

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.

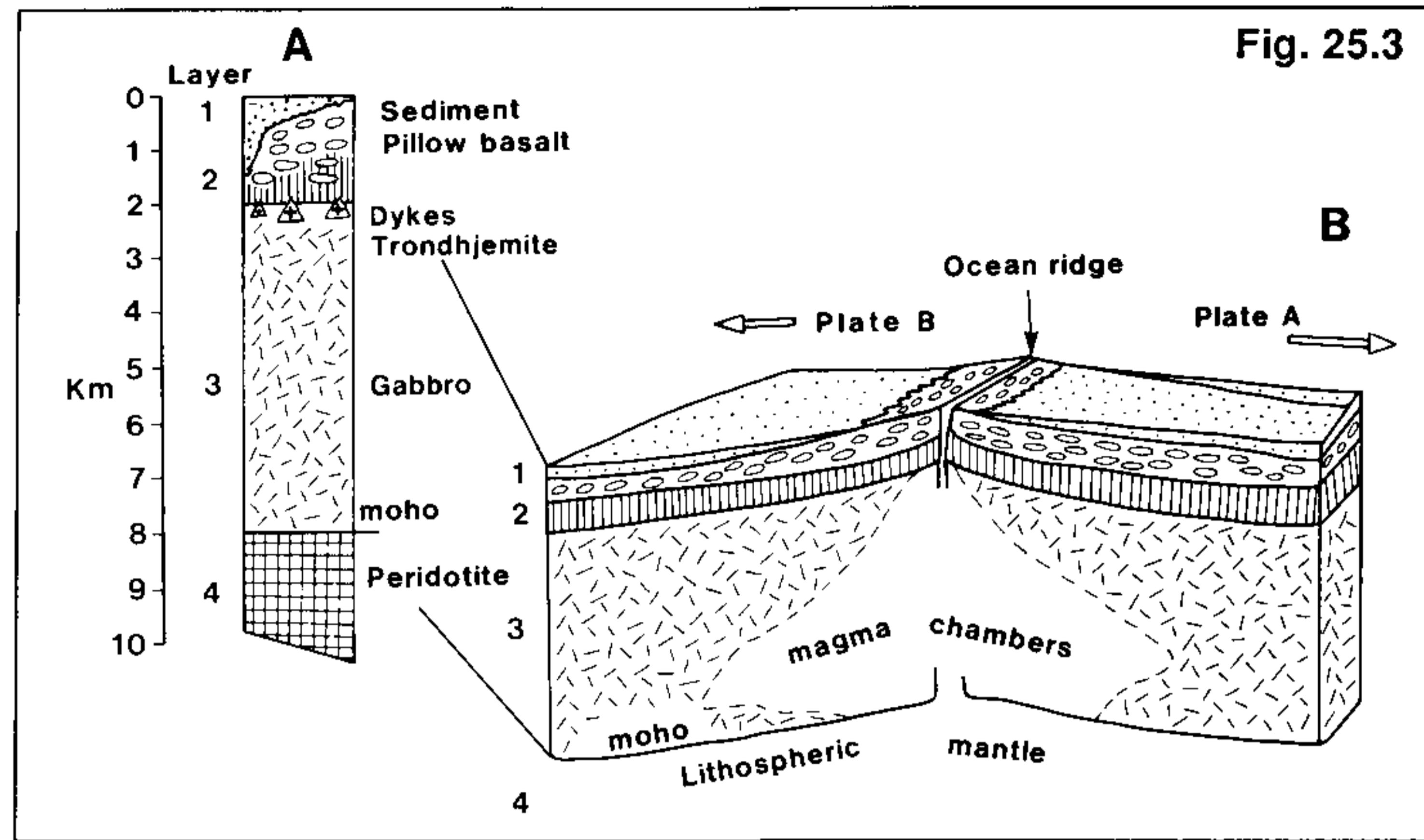


Fig. 25.3

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,

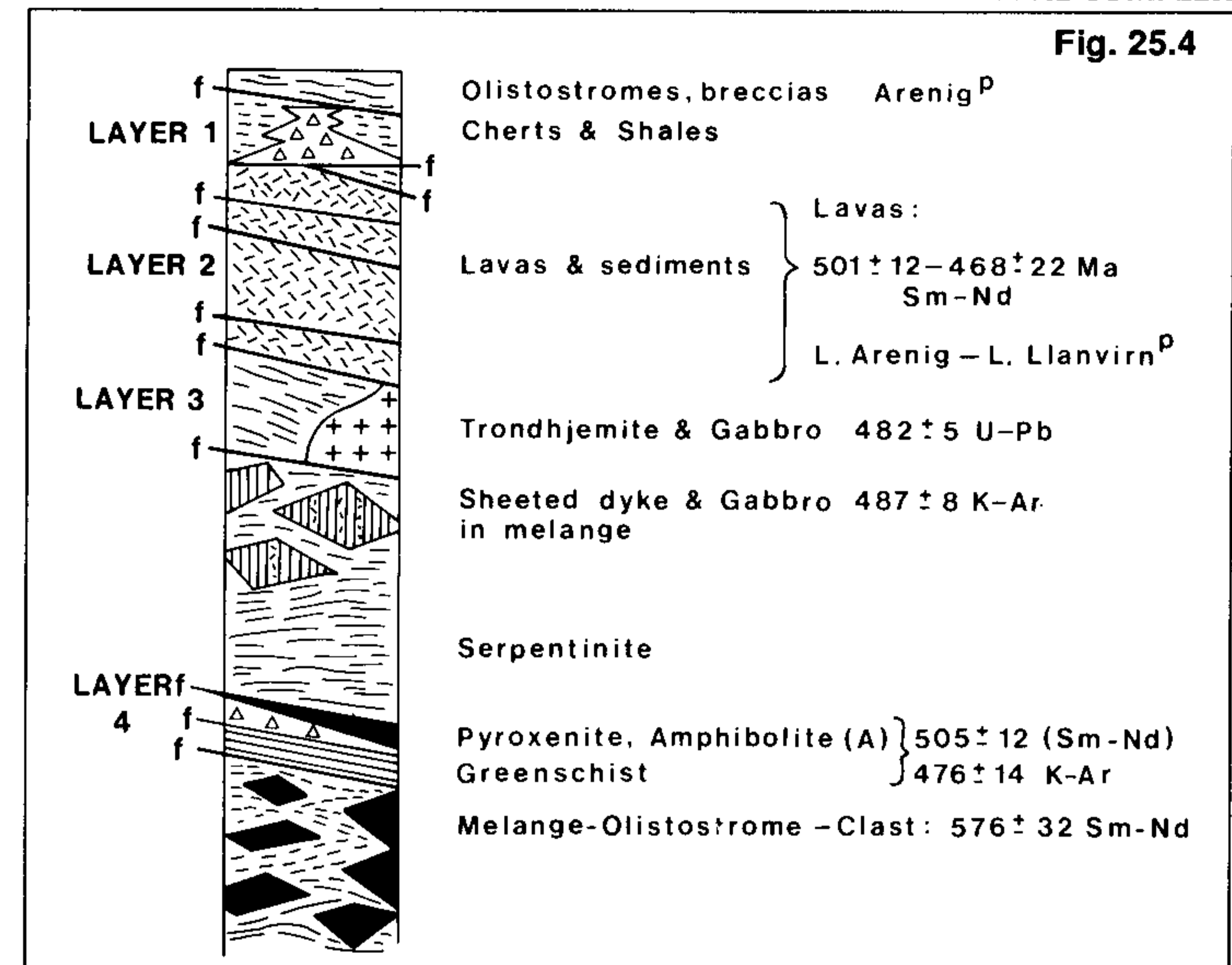


Fig. 25.4

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 is 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

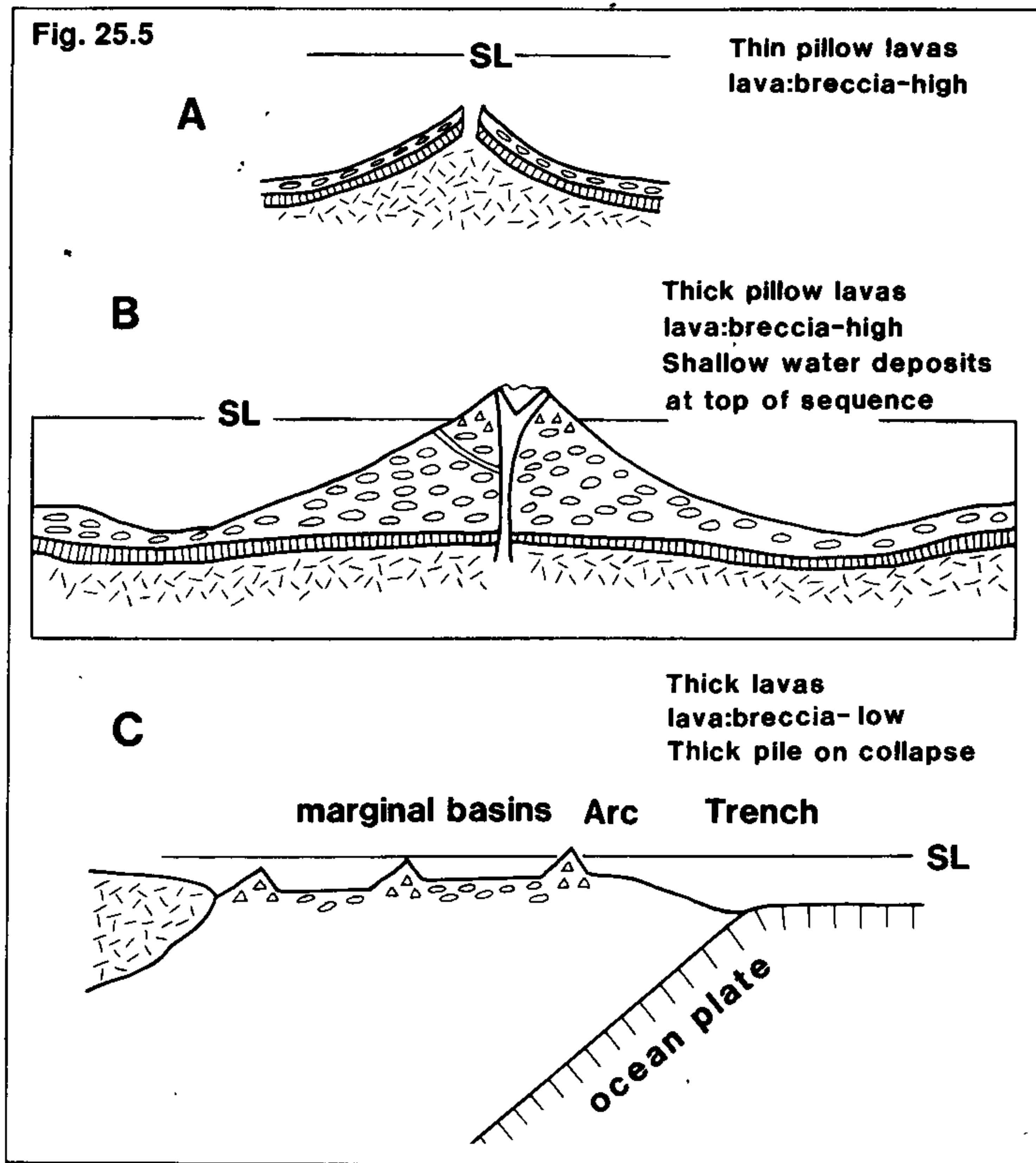


FIGURE 25.5. Diagram showing the various ways in which ophiolites form, together with some of the main characteristics which typifies each one. **A.** Formation at a spreading ridge; **B.** At an ocean seamount, **C.** At a marginal basin-arc. SL = sea level. The symbols are as for Figures 25.3 and 25.4.

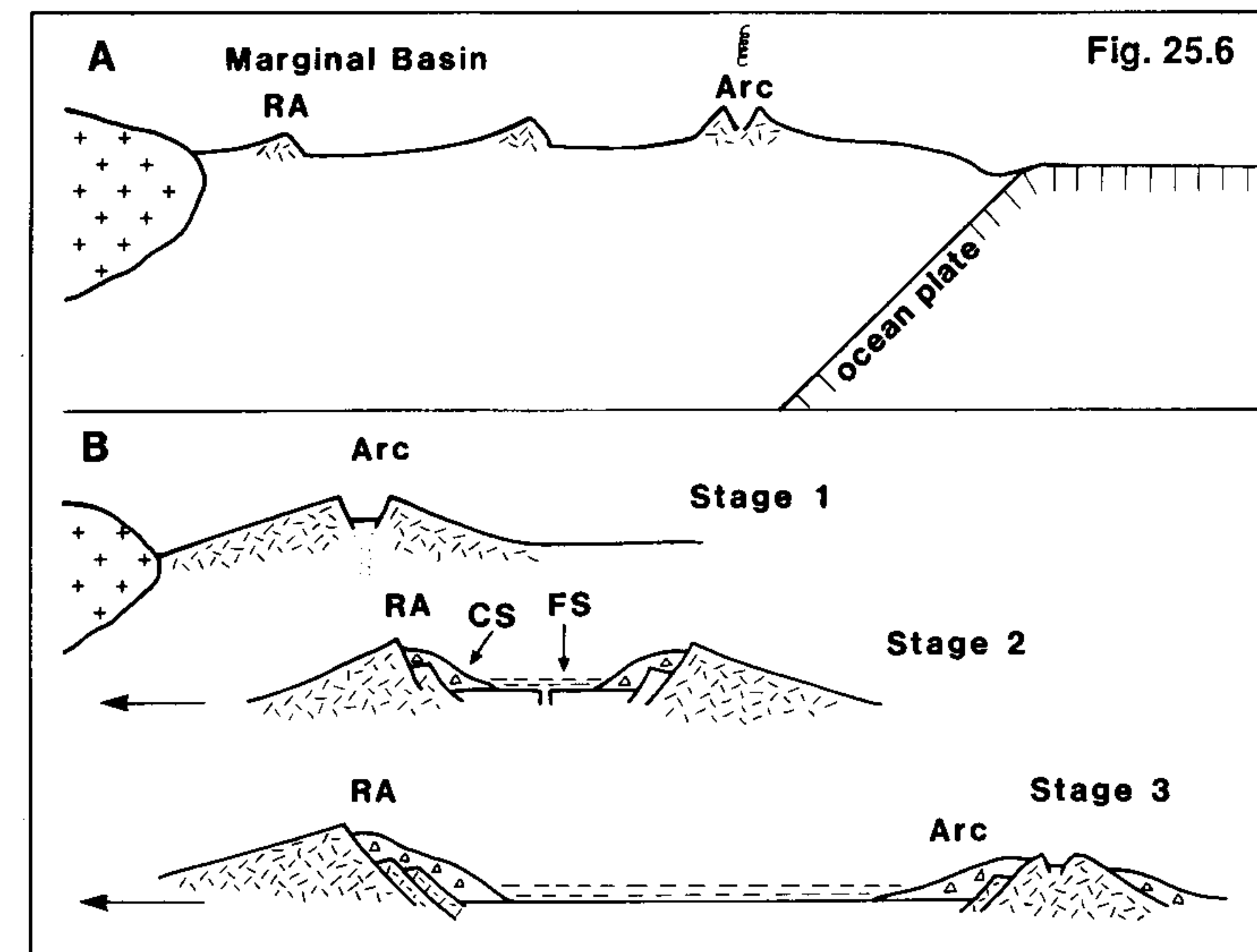
initially the pile is dominated by pillows which lack evidence of explosive activity, to be followed at the top of the pile by a relatively thin interval of shallow water and intertidal flows and finally the subaerial flows as the cone emerges to form an island (Fig.25.5B). As these lavas are built on 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 splitting 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 be 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 olistostrome-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, an essential difference 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

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- 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.

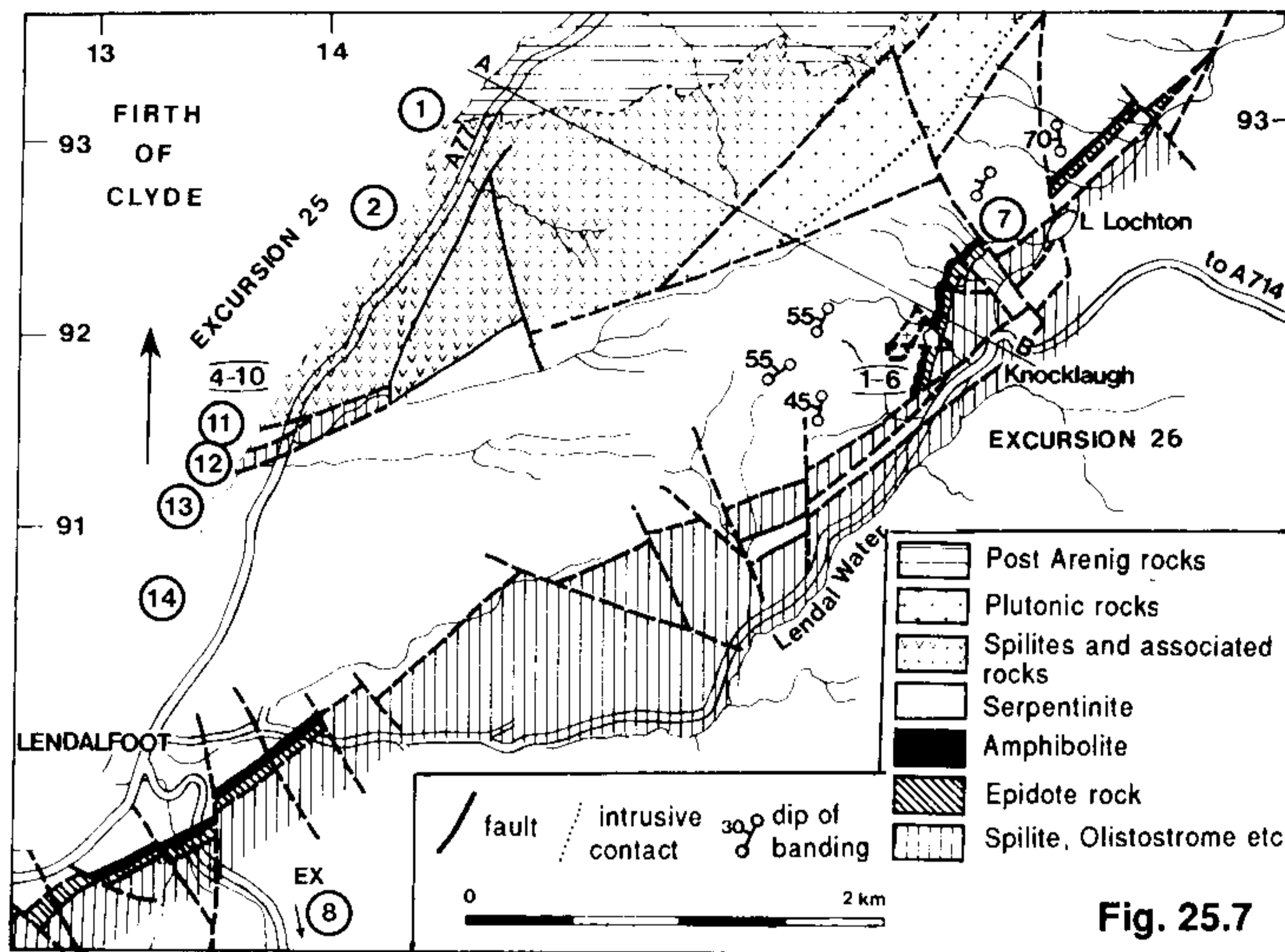


FIGURE 25.7. Simplified map of the northern part of the Ballantrae Complex, with positions of localities mentioned in Excursions 25 and 26.

The wave-washed surfaces at this outcrop are very polished and on them the dark lavas show indistinct 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 particularly 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

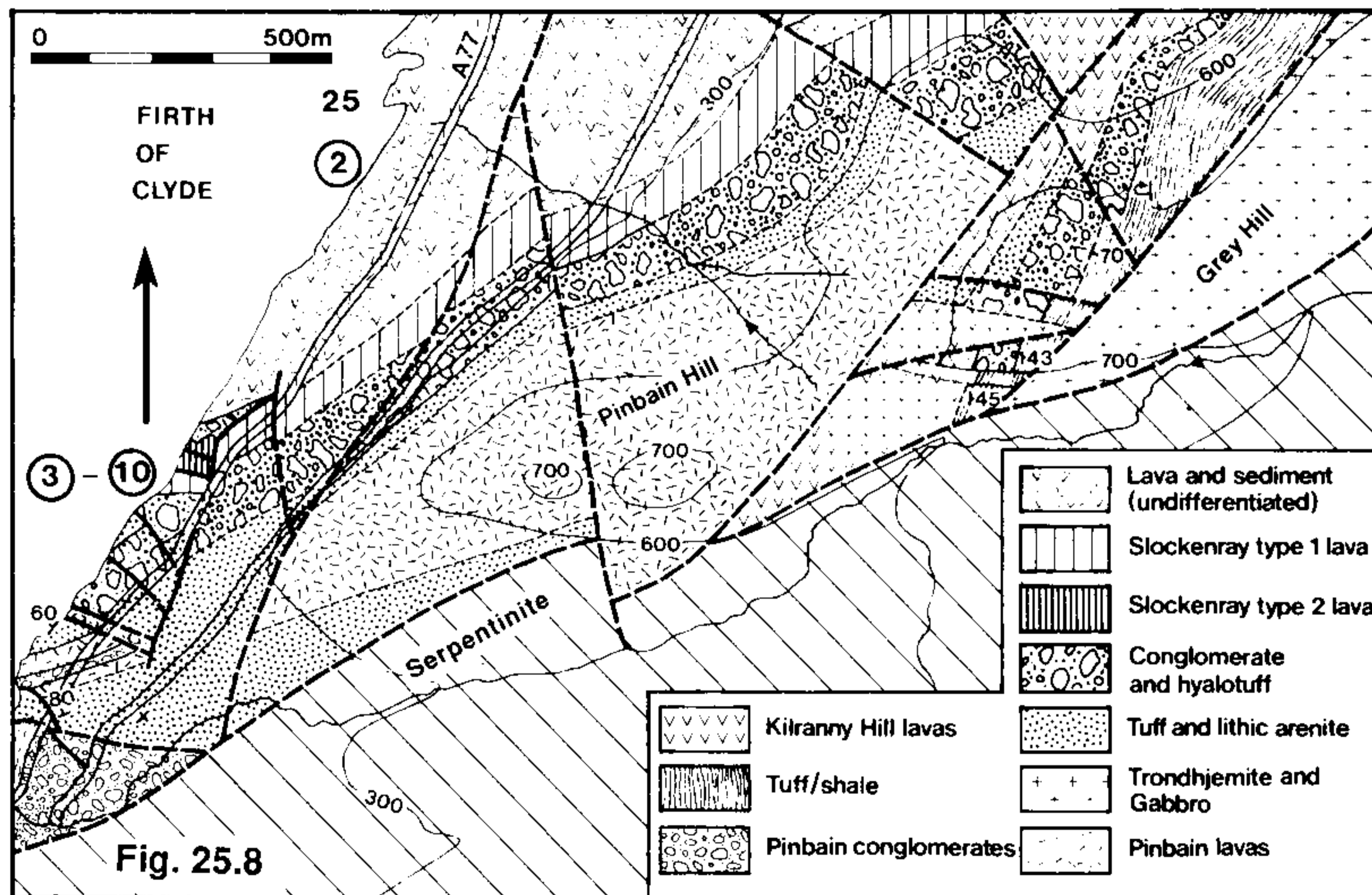


Fig. 25.8

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

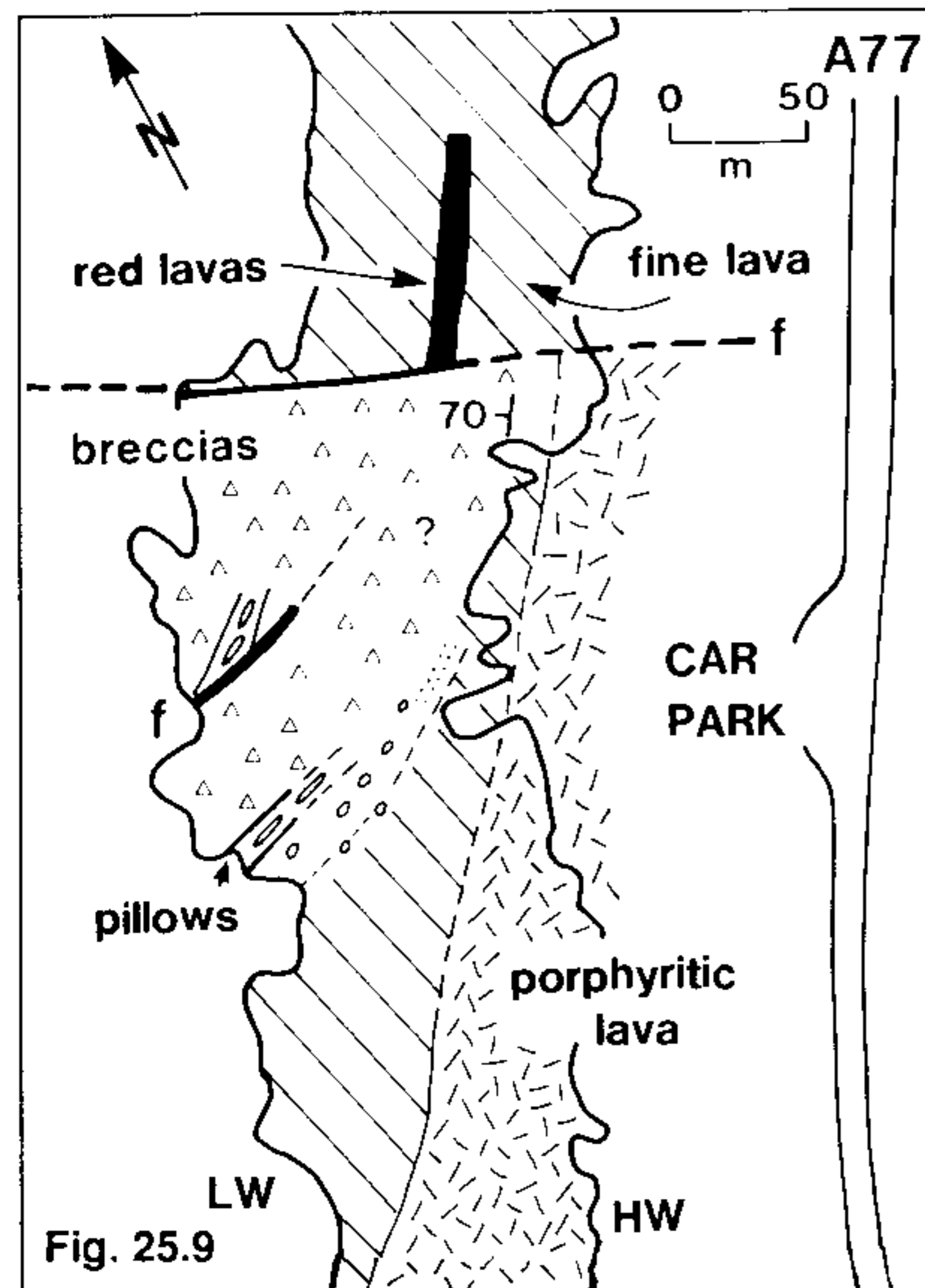


FIGURE 25.9. Sketch map to illustrate the geology at Locality 2, Pinbain Block. LW and HW refer to low and high water marks.

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 between 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

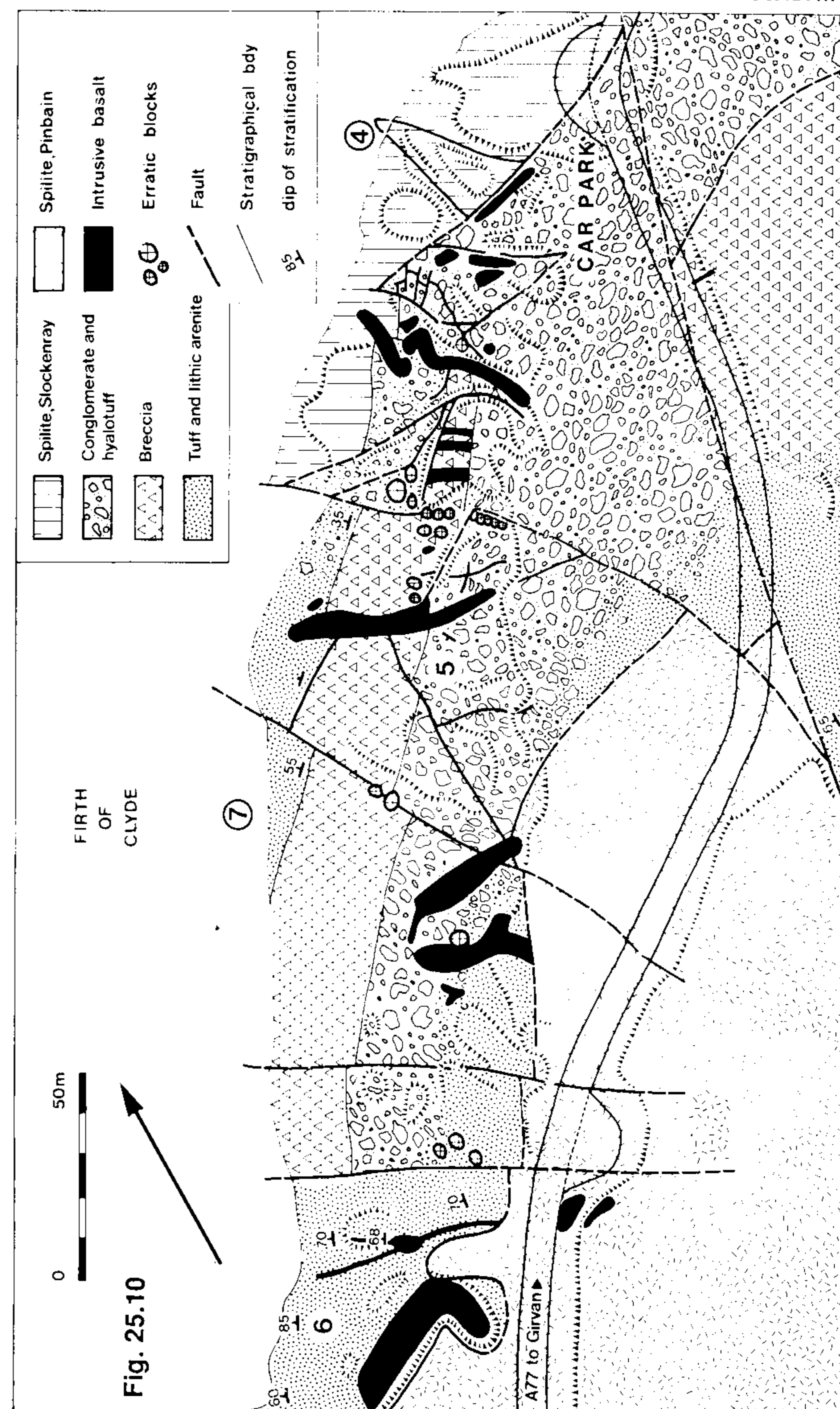


FIGURE 25.10. Map of the southern end of Slockenray. The north margin of map begins at the headland which divides this map from Figure 25.12, and the two lava types - porphyritic and dark aphyric (not subdivided in the map). They both belong to the Slockenray spilite of the caption.

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 porphyritic 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 plagioclase 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

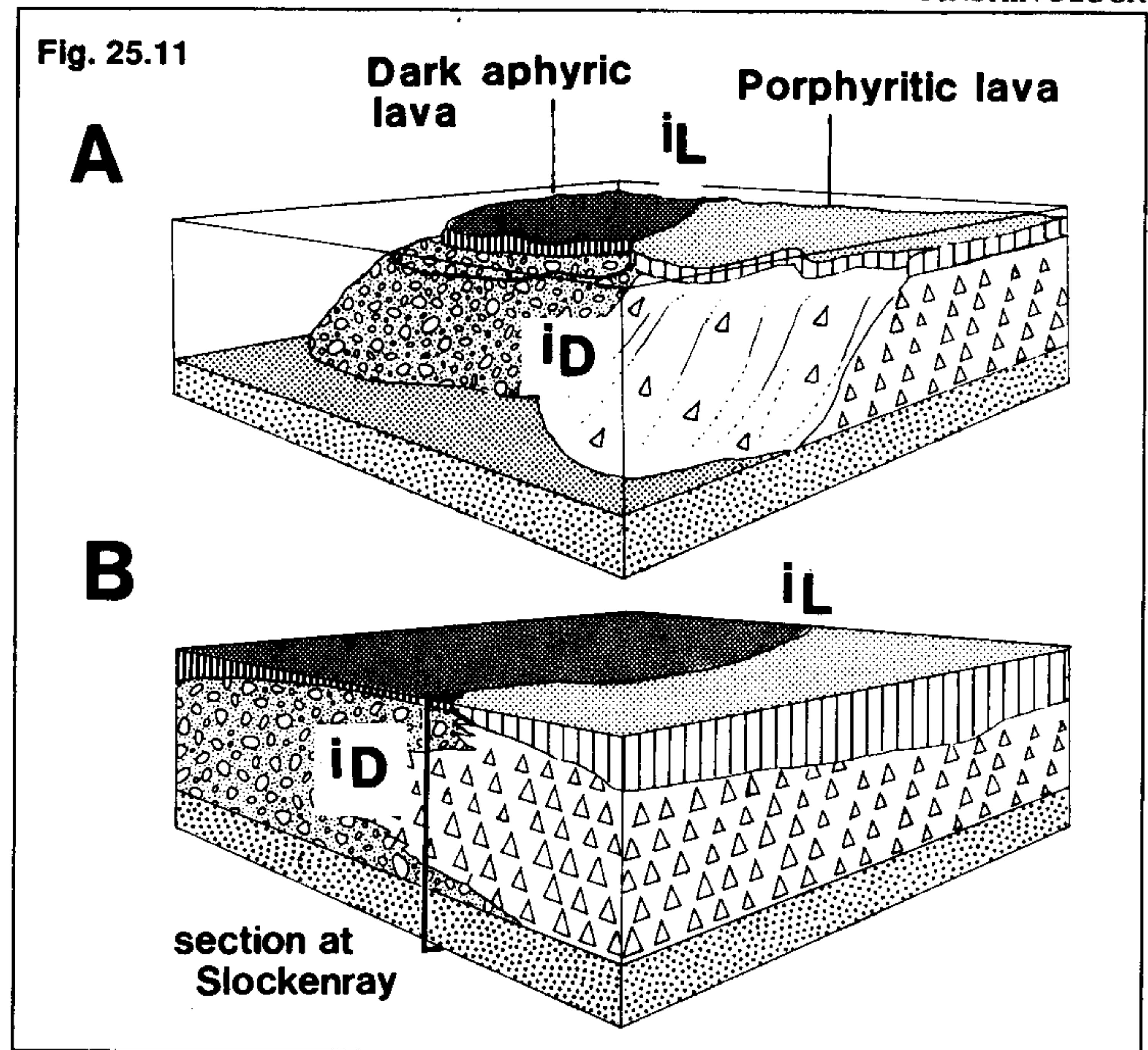


FIGURE 25.11. Explanation of the Slockenray sequence. **A.** Two lava flows, one porphyritic, the other dark aphyric, are both simultaneously extruded and flow together towards the coastline where they both build out delta cones adjacent to each other, each cone being sourced by their individual lava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

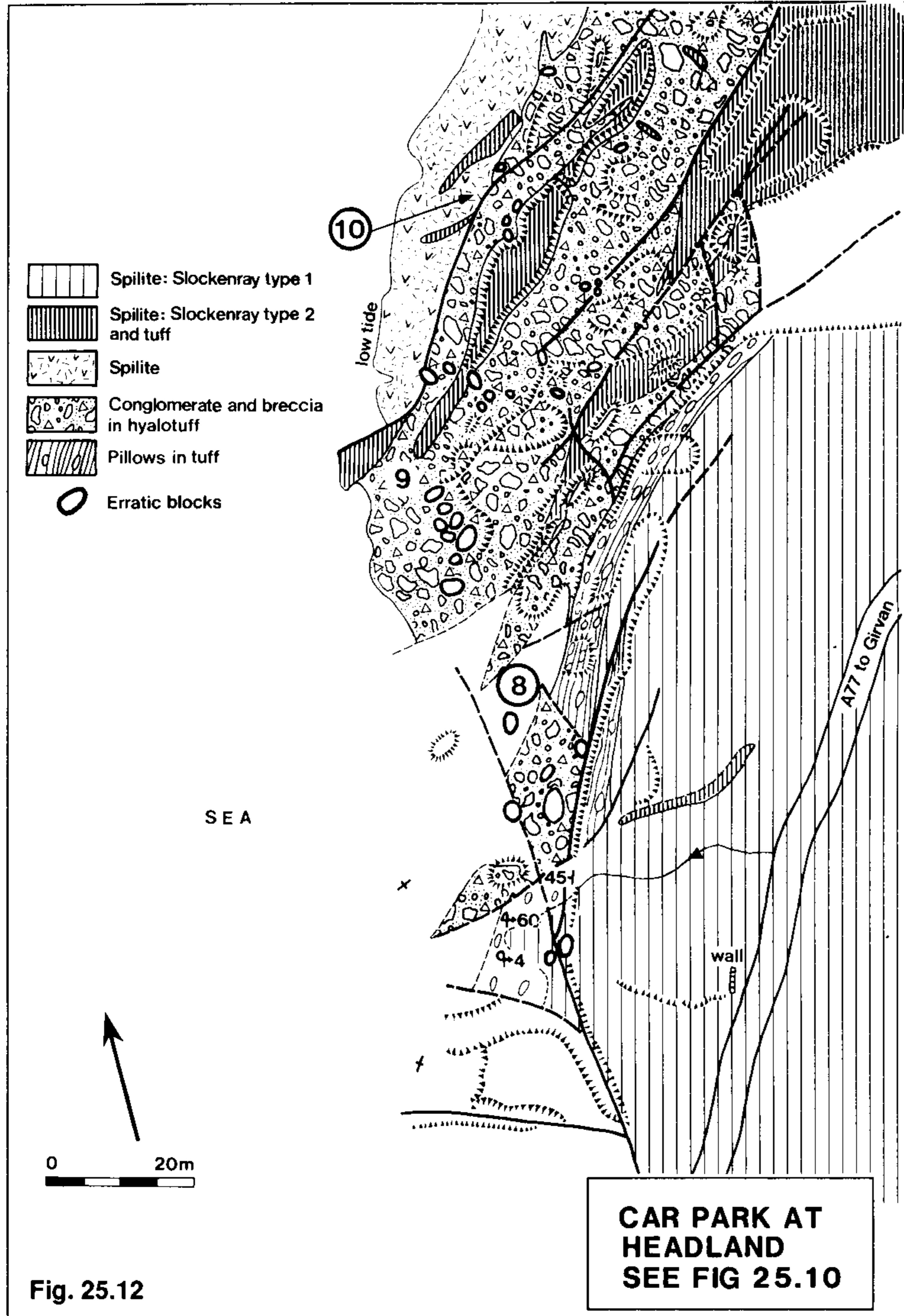


Fig. 25.12

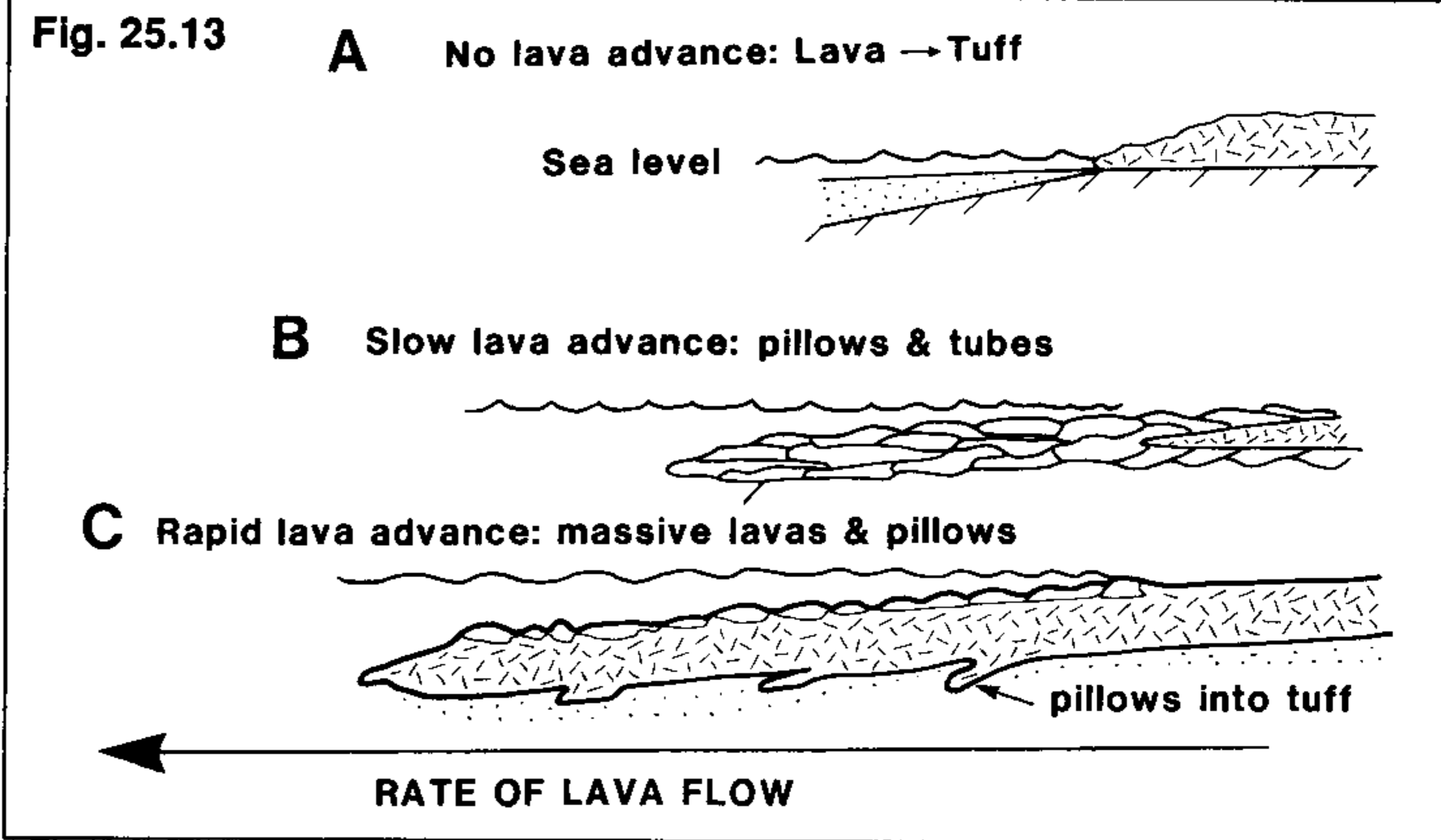


FIGURE 25.13. Explanation of the development of various lava structures and tuffs in lavas which enter the sea from the land. **A**, the lava front is moving slowly and as it enters the sea where it is rapidly chilled, all of it is converted to tuff at the shoreline. Waves and currents move the tuff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

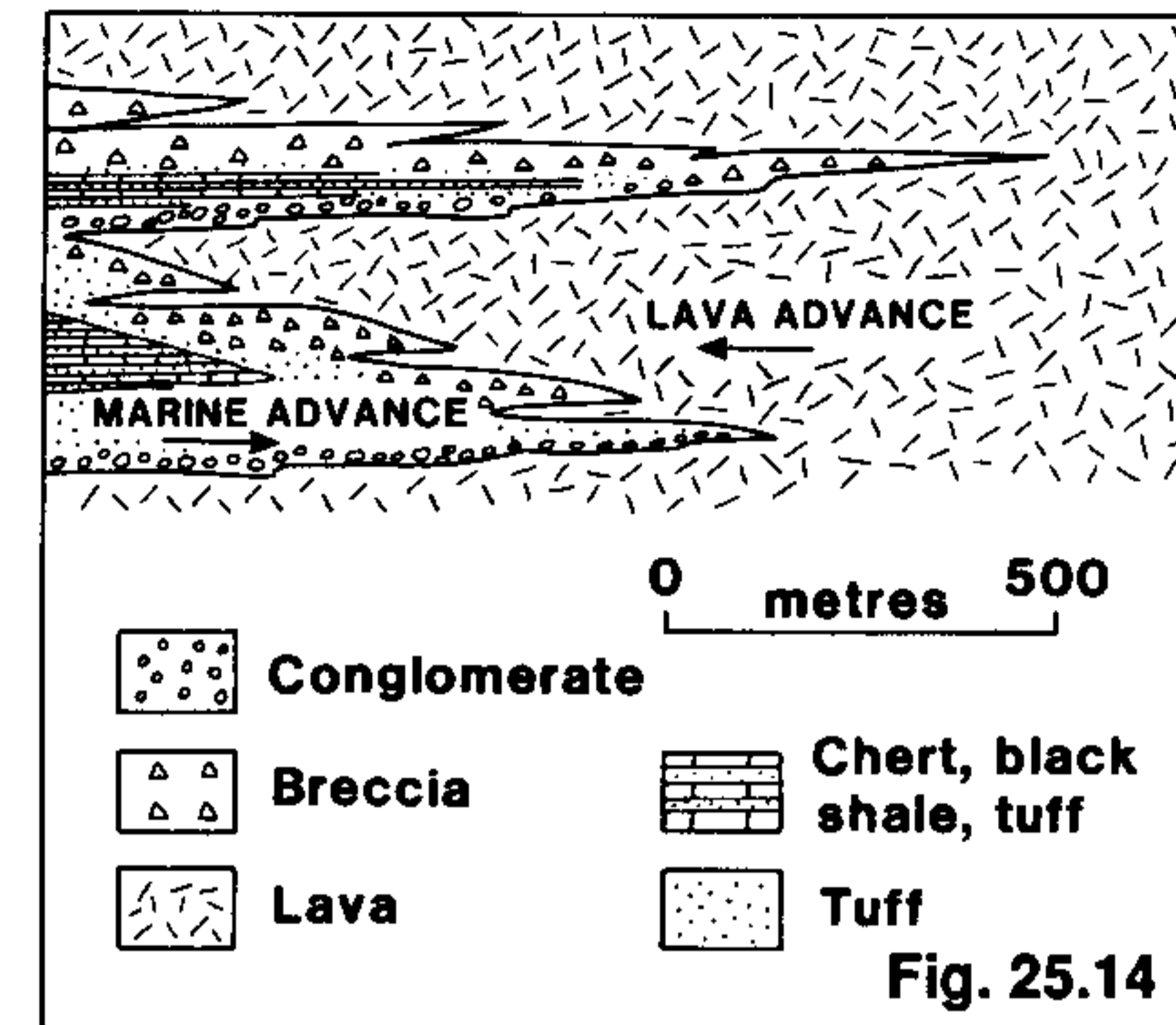


FIGURE 25.14. Explanation of the sediment-lava cycles in the Pinbain Block and elsewhere. When the rate of lava extrusion is rapid or the rate of sea-level change is slow, the lavas advance into the sea. Because of seawater-lava interactions, where the lavas break down by explosive or erosional activity,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

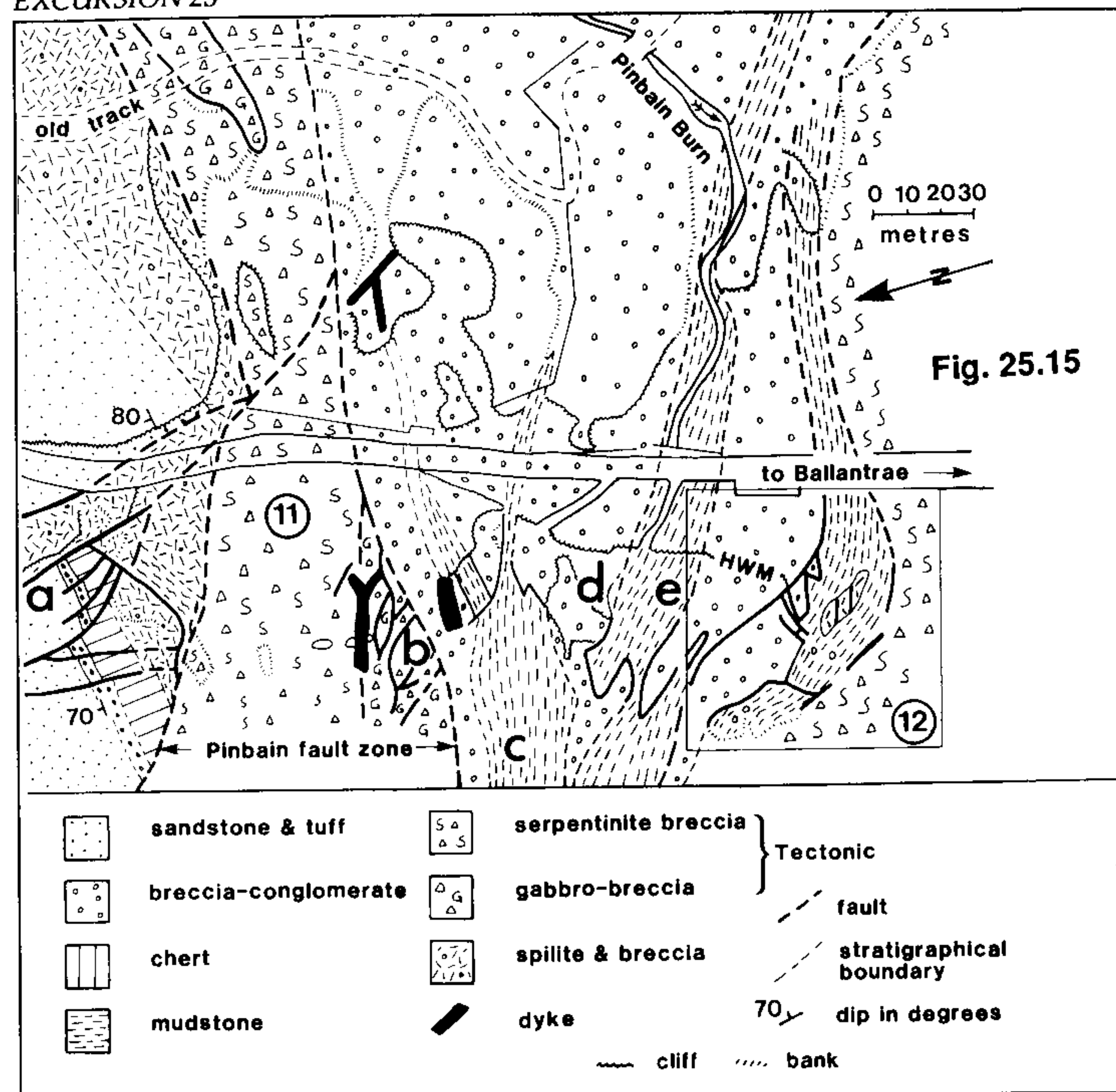


Fig. 25.15

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 interstratified 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

