctic Canadian biostratigraphy. *Geol Assoc. Can., Spec. Pap.* **18.**
OWEN, A. W. 1986. The uppermost (Hirnantian) trilobites of Girvan, SW Scotland with a review of coeval trilobite faunas. *Trans. R. Soc. Edinburgh: Earth Sciences*, **77**, 231-239.
OWENS, R. M. 1973. British Ordovician and Silurian Proetidae (Trilobita). *Palaeontogr. Soc. [Monogr.], London* . 1-98, 15 pls.
PEACH, B.N. and HORNE, J. 1899. The Silurian rocks of Britain,1: Scotland *Mem Geol.Surv. U.K.*
REED, F. R. C. 1903-6. The Lower Palaeozoic trilobites of the Girvan district, Ayrshire. *Palaeontogr. Soc. [Monogr.], London.* Pt 1 (1903), 1-48, pls 1-6, Pt 2 (1904), 49-96, pls 7-13, Pt 3 (1906), 97-186, pls 14-20.
________ 1914. The Lower Palaeozoic trilobites of Girvan. Supplement. *Palaeontogr. Soc. [Monogr.], London.* 1-56, 8 pls.
________ 1917. The Ordovician and Silurian brachiopods of the Girvan district. *Trans. R. Soc. Edinburgh,* **51**, 795-998, 24 pls.
________ 1931. The Lower Palaeozoic trilobites of Girvan. Supplement No 2. *Palaeontogr. Soc. [Monogr.], London.* 1-30.
________ 1935. The Lower Palaeozoic trilobites of Girvan. Supplement No 3. *Palaeontogr. Soc. [Monogr.], London.* 1-64, 4 pls.
RUSHTON, A. W. A. and TRIPP,R. P. 1979. A fossiliferous lower Canadian (Tremadoc) boulder from the Benan Conglomerate of the Girvan district. *Scott. J. Geol.*, **15**, 321-327.
________, STONE, P., SMELLIE, J.L. and TUNNICLIFF, S.P. 1986. An early Arenig age for the Pinbain sequence of the Ballantrae Complex. *Scott. J. Geol.* **22**, 41-54.
SMELLIE, J.L. 1984. Accretionary lapilli and highly vesiculated pumice in the Ballantrae ophiolite complex: ashfall products from subaerial eruptions. *Rep.Br. Geol. Surv.* **16**, 36-40.
SPRAY, J.G. and WILLIAMS, G.D. 1980. The sub-ophiolite metamorphic rocks of the Ballantrae igneous complex, SW Scotland. *J.Geol. Soc.Lond.* **137**, 359-368.
STONE, P. 1984. Constraints on genetic models for the Ballantrae complex, SW Scotland. *Trans. R.Soc.Edinb. Earth Sci.* **75**, 189-191.
________, and RUSHTON, A.W.A. 1983. Graptolite faunas from the Ballantrae ophiolite complex and their structural implications. *Scott. J. Geol.* **19** 297-310.
________, and SMELLIE, J.L. 1988. *Classical areas of British geology: The Ballantrae area: a description of the solid geology of parts of 1:25 000 sheets NX 08, 18 and 19.* (London: HMSO for British Geological Survey.) B.G.S.
THIRLWALL, M.F. and BLUCK, B.J. 1984. Sr-Nd isotope and geological evidence that the Ballantrae "ophiolite", SW Scotland is polygenetic. in Gass, I.G., Lippard, S.J., and Shelton, A.W. (eds) *Ophiolites and oceanic lithosphere. Spec. Pub. Geol. Soc. Lond.* No **13**, 215-230.

TOGHILL, P. 1970. Highest Ordovician (Hartfell Shales) graptolite faunas from the Moffatarea, south Scotland. *Bull. Brit. Mus. nat. Hist. (Geol.)*, **19**, 1-26, 16 pls.
TRELOAR, P.J., BLUCK,B.J., BOWES, D.R. and DUDEK,A. 1980. Hornblende-garnet metapyroxenite beneath serpentinite in the Ballantrae complex of S.W.Scotland, and its bearing on the depth of provenance of obducted ocean lithosphere. *Trans. R. Soc. Edinb. Earth Sci.* **71**, 201-212.
TRENCH, A., BLUCK, B. J. and WATTS D. R. 1988. Palaeomagnetic studies within the Ballantrae Ophiolite; southwest Scotland: magnetotectonic and regional tectonic implications. *Earth and Planetary Science Letters*, **90**, 431-448.
TRIPP, R. P. 1954. Caradocian trilobites from mudstones at Craighead Quarry, near Girvan, Ayrshire. *Trans. R. Soc. Edinburgh*, **62**, 655-693, 4 pls.
________ 1962. Trilobites from the *confinis* Flags (Ordovician) of the Girvan district, Ayrshire. *Trans. R. Soc. Edinburgh*, **65**, 1-40, 4 pls.
________ 1965. Trilobites of the Albany division (Ordovician) of the Girvan district, Ayrshire. *Palaeontology*, **8**, 577-603, pls 80-83.
________ 1967. Trilobites of the upper Stinchar Limestone (Ordovician) of the Girvan district, Ayrshire. *Trans. R. Soc. Edinburgh*, **67**, 43-93, 6 pls.
________ 1976. Trilobites from the basal *superstes* Mudstones (Ordovician) at Aldons Quarry, near Girvan, Ayrshire. *Trans. R. Soc. Edinburgh*, **69**, 369-423, 7 pls.
________ 1979. Trilobites from the Ordovician Auchensoul and Stinchar Limestones of the Girvan district, Strathclyde. *Palaeontology*, **22**, 37-40, pls 37-40.
________ 1980a. Trilobites from the Ordovician Balclatchie and lower Ardwell groups of the Girvan district, Scotland. *Trans. R. Soc. Edinburgh: Earth Sciences*, **71**, 147-157, 1 pl.
________ 1980b. Trilobites from the Ordovician Ardwell Group of the Craighead Inlier, Girvan district, Scotland. *Trans. R. Soc. Edinburgh: Earth Sciences*, **71**, 123-145, 4 pls.
WILLIAMS, A. 1959. A structural history of the Girvan district,SW Ayrshire. *Trans. R. Soc. Edinburgh*, **63**, 629-667.
________ 1962. The Barr and lower Ardmillan Series (Caradoc) of the Girvan District, south-west Ayshire, with description of the Brachiopoda. *Mem. Geol. Soc. London*, **3**, 1-267, 25 pls.
WILLIAMS, S. H. 1987. Upper Ordovician graptolites from the *D. complanatus* Zone of the Moffat and Girvan districts and their significance for correlation. *Scott. J. Geol.* **23**, 65-92.
WILKINSON, J.M. and CANN, J.R. 1974. Trace elements and tectonic relationships of basaltic rocks in the Ballantrae igneous complex, Ayrshire. *Geol Mag.* **111**, 35-41.