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P. Gutteridge

Controls on carbonate sedimentation in a Brigantian intrashelf basin (Derbyshire)

P. GUTTERIDGE

SUMMARY: Reactivation of basement faults underlying the Derbyshire carbonate shelf caused the development of an intrashelf basin at the beginning of the Brigantian. The Station Quarry Beds were deposited in a contemporaneous syncline between the Longstone Edge Monocline and the Taddington — Bakewell Anticline. Growth of the Cressbrook Uplift caused the development of an angular uncomformity which separated the Station Quarry Beds from the overlying limestones. Deposition within this basin was by bioclastic calcisiltites and bioclastic turbidites which were sourced from surrounding shelf areas. Peritidal conditions were established in the basin during eustatic regressions. During the early Brigantian the northern basin margin was formed by an homoclinal carbonate ramp which later developed into a distally steepened ramp due to growth of the Longstone Edge Monocline. Shelf carbonates prograded across the northern margin on at least three occasions in response to varying subsidence rates within the basin. The Taddington — Bakewell Anticline controlled the southern margin during the early Brigantian; basinal conditions later spread southwards across this anticline. During the early Brigantian the intrashelf basin was connected to the Edale Basin to the east; however, the development of the Edensor Anticline during the late Brigantian caused partial isolation of the intrashelf basin.

Dinantian sedimentation on the Derbyshire Dome has been discussed by Gutteridge (1987) and Gawthorpe *et al.* (this volume) who indicated that a change in sedimentation took place at the Asbian/Brigantian boundary. At this time the influence of tectonic features on sedimentation became apparent, volcanic activity reached a peak and an intrashelf basin developed over part of the carbonate shelf. The objectives of this paper are to describe the controls on sedimentation in this intrashelf basin and to use this information, together with gravity data, to refine the model of basement structure presented by Gutteridge (1987). The study area and the localities referred to in the text are shown in Figure 1.

Previous work on Brigantian limestones (Walkden 1970, 1977; Stevenson & Gaunt 1971; Butcher & Ford 1973; Brown 1973; Adams and Cossey 1978; Pazdzierski 1982; Gutteridge 1983, 1987; Aitkenhead *et al.* 1985; Currie 1987) has shown that they were deposited in a variety of environments, and it is generally accepted that this diversity reflects the development of a deeper-water area within the carbonate shelf.

Since the sediments described here are interpreted as the deposits of an intrashelf basin the distinction between an intrashelf basin and the major basinal environments known from the Dinantian (e.g. the Bowland Basin, described by Gawthorpe 1986, 1987) should be made clear. In this case, the intrashelf basin developed as a result of fault-controlled subsidence beneath part of a pre-existing carbonate shelf.

The extent of basin development is determined by the rate of subsidence within the basin. If subsidence rate is relatively slow, carbonate sedimentation may keep pace with subsidence and a carbonate shelf maintained. A possible example of this was described by Somerville (1979a) and Gray (1981) from the Asbian of North Wales. They inferred that fault-bounded embayments in the North Wales shelf were subsiding faster than the surrounding shelf areas. In these embayments shelf carbonate cycles are thicker than in the surrounding areas and incorporate a basal argillaceous subtidal phase which is absent in cycles deposited on the surrounding shelf areas. In such cases subsidence was not fast enough to disrupt the continuity of the carbonate shelf, but was manifest as a thickening of shelf cycles and a change in the style of cyclicity. If subsidence rate outstrips the sedimentation rate so that the shelf subsides below the depth at which carbonate production can be sustained, a deep basin may be established with the development of rimmed margins at the surrounding shelf edges (Read 1982, 1985; Rees 1986). Subsidence rates in intrashelf basins are inferred to lie between the cases outlined above, so that water depth does not exceed that required to supress carbonate production and the sedimentation rate keeps pace, at least periodically, with the subsidence rate.

1. FACIES

Nine facies have been recognised and their distribution is shown in Figure 8.

1.1. Pale facies

1.1.1. Description

This facies comprises repeated cycles 0.5 m to 4 m in thickness whose boundaries are defined by emergent surfaces. A cycle consists of a basal coarse bioclastic grainstone/packstone. Bioclasts are disarticulated, fragmented, abraded and occasionally have oncolitic coatings. This grades into a bioturbated bioclastic packstone/wackestone which contains discontinuous layers of coarse bioclasts and occasional colonial corals. Cross-bedding is rare in this unit. The upper parts of cycles are fine-grained, sorted bioclast peloid grainstones with rhizocretions and alveolar textures. Wackestones with a restricted fauna and fenestrae are locally present at the top of some cycles. Chert is rarely present.

1.1.2. Interpretation

This facies comprises repeated shallowing-upwards cycles. The basal coarse bioclastic unit of each cycle probably accumulated in shallow-turbulent water during reflooding of the shelf following emergence. A lower energy subtidal environment was established at the height of the transgression with deposition taking place near wave-base. Occasional winnowing by storms led to the development of bioclastic lag deposits which were later mixed with the surrounding sediment by bioturbation. Cross-bedding indicates occasional reworking by currents, although its rarity probably reflects intensive bioturbation. A diverse bioclast suite indicates full marine salinity. The sorted peloid grainstone with rhizocretions and alveolar texture found at the top of cycles was deposited in a well reworked shallow subtidal or shoreface environment which was partly emergent. The local fenestral wackestones present at the top of some cycles are interpreted as former intertidal deposits. The pale facies is similar to other Asbian and Brigantian carbonate sequences comprising repeated shallowing-upwards cycles found in northern England and North Wales interpreted by Somerville (1979b), Bridges (1982), Walkden (1987) and Horbury (this volume) to have been deposited on flat-topped carbonate shelves.

1.2. Evenly bedded facies

1.2.1. Description

This facies is a bioclastic packstone/wackestone with a strong bituminous smell. Bedding is picked out by the presence of planar fissile limestones which alternate with hard limestones (terminology of Bathurst 1987). Fissile limestone vary in thickness from 0.01 m to 0.05 m and consist of pressure dissolution seams which weather to produce shaley partings. Hard limestones vary in thickness from 0.05 m to 0.3 m (Fig. 2a). This facies is

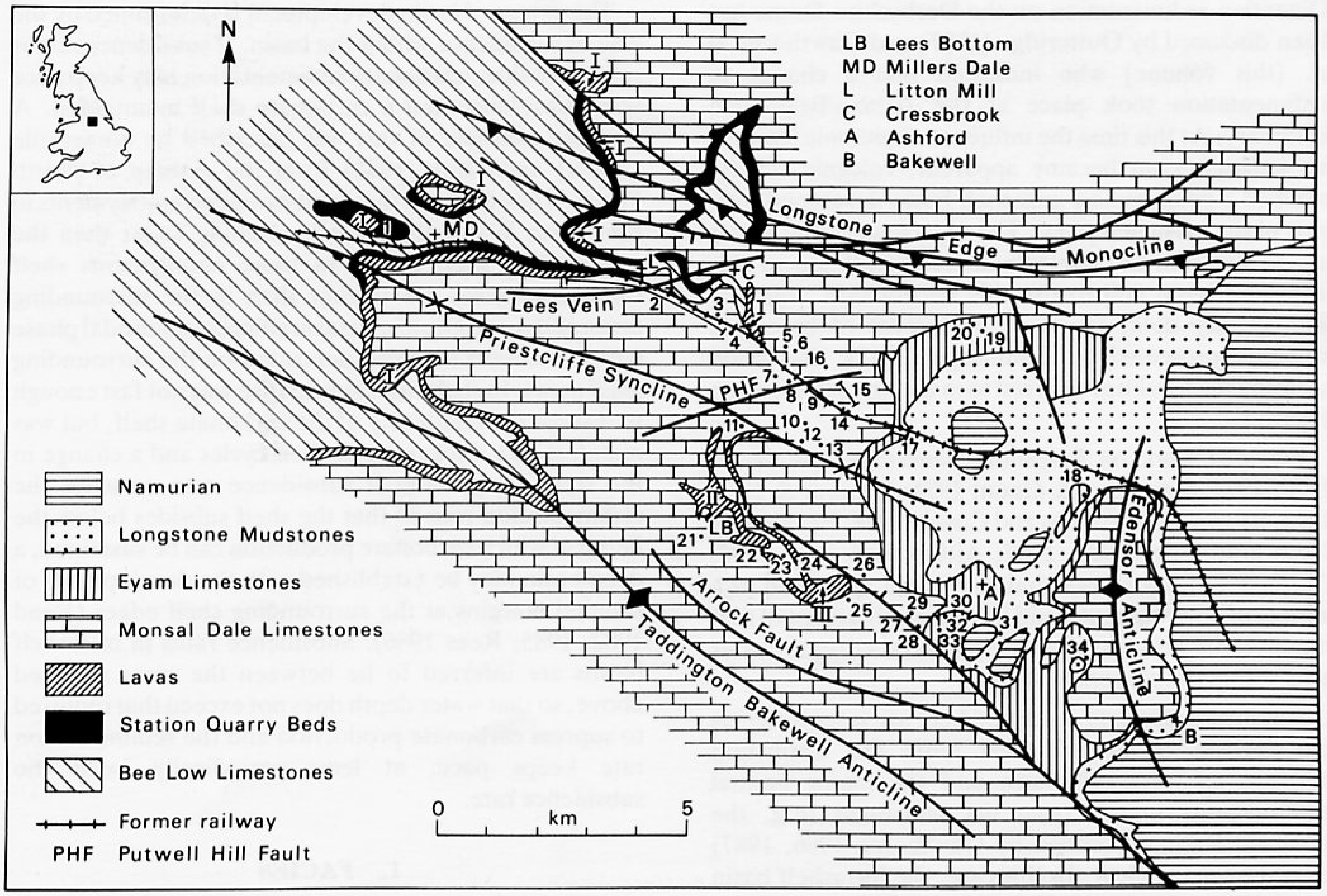

Fig. 1. Geology of the intrashelf basin area. Numbers are localities referred to in the text: I. Upper Millers Dale Lava and Cressbrook Lava; II. Lees Bottom Lava; III. Shacklow Wood Lava.

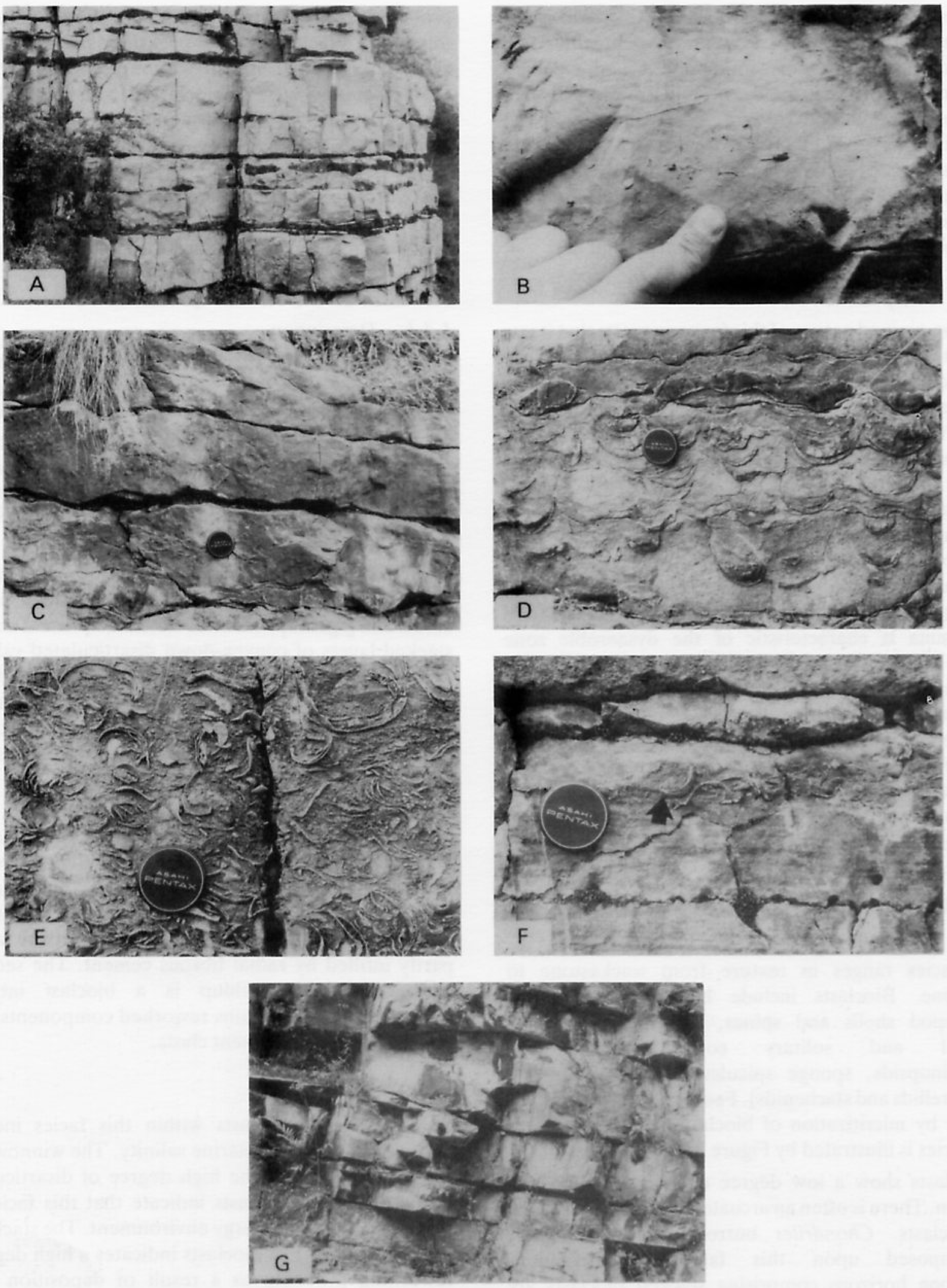

Fig. 2. a. Evenly bedded facies at Upper Dale (Locality 5). Length of hammer 0.3 m; b. Evenly bedded facies showing matrix-supported cluster of whole bioclasts. Locality 4. Finger is 1 cm in width; c. Bioclastic wackestone/packstone facies. Locality 5. Lens cap is 5 cm in diameter; d. Bioclastic grainstone/packstone facies, showing highly stacked gigantoproductid valves. Locality 2; e. Chaotic layer (Interval B) in resedimented coarse bioclastic grainstone interpreted as suspension deposition from a high density turbidite. Locality 8; f. Planar laminated, sorted bioclastic sand (Interval C) overlain erosively (contact arrowed) by a layer of stacked brachiopod valves (Interval A) grading into sorted bioclastic sand (Interval C). Bedload deposition of successive low-density turbidites. Locality 8; g. Thickly bedded, graded bioclastic calcisiltites with erosive bases, interpreted as low-density turbidites. Locality 9.

composed of bioclastic calcisiltite consisting predominantly of fragmentary brachiopods, echinoderms, bryozoans, foraminifera and calcified molluscs. Whole gastropods, crinoid ossicles, bivalve and brachiopod valves are rare and occur in discontinuous stacked layers or as matrix-supported clusters (Fig. 2b). Sediment mottling defined by a patchy distribution of siliciclastic mud is present. *Chondrites* burrows are also present. Chert is abundant and occurs as tabular bodies up to 0.1 m in thickness (Fig. 2a).

1.2.2. Interpretation

The fine-grained nature of this facies suggests that it was deposited by settling at depths below wave-base. Occasional layers of stacked whole bioclasts may represent influxes of coarse bioclasts or concentrations due to winnowing. The scarcity of whole fossils suggests a restricted environment due either to unusual salinity or to low oxygen levels. The range of fossils present indicates normal marine salinity. Mottling of sediment was probably caused by limited bioturbation. The common occurence of *Chondrites* suggests a poorly oxygenated environment (Bromley & Ekdale 1984). The scarcity of shelly fauna, in addition to the limited ichnofauna is characteristic of the dysaerobic zone (Byers 1977). The bioclasts present suggests that the sediment was derived from the surrounding shelf areas. The mechanism of deposition was probably similar to the deposition of periplatform carbonates around modern carbonate platforms proposed by Boardman & Neumann (1984) where fine-grained sediment is winnowed and transported off platforms by storm and tidal currents and deposited in adjacent basinal areas.

1.3. Bioclastic wackestone/packstone facies

1.3.1. Description

This facies ranges in texture from wackestone to packstone. Bioclasts include bivalves, gastropods, brachiopod shells and spines, crinoids, bryozoans, colonial and solitary corals, foraminifera, saccaminopsids, sponge spicules and skeletal algae (ungdarellids and stachenids). Faecal pellets and peloids formed by micritization of bioclasts are also present. This facies is illustrated by Figure 2c.

Bioclasts show a low degree of disarticulation and abrasion. There is often an arcuate or random alignment of bioclasts. *Chondrites* burrows are occasionally superimposed upon this fabric. Discontinuous grainstone horizons comprising stacked bioclasts are surrounded by matrix-supported bioclasts which display a random or arcuate orientation. Chert occurs as isolated rounded and tabular nodules and as a marginal replacement of bioclasts.

1.3.2. Interpretation

The diverse range of bioclasts indicates normal marine salinity. The random and arcuate alignment of bioclasts is attributed to bioturbation. Faecal pellets were probably produced by gastropods which are abundant in this facies. Wackestone and packstone textures indicate a low-energy environment with infrequent reworking. Discontinuous layers of bioclasts are interpreted as lag deposits which were subsequently mixed with the surrounding sediment by bioturbation. Deposition therefore took place in generally low-energy conditions, but occasional higher-energy conditions caused winnowing. This suggests deposition was at depths below normal wave-base but above storm wave-base.

1.4. Bioclastic grainstone/packstone facies

1.4.1. Description

This facies consists of moderately sorted medium- to coarse-grained bioclastic peloid sand. Bioclasts include crinoids, brachiopod shells and spines, bryozoans, corals, foraminifera and skeletal algae (green algae and stachenids). Peloids formed by micritization of bioclasts are also present. Chert has the same mode of occurrence as in the bioclastic wackestone/packstone facies (Section 1.3.).

Bioclasts are disarticulated, highly fragmented and abraded. Coarser bioclasts include abraded solitary corals and gigantoproductids which are present either as stacked layers of convex-down disarticulated valves or as convex-down articulated shells sitting inside one another (Fig. 2d). In some cases erosional surfaces are overlain by layers of stacked, disarticulated gigantoproductid valves which pass up into well sorted, cross-stratified crinoidal sand.

A bryozoan bioherm is present at locality 10, which consists of a framework of *in situ* stick-type, fenestrate and encrusting bryozoans and vermetid gastropods. Fragmented and abraded bioclasts and peloidal carbonate mud have been trapped within the framework. There is a high framework porosity which is partly infilled by radial fibrous cement. The sediment surrounding the buildup is a bioclast intraclast packstone which contains reworked components of the build-up including cement clasts.

1.4.2. Interpretation

The variety of bioclasts within this facies indicates deposition in normal marine salinity. The winnowing of carbonate mud and the high degree of disarticulation and abrasion of bioclasts indicate that this facies was deposited in a high-energy environment. The stacking of shells and sorting of bioclasts indicates a high degree of reworking probably as a result of deposition above wave-base. The occurrence of cross-stratification indicates the influence of currents.

1.5. Resedimented carbonates

Three types of resedimented carbonates have been recognised:

1.5.1. Graded coarse bioclast/intraclast grainstone

Bioclasts in these limestones are abraded, disarticulated and range in size from whole to highly comminuted. Silicification occurs as a peripheral replacement of bioclasts, whereas chert nodules are rare. The beds commonly have undulatory erosional bases. Their structure is composite comprising one or more of the following intervals (Fig. 3):

Interval A. Stacked shells which are aligned subparallel to bedding. This interval often overlies an erosive surface and grades into either interval B or C (Fig. 2f).

Interval B. This consists predominantly of randomly-oriented whole brachiopod valves and large, articulated crinoid stems (Fig. 2e). There is no preferential alignment of bioclasts. Very coarse bioclasts such as reworked colonial corals occur at the top of this interval (Fig. 3).

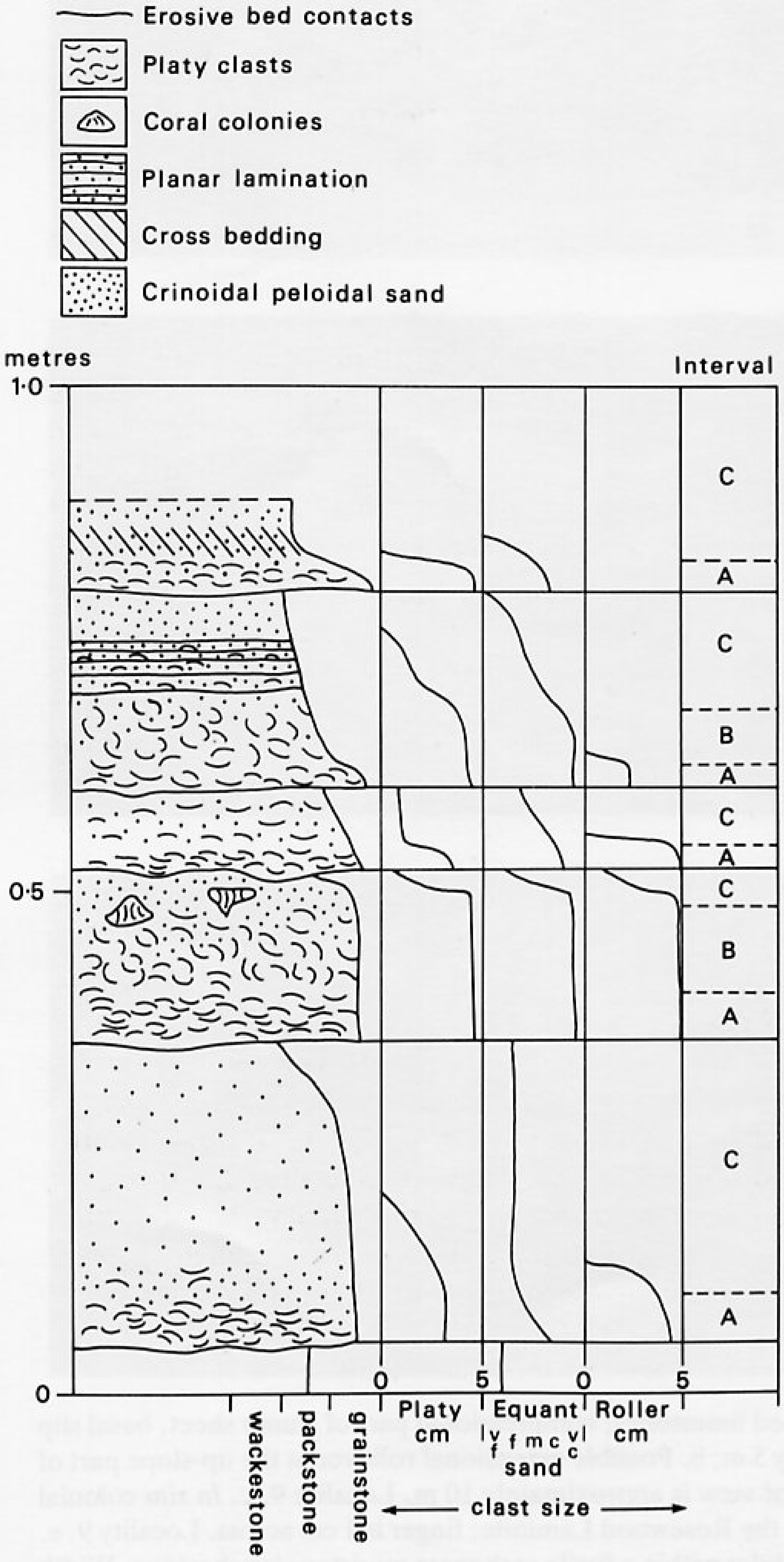

Fig. 3. Log through graded coarse bioclastic grainstone showing composite nature of the beds. Locality B.

Interval C. This consists of planar or cross-bedded, sorted crinoidal peloidal sand (Fig. 2f). Occasional shells present within this interval, lie parallel to lamination. Possible escape burrows are present in this interval. Interval C occasionally grades up into shale.

The concentration of differently-shaped bioclasts within each interval suggests that some mechanism of selective deposition was operating. There is a general fining-upwards within each bed indicating a decrease in current strength. Intervals A and C suggest systematic rather than rapid deposition probably as a result of bedload processes. The chaotic nature of interval B suggests rapid deposition possibly due to the dumping of a suspended load. The occurrence of very coarse bioclasts at the top of interval B is suggestive of inverse grading found in grainflows and high density turbidites by Walker (1975). These beds were deposited by a variety of processes which acted during a single event and are interpreted as a combination of high and low density turbidites (Lowe 1982). These beds contain bioclasts and peloids which indicate a shallow water provenance. Dispersal directions are uncertain as there are few palaeocurrent indicators. These turbidites were sourced from the surrounding shelf areas and intrabasinal highs.

1.5.2. *Bioclastic calcisiltites*

These are impure limestones 0.1 m — 2 m thick comprising fine sand to silt-sized comminuted bioclasts (Fig. 2g). Whole fossils are rare, but occasional brachiopod valves, thin-shelled bivalves and trilobite carapaces lie parallel to bedding. Chert is rare in these limestones occurring as nodules and tabular bodies. The bases of these beds are sharp and sometimes erosional. The beds are graded which is defined by an upward decrease in maximum grain-size, upward decrease in the abundance of bioclasts coarser than fine sand size, and an upward increase in the amount of siliciclastic mud. Interbedded shales contain an *in situ* fauna of corals and brachiopods and a drifted fauna of goniatites, ostracods and trilobites (Butcher & Ford 1973). The limestones contain a reworked shallow marine fauna including crinoids, bryozoans, and productoid brachiopods. The ichnofauna in the limestones includes *Thallassinoides, Zoophycos* and *Chondrites*. Sediment mottling defined by a patchy distribution of siliciclastic mud is present.

The petrography of these beds indicates they were derived from a shallow marine environment. The lack of reworking and the interbedded shales suggests deposition in a low-energy environment below wave-base. The sharp and occasionally erosive base and the grading suggest deposition from a waning flow, probably as low-density turbidites.

The difference in depositional process and components of the two types of resedimented carbonates may be related to differing source areas or proximal to distal trends. Although the ichnofauna of the turbidites suggests oxygen levels within the basin were low, the presence of an *in situ* fauna of corals and brachiopods in the interbedded shale implies that the basin was periodically oxygenated.

1.5.3. Slumps

Horizons of recumbent folding and disrupted bedding, interbedded with undisturbed beds, are interpreted as the compressional parts of slump sheets (Fig. 4a). A discontinuity seen at Locality 9 is interpreted as an extensional rollover developed at the trailing edge of a slump sheet (Fig. 4b).

All slumps are overlain by erosive based, graded beds of coarse bioclast intraclast grainstone interpreted as high density turbidites. The petrography of these beds is similar to the underlying slump and their common association with slumps suggest that the generation of the turbidite was linked to the generation of the slump.

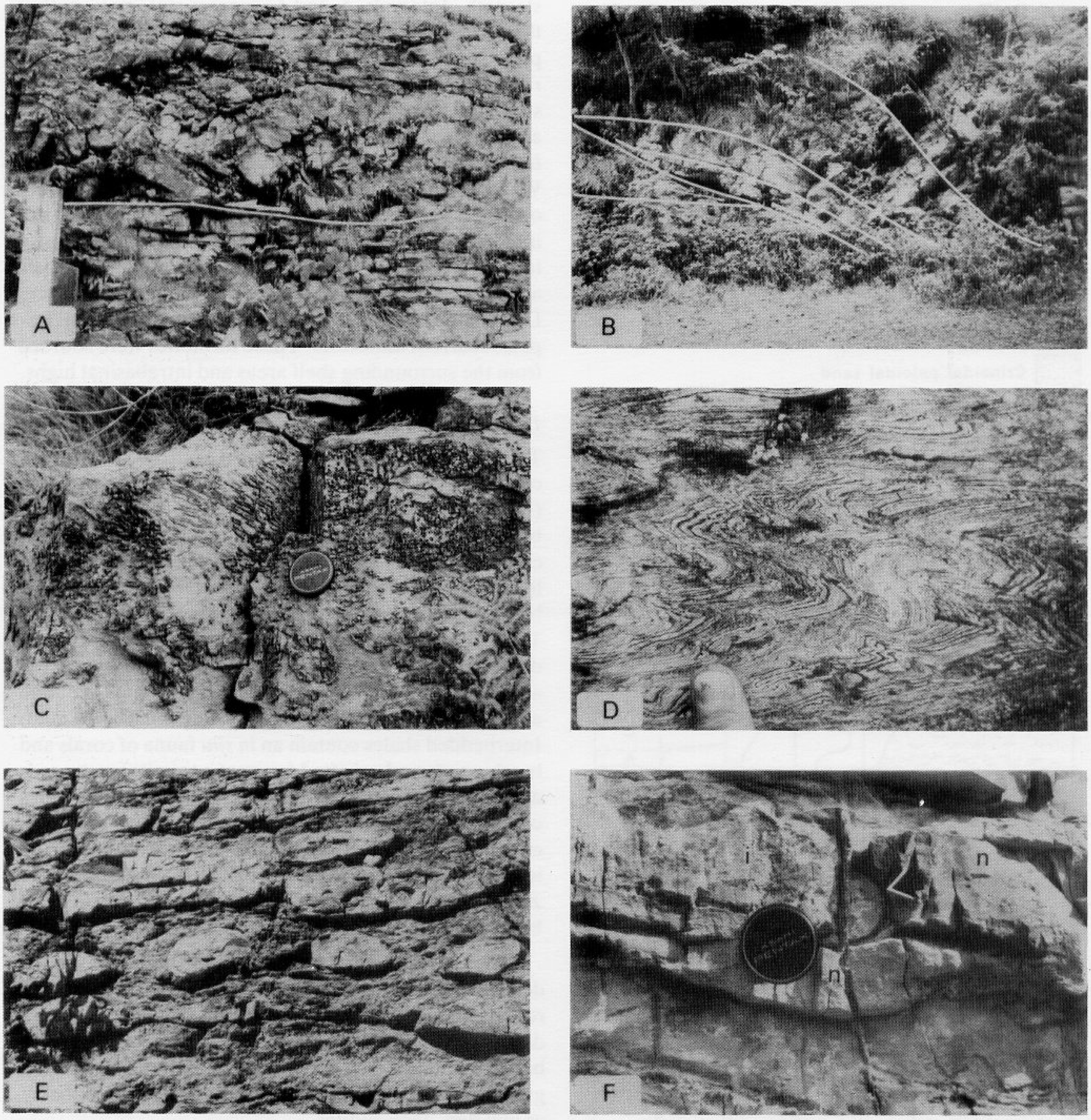

Fig. 4. a. Horizon of recumbent folding interbedded with undisturbed limestones; compressional part of slump sheet, basal slip plane is indicated. Locality 4. Height of view is approximately 5 m; b. Possible extensional rollover at the up-slope part of a slump sheet; interpretation drawn on photograph. Width of view is approximately 10 m. Locality 9. c. *In situ* colonial coral in Hob's House Coral Bed. Locality 10. d. Slumping in the Rosewood Laminite; finger is 1 cm across. Locality 9. e. Nodular facies showing horizons of sub-spherical lithified nodules within a fissile carbonate mudstone/wackestone; Width of view is approximately 1 m. Locality 19. f. Nodular facies showing tabular nodules (n) interbedded with fissile sediment (i); margin of nodule (arrowed) is formed by a brachiopod valve. Locality 19.

Gawthorpe & Clemmey (1985) suggested that partly lithified slump sheets may disaggregate to form debris flows. The slumps found in this basin have not disaggregated and the turbidite was deposited after emplacement of the slump sheet. Retrogressive slope failure is suggested as the common cause of slumping and turbidite generation in this case (Pickering 1979). It is not, however, suggested that all tubidites were generated by slumping, as it is possible that some were generated by storm activity (Aigner 1985). Palaeoslopes were deduced from slump structures using the method of Woodcock (1979); these are shown in Figure 5.

1.6. Coral beds

1.6.1. *Description*

Three coral beds within the succession (Fig. 8) have been named (from the lowest up) the Upper Dale, Hob's House and Whitecliff coral beds. Lists of coral species are given by Butcher & Ford (1973), Cossey (1983) and Aitkenhead *et al.* (1985). Coral beds range in thickness from 0.3 m to 2 m. The majority of coral colonies are abraded and inverted, and in some cases colonies rest on each other with a zone of broken corallites at the contact. Other bioclasts within the coral beds include brachiopod valves and crinoid ossicles and stems. Coral beds occasionally rest on scoured surfaces and are graded (e.g. the Upper Dale Coral Bed at Locality 4). Most of the corals within the Whitecliff and Hob's House coral beds appear to be preserved *in situ.* The Hob's House Coral Bed contains large colonies several metres across, whose growth commenced as symmetrical bushes but later developed laterally spreading branches (Fig. 4c). At the top of the Hob's House Coral Bed an accumulation of broken corallites and rolled solitary corals indicates an increase in reworking. The Whitecliff Coral Bed is a concentration of colonies within the bioclastic grainstone/packstone facies, many of which appear to be preserved *in situ.*

1.6.2. *Interpretation*

Erosively based, graded coral beds containing reworked colonies indicate transportation of colonies into deeper parts of the basin, probably by debris flow. Cossey (1983) suggested that distance of transport of these corals was not great. Coral beds containing *in situ* colonies occurring in association with the bioclastic grainstone/packstone facies (sections 1.4. and 2.2.) were deposited in a well reworked subtidal environment which was close to wave-base. The widespread development of coral beds reflects the occurrence of these conditions over much of the basin. Brown (1973) suggested that the coral beds encroached into the deeper parts of the intrashelf basin from surrounding shelf areas during regressions.

1.7. Laminite facies

1.7.1. *Description*

This facies is a dolomitized limestone which is laminated on a millimetre to centimetre scale (Fig. 4d). There is evidence of soft sediment deformation which has been used to determine palaeoslope direction (Fig. 5). The four occurrences of this facies have been named (from the lowest up) the Monsal, Rosewood (also known as the Rosewood Marble), Ashford and Headstone laminites (Fig. 8). Two types of lamination are present:

1. Carbonate mudstone/dolomicrite laminae are up to several centimetres in thickness and are composed of calcite and dolomite micrite displaying an homogeneous, peloidal or graded texture. Thicker laminae contain laminar, tubular and irregular fenestrae. Occasional spar-filled, coffin-shaped and hexagonal pseudomorphs occur in layers, rosette-like clusters and in association with irregular fenestrae. These pseudomorphs are often slightly compacted. Bedding surfaces of laminae occasionally have a polygonal network of ridges a few millimetres high. The lamination is defined either by partings on which terrestrial plant fragments are concentrated or by the interbedded grainstone laminae described below.
2. Grainstone laminae are up to 5 mm in thickness. Two types have been recognised. Firstly, those containing sorted, comminuted micritized bioclasts including bryozoans, brachiopods and foraminifera. Peloids formed by micritization are also present. Secondly, those containing an unusual bioclast assemblage described by Gutteridge (1983) which includes tubular microfossils, smooth ostracods, bivalves with mixed calcite/aragonite shells and faecal pellets. Some mat-like organisms similar to ?*Aphralysia* and ?*Renalcis* are also present. Both types of grainstone laminae are partly replaced by a dolomite mosaic consisting of crystals 30μm to 90μm in size. This phase of dolomitization post-dates the origin of the dolomicrite.

Other lithologies included in the laminite facies are

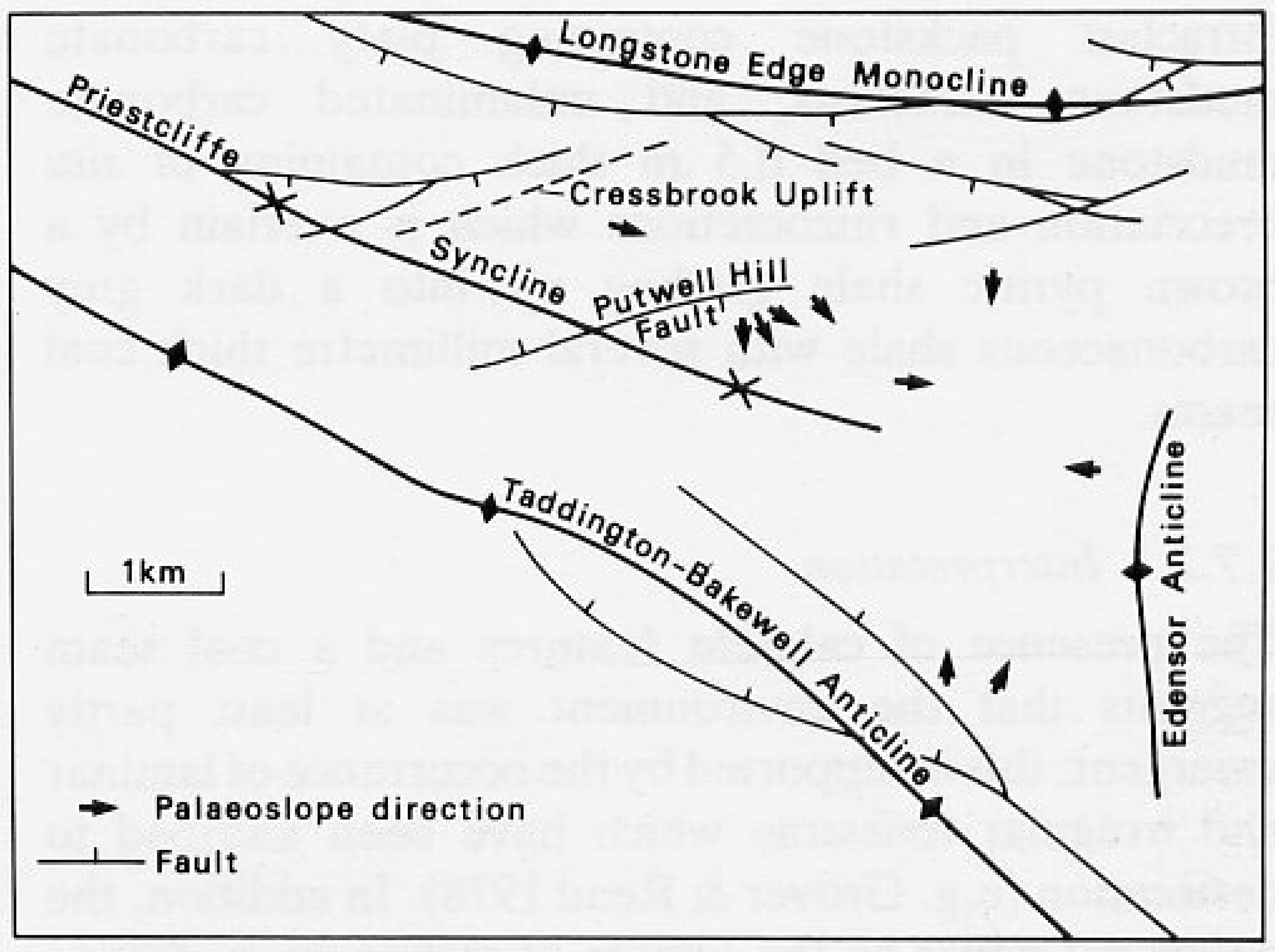

Fig. 5. Palaeoslopes within the intrashelf basin as inferred from slump structures.

intraclast packstone containing platy carbonate mudstone intraclasts, and unlaminated carbonate mudstone in a bed 0.5 m thick containing *in situ* brecciation and rhizocretions which is overlain by a brown pyritic shale grading up into a dark grey carbonaceous shale with several millimetre thick coal seams.

1.7.2. Interpretation

The presence of calcrete features and a coal seam suggests that the environment was at least partly emergent; this is supported by the occurrence of laminar and irregular fenestrae which have been ascribed to desiccation (e.g. Grover & Read 1978). In addition, the polygonal ridges on the surface of carbonate mudstone and dolomicrite laminae are interpreted as compacted desiccation curls formed by the drying out of algal mats (Ginsberg *et al.* 1977). The occurrence of tubular fenestrae, interpreted by Grover & Read (1978) as root moulds or burrows, in association with other types of fenestrae is suggested by Shinn (1983) to be an indicator of intertidal or supratidal conditions. The occurrence of terrestrial plants also suggests proximity to an emergent area. The coffin-shaped and hexagonal spar-filled moulds are interpreted to be pseudomorphs after evaporites. These pseudomorphs are distinct from those described by Adams & Cossey (1978) from the Rosewood Laminite which are of late diagenetic origin. Because the pseudomorphs are often partly compacted, the evaporites grew and dissolved whilst the surrounding sediment was still soft. The precipitation and dissolution of evaporites requires salinity to fluctuate between hypersaline and fresh (Butler 1969; Logan 1974). Conditions of abnormal salinity are also supported by the presence of the unusual bioclast assemblage in the grainstone laminae which are interpreted as the deposits of hypersaline or brackish intertidal ponds. Grainstones with marine bioclasts indicate a connection with marine conditions and were probably deposited as storm washover on to the intertidal areas. Chert is common in this facies and is of late diagenetic origin. The association between chert nodules and slump structures suggested by Butcher & Ford (1973) is probably fortuitous.

The Monsal Laminite (Fig. 8) is at least 2 m thick and consists of laminated carbonate mudstone with abundant tubular, laminar and irregular fenestrae with occasional ?*Renalcis* bafflestone. Using the criteria for recognising the degree of subaerial exposure of peritidal sediment proposed by Ginsberg *et al.* (1977) and Smosna & Warshauer (1981), the Monsal Laminite is interpreted as an intertidal or supratidal deposit with occasional intertidal ponds. The Monsal Laminite shows no evidence of soft sediment deformation and is probably *in situ*.

The Rosewood Laminite is 0.7 m to at least 1 m in thickness and consists of interlaminated carbonate mudstone and dolomitized grainstone which contain elements of the unusual bioclast assemblage and occasional bioclasts indicative of normal marine salinity. This laminite has been interpreted by Walkden (1970) as a deep-lagoonal deposit with alternating deposition of carbonate mud and dolomite due to seasonal climatic variation, and by Adams & Cossey (1978) as an offshelf storm laminite. These interpretations are rejected because dolomitisation post-dates cementation of the grainstone laminae. Thus the dolomite is not a primary deposit, and the rarity of marine bioclasts in the grainstone laminae make them unlikely to be reworked shallow-marine sediment. The Rosewood Laminite is interpreted as an intertidal deposit with intertidal ponds which received occasional storm wash-over from an adjacent marine area. This laminite shows abundant evidence of slumping, which indicates that it is allochthonous.

The Ashford Laminite is 0.03 m in thickness. It consists of carbonate mudstone laminae separated by partings or thin grainstone laminae containing elements of the unusual bioclast assemblage. This is interpreted as an intertidal deposit. The laminite shows evidence of soft sediment deformation and is allochthonous.

The Headstone Laminite is 2.5 m to 9 m in thickness; however, this laminite consists of several slump sheets, so there is some stratigraphical repetition and the original thickness is not known. It consists of laminated dolomicrite and carbonate mudstone containing irregular and laminar fenestrae and evaporite pseudomorphs. Grainstone laminae containing bioclasts indicative of restricted and normal marine salinity are present. A coal and seat-earth overlying a calcretised limestone are present at Locality 17. This laminite has been interpreted by Brown (1973) as a lacustrine deposit; however, the presence of marine bioclasts does not support this interpretation. The laminite was probably deposited in a variety of supratidal and intertidal flat and pond environments. The occurrence of the intraclast packstone at Locality 32 suggests reworking of the tidal flat, probably in an intertidal channel.

The Headstone Laminite is allochthonous, as a number of slump sheets can be identified. At localities 17 and 33 slumped units are overlain erosively by resedimented graded calcarenite containing intraclasts derived from the laminite. This association is interpreted as a turbidite generated by retrogressive slope failure following slumping (Section 1.5).

1.8. Spiriferid facies

1.8.1. Description

This facies is between 0.15 m and 0.3 m in thickness and it overlies the Headstone Laminite. It is a bioclastic packstone crowded with spiriferid valves. Other bioclasts include disarticulated crinoid ossicles, brachiopod spines and bryzoans. At localities 18 and 32 this facies is graded and all the spiriferids are disarticulated, whereas at Locality 17 the facies displays no internal grading with an apparently uniform distribution of spiriferid valves.

1.8.2. Interpretation

This facies reflects the re-establishment of marine conditions following peritidal sedimentation. The variation in degree of reworking probably reflects differing degrees of reworking in the shallow subtidal environment in different parts of the basin.

1.9. Nodular facies

1.9.1. Description

This is a bioclastic calcisiltite with a wackestone / carbonate mudstone texture. There are occasional discontinuous layers of whole, stacked, disarticulated brachiopod valves and crinoid ossicles. Occasional articulated brachiopods are preserved in life position with spines attached. The sediment is mottled but no distinct burrow forms are present. This facies has a well developed nodular bedding. The shape of these nodules ranges from isolated rounded nodules up to 0.1 m in thickness, with thickness-to-length ratios of 1:2 to 2:3 (Fig. 4e, f) to tabular and multilayered nodules. They occur in layers with a vertical spacing of between 0.1 m to 0.5 m. There is no textural difference between the nodules and the sediment between the nodules. There is no evidence of boring or encrustation of the surfaces of the nodules. Layers of stacked bioclasts can be traced through the margins of nodules showing that the margins of the nodules transgress bedding. The sediment between the nodules shows differential compaction around the nodules. This facies is cut by a syn-sedimentary listric normal fault with southward downthrow whose fault plane is marked by a zone of broken nodules, some of which may be refitted. The attitude of these nodules shows that some have been rotated within the fault plane.

1.9.2. Interpretation

The fine-grained sediment and the absence of reworking indicates deposition in a low-energy environment below wave-base. Layers of stacked bioclasts represent either influxes or concentrations due to winnowing. The limited bioturbation indicates a poorly oxygenated environment, although the occurrence of *in situ* brachiopods implies some oxygenation. The absence of encrustation or boring of the surface of nodules suggests that they formed beneath the sediment surface. Fracturing and rotation of nodules within the contemporary fault plane and the differential compaction around the nodules shows that they were lithified before the surrounding sediment. The cross-cutting relationship between the nodules and bedding also suggests that they were formed by early cementation of the sediment. This facies may be of similar origin to nodular chalks formed by early cementation of periplatform sediments described from the north slope of the Little Bahama Bank by Mullins *et al.* (1980, 1985). The presence of a syn-sedimentary fault with southward downthrow indicates deposition on a southward-dipping palaeoslope.

2. FACIES ASSOCIATIONS

A facies association comprises two or more facies which commonly occur together. Three associations are recognised, and their distribution is shown in Figure 8.

2.1. Association 1: Evenly bedded — resedimented carbonate association

The evenly bedded facies consists of fine-grained bioclastic sediment which was winnowed from surrounding shelf areas and deposited below wave-base in the intrashelf basin. Resedimented carbonates were deposited as low and high density turbidites generated by slumping or storm activity and were derived from surrounding shelf areas and intrabasinal highs. The restricted ichnofauna of this association suggests conditions of low oxygen levels, but the presence of occasional *in situ* corals and brachiopods in shales interbedded with turbidites suggests that the basin was at least periodically oxygenated. Reworked coral beds occur within this association.

2.2. Association 2: Resedimented carbonate – bioclastic wackestone/packstone – bioclastic grainstone/packstone – coral bed association

Resedimented carbonates in this association form a coarsening-upwards sequence in which the bed thickness increases upwards. This sequence is interpreted as progressively proximal storm deposition. The resedimented carbonates are overlain transitionally by the bioclastic wackestone/packstone facies which was deposited in a subtidal environment below wave-base but occasionally subject to storm reworking. The bioclastic grainstone/packstone facies was deposited above wave-base with occasional development of cross-bedded bioclastic grainstone shoals. The coral beds contain corals which are mainly preserved *in situ* and can be traced laterally into association 1 where the corals are reworked. This association is interpreted as a gradual shallowing of the basinal environment, probably as a result of progradation of a carbonate ramp.

2.3. Association 3: Nodular – bioclastic wackestone/packstone – bioclastic grainstone/packstone association.

The nodular facies is interpreted as periplatform bioclastic sediment deposited below wave-base in near-anoxic conditions. Direction of slumping shows that deposition took place on a southward dipping slope and the lack of evidence indicating rapid deposition or undercutting of the slope by scour suggests that this slump may have been seismically triggered. The bioclastic wackestone/packstone and bioclastic grainstone/packstone facies were deposited in shallower water below and above wave-base respectively. This association is interpreted as a shallowing of the intrashelf basin. The occurrence of slumping involving deep-water sediment suggests that this association may represent deposition on a distally steepened carbonate ramp.

3. BASIN INITIATION

The sequence of events associated with the initiation of the intrashelf basin can be inferred from the Millers Dale — Litton Mill area (localities 1 to 4). Detailed mapping of this area supports Walkden's (1977) interpretation of the events at the Asbian/Brigantian boundary as follows:

1. Sub-aerial exposure of the Bee Low Limestones.
2. Deposition of the Station Quarry Beds in a basin formed between the developing Longstone Edge Monocline and the Taddington — Bakewell Anticline. The Station Quarry Beds were deposited initially in a shallow subtidal environment with local emergent grainstone shoals built up above sea-level by reworking of subtidal sediments. The remainder of the Station Quarry Beds comprise the bioclastic wackestone/packstone facies which was deposited in a restricted low-energy subtidal environment. The full extent of this basin is not known as the occurrence of the Station Quarry Beds has not been demonstrated to the east. Walkden (1977) considered the Station Quarry Beds to be the marginal deposits of a basin which lay to the east.
3. Deposition of the Station Quarry Beds was terminated by uplift along an ENE-WSW trending structure coincident with the Lees Vein (Fig. 6). This axis is referred to as the Cressbrook Uplift. The Station Quarry Beds are overstepped along a line parallel to the Cressbrook Uplift (Fig. 6) which resulted in further subaerial exposure of the underlying Bee Low Limestones producing karstic pits up to 2 m deep (e.g. at Locality 2, Fig. 7).
4. Extrusion of the Upper Millers Dale Lava took place in two episodes. The second flow was correlated by Walkden (1977) with a bentonite clay approximately 7 m above the base of the Monsal Dale Limestones. The two episodes of lava extrusion were thus separated by deposition of the limestones underlying the bentonite.
5. The bioclastic grainstone/packstone facies was deposited in the form of a wedge draped over the flow front of the second lava flow (Fig. 7). The coarse intraclastic and bioclastic nature of this sediment, which contains oncolites, indicates that it is a high-energy sediment. Grading and stacking of bioclasts within the wedge suggests that the sediment was transported. The wedge is interpreted as bioclastic sediment produced in a shallow water environment over the lava flow, then washed off the lava and deposited in a wedge banked up against the flow front.

The Upper Millers Dale Lava (Fig. 6) is absent over the Cressbrook Uplift, which is interpreted to have formed a positive feature around which the lava flowed.

Figure 7 shows a facies change in limestones overlying the Cressbrook Uplift together with an interpretation of their environments of deposition. The bioclastic grainstone/packstone facies overlies the crest of the Cressbrook Uplift, and the bioclastic wackestone/packstone facies and the evenly bedded facies occupy progressively distal positions down the flanks of the Uplift. Facies boundaries are parallel to the Uplift. This is interpreted as a transition from shallow, high-energy conditions over the crest of the Uplift to progressively deeper, lower energy and near-anoxic conditions down the flank. The transition from the bioclastic grainstone/

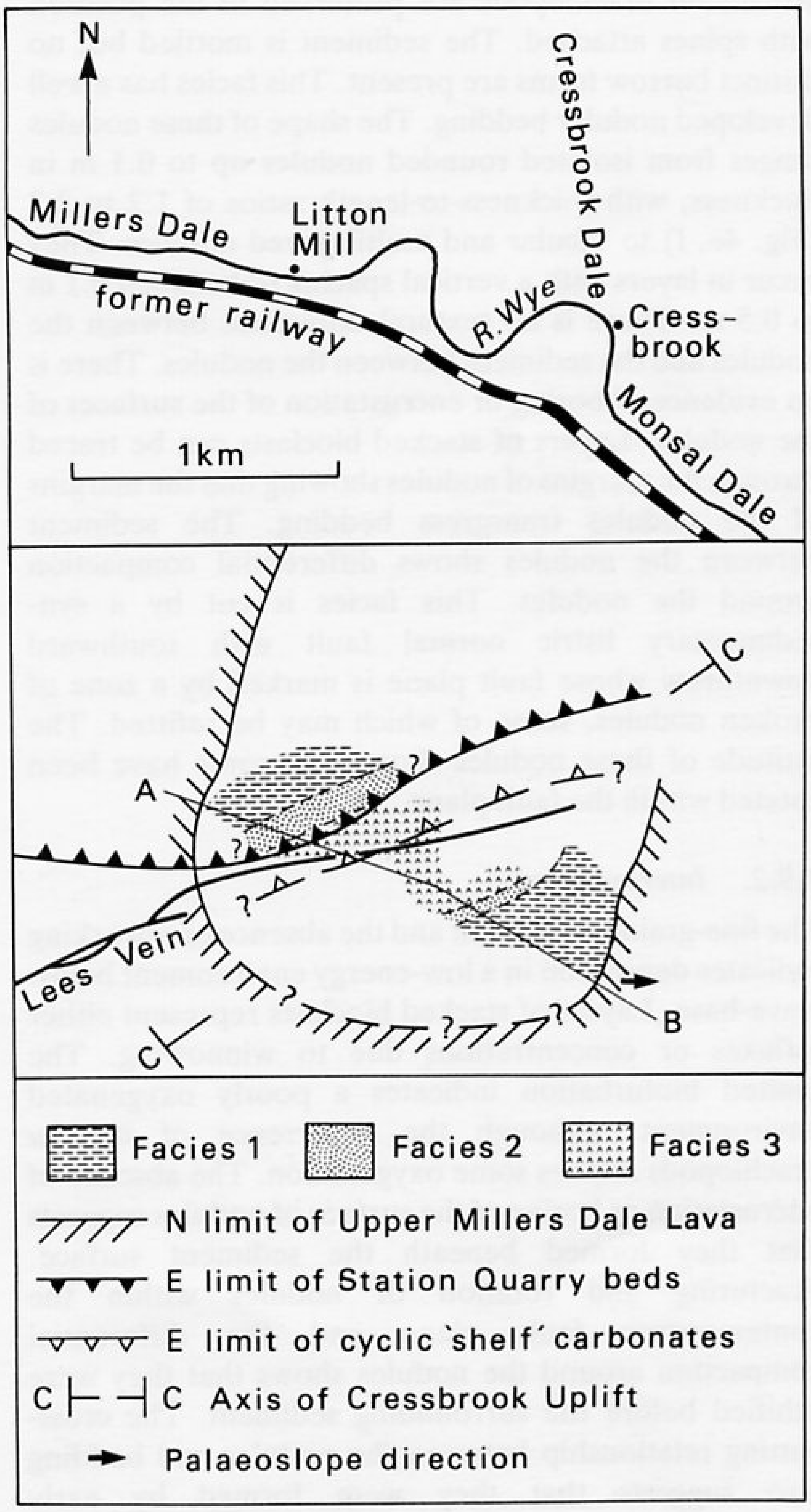

Fig. 6. Distribution of the Upper Millers Dale Lava and facies in limestones overlying the Cressbrook Uplift, showing the influence of the uplift on sedimentation: A-B. line of section shown in Figure 7.

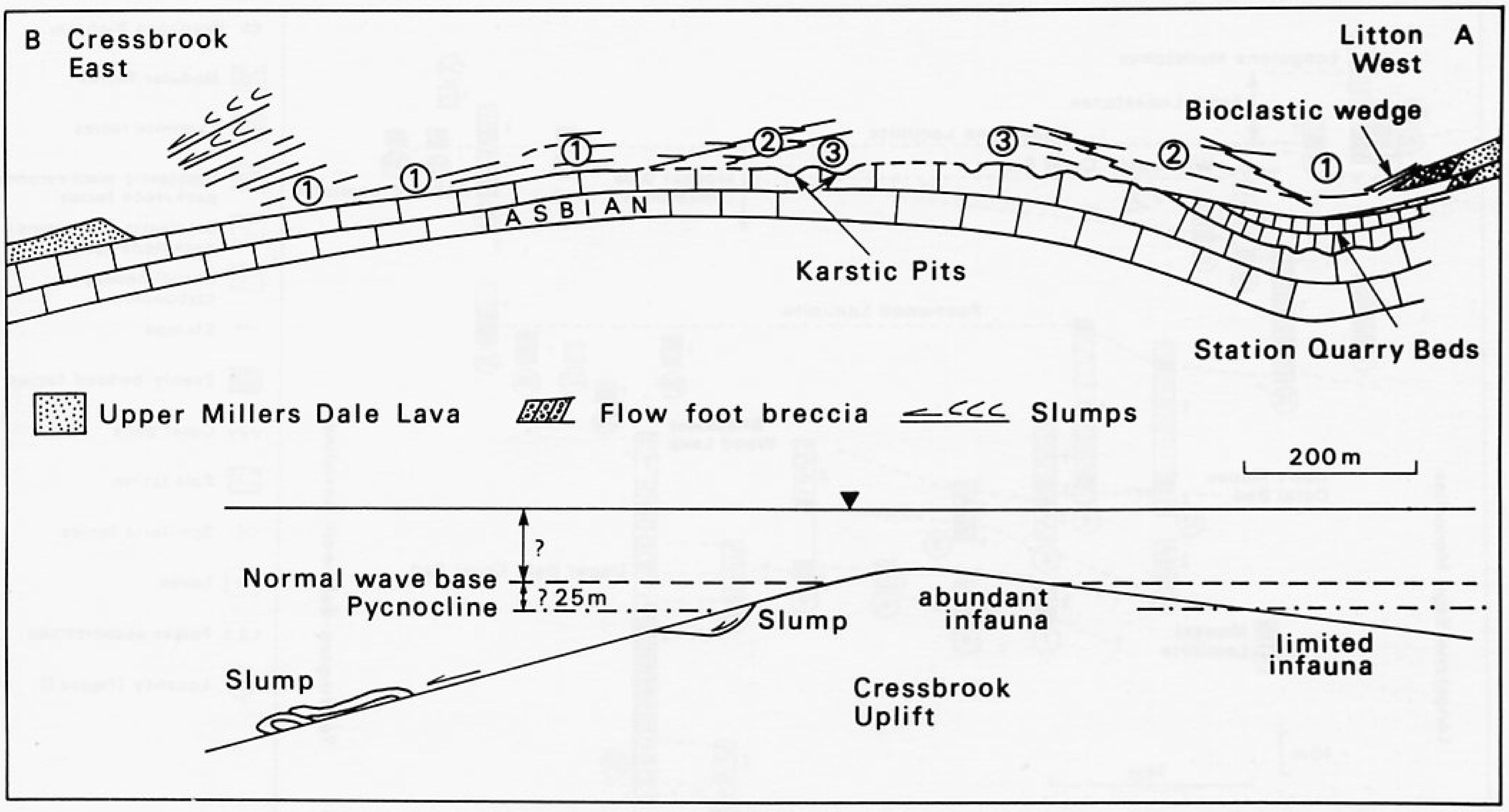

Fig. 7. Measured section across the Cressbrook Uplift showing facies change and the interpreted environment of deposition: 1. Evenly bedded facies; 2. Bioclastic wackestone/packstone facies; 3. Bioclastic grainstone/packstone facies; line of section is shown in Figure 6.

packstone facies to the bioclastic wackestone/packstone facies is attributed to the intersection of normal wave-base with the sea floor; the transition from the bioclastic wackestone/packstone facies to the evenly bedded facies is attributed to the intersection of the pycnocline with the sea floor (Fig. 7).

The Cressbrook Uplift sheltered and partly isolated the western part of the intrashelf basin. The vertical transition from the restricted sediments of this partly isolated basin to the overlying pale facies shows that this basin was later infilled by progradation of surrounding shelf carbonates. The transition from the pale facies to the bioclastic wackestone/packstone and evenly bedded facies of the main part of the intrashelf basin takes place across the Cressbrook Uplift; this suggests that the Uplift formed the western margin of the intrashelf basin during at least part of the Brigantian (Fig. 6).

4. BASINAL SEDIMENTATION

The distribution of the facies and facies associations described in sections 1 and 2 is shown by Figure 8. Facies Association 1 is the most common association suggesting that deposition in the basin was dominated by periplatform sediments winnowed from the surrounding shelf and resedimented bioclastic carbonates sourced from the surrounding shelf area and intrabasinal highs.

Palaeoslopes (Fig. 5) show an association with present-day structures, including the Lees Vein (coincident with the Cressbrook Uplift), Longstone Edge Monocline, Taddington — Bakewell Anticline and Edensor Anticline. This association suggests that precursors of these structures may have influenced sedimentation within the basin.

The site of the intrashelf basin is thought to have been marked by a faster rate of subsidence relative to the surrounding shelf areas, rather than a marked difference in water depth. The following lines of evidence provide a minimum estimate of the water depth in the basin:

1. Facies Association 2 is interpreted as a transition from deep sub-wave-base sediments deposited above wave-base due to deposition on a carbonate ramp. The thickness of this association is, thus, a minimum estimate of water depth in the basin. The best exposed example has a thickness of 15 m from the top of the evenly bedded facies to the base of the bioclastic grainstone/packstone facies.
2. The facies change across the Cressbrook Uplift (Fig. 7) provides an estimate of the depth between wave-base and the pycnocline as this is a function of the outcrop width of the bioclastic wackestone/packstone facies and the depositional slope. Assuming the dips of limestones down the flanks of the Cressbrook Uplift are maximum values for the depositional slope (7°) and the outcrop width is 200 m, then the maximum depth between wave-base and the pycnocline was 25 m.

The development of near-anoxic conditions in this basin reflects the sheltered position of an intrashelf basin. Wave and current activity would have been strongly damped in the surrounding shallow-shelf areas producing a high wave-base over the intrashelf basin. This would have reduced the effectiveness of vertical

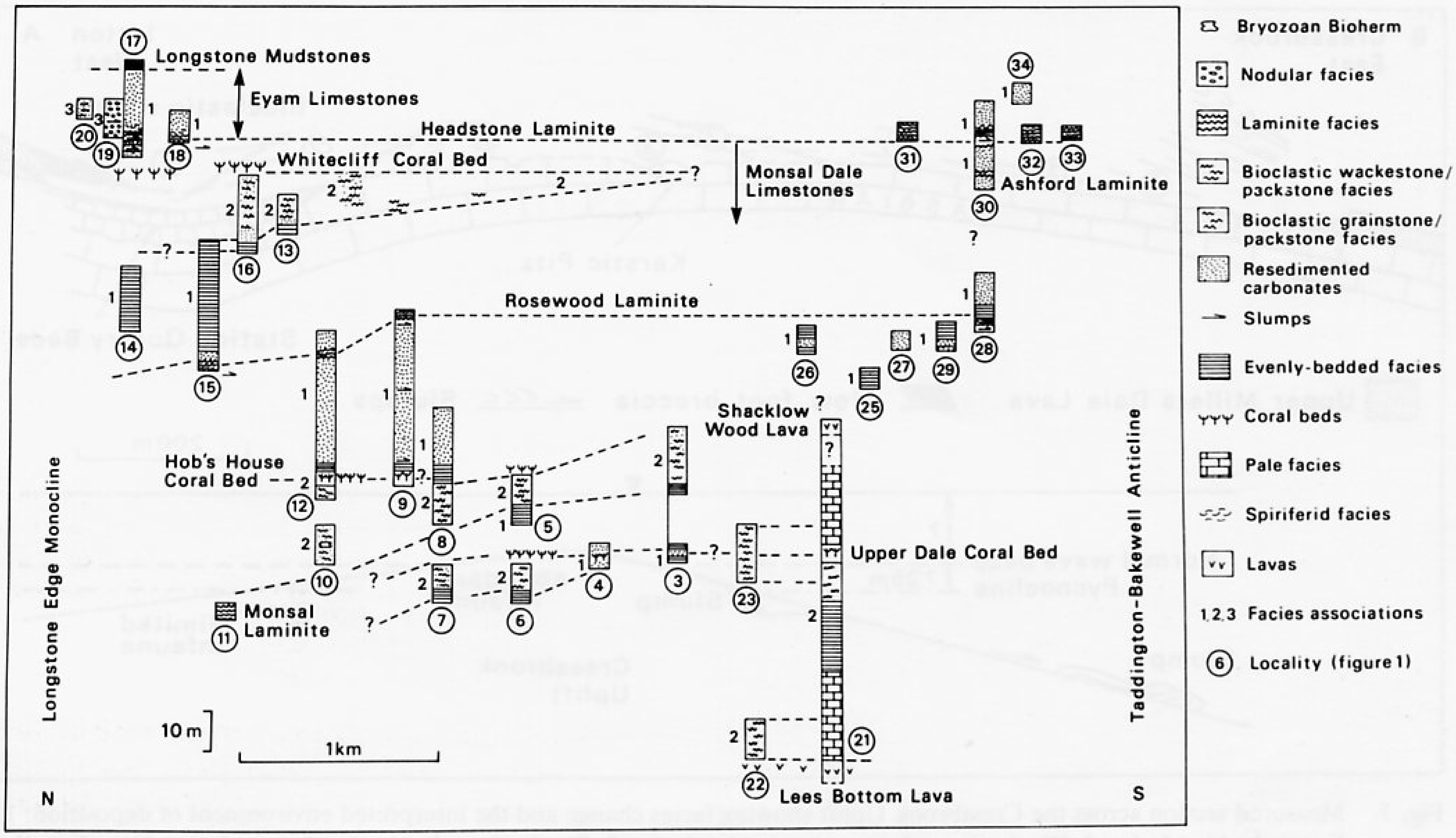

Fig. 8. Distribution and correlation of facies within the intrashelf basin.

mixing in the water column, promoting the development of shallow anoxic conditions (Byers 1977).

The interpretation of the laminite facies as a peritidal deposit implies the intrashelf basin was periodically emergent. There is no evidence of shallowing in the sediments underlying the laminite facies, and so the emergence was not caused by progradation. The Headstone Laminite was deposited at a time when much of the surrounding shelf area was emergent (Gutteridge 1983). The absolute depth of the intrashelf basin is not known but is thought to be at least 25 m. A drop in base level by this amount is more likely to be due to eustatic regression rather than tectonic uplift. The laminite facies occurs at four levels in the succession, which indicates that four regressions of sufficient magnitude to expose the intrashelf basin took place during the Brigantian.

The presence of volcanic units in the succession might suggest that sedimentation was influenced by vertical movement associated with near-surface movement of magma (e.g. Le Bas 1980), or by the development of topographic features built up by extrusive bodies. Geochemical work by MacDonald *et al.* (1984) on the Derbyshire volcanics indicates that there was insignificant fractionation of magma shallower than 45 km, suggesting that shallow magma chambers were unlikely to have been present. This makes the possibility of vertical movements associated with volcanism unlikely. Extrusion of the Upper Millers Dale Lava within the basin created a topographic high which formed an area of shallow-water sedimentation (Section 3). It is not, however, possible to demonstrate any topographic influence on sedimentation by other lavas owing to poor exposure.

5. NATURE OF THE BASIN MARGINS

5.1. The northern margin

Figure 8 shows that Facies Association 2 occurs at three levels within the succession; it thickens towards the E-W trending Longstone Edge Monocline (Fig. 1) and thins

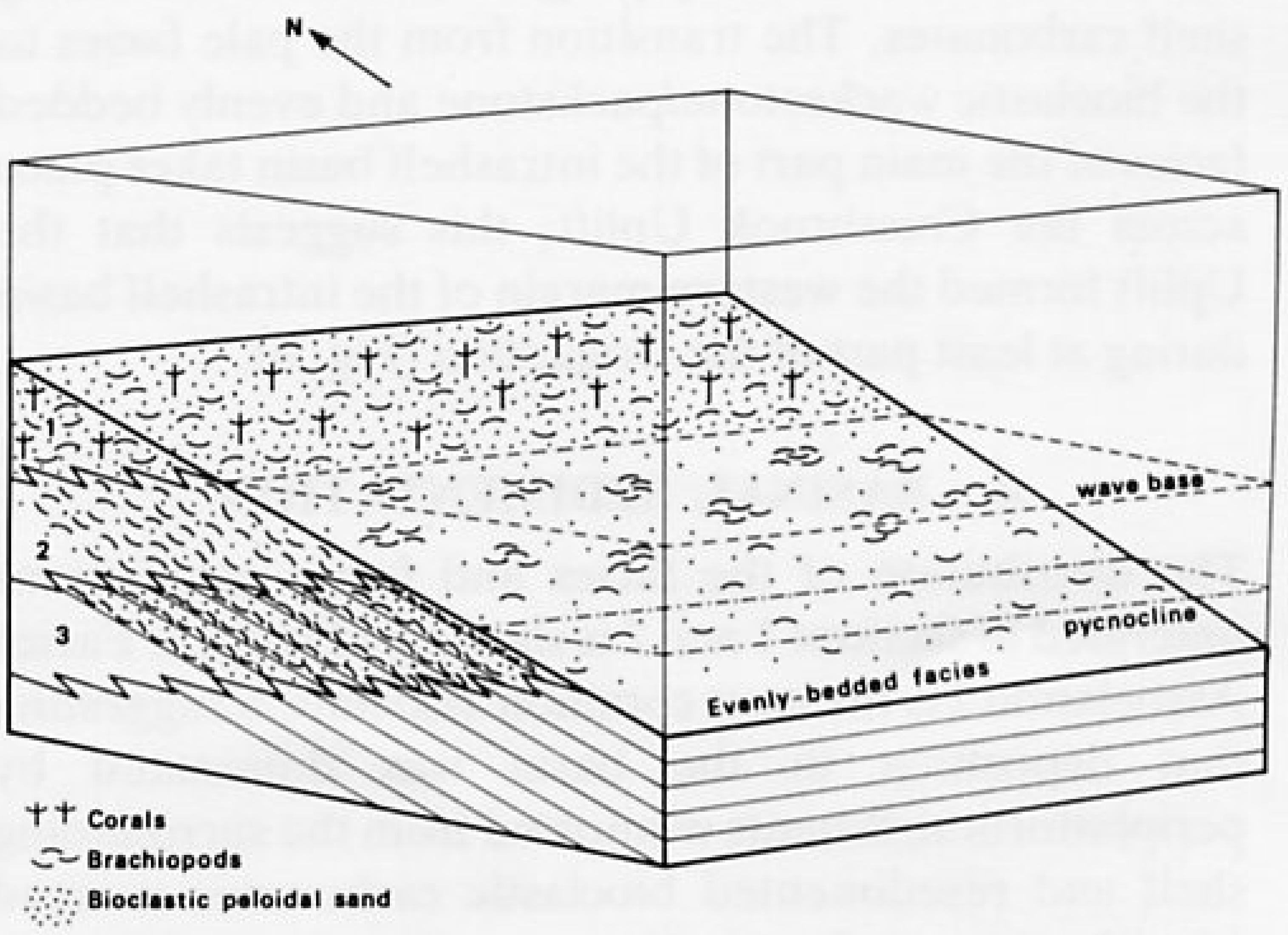

Fig. 9. Sedimentary model of southwards-dipping carbonate ramp at the northern margin of the intrashelf basin when the Longstone Edge Monocline was inactive; 1. Bioclastic grainstone/packstone facies; 2. Bioclastic wackestone/packstone facies; 3. Resedimented carbonates.

and pinches out to the south. Stevenson & Gaunt (1971) and Brown (1973) showed that, to the north of the Longstone Edge Monocline, the Monsal Dale Limestones are similar to the pale facies whose depositional environment has been interpreted as a carbonate shelf (Section 1.1). The northern margin of the basin is thus interpreted as a southward dipping carbonate ramp (Fig. 9).

Facies Association 2 is abruptly overlain by Facies Association 1 indicating that the rate of subsidence outpaced sedimentation (otherwise a deepening sequence overlying facies association 2 would have been preserved). The cause of progradation followed by abrupt deepening is not known but may be due to variation of subsidence rate in the basin. There is no evidence of growth of the Longstone Edge Monocline during deposition of the Monsal Dale Limestones. No slumping has been found; also Aitkenhead *et al.* (1985, p.26) showed that the trend of the Longstone Edge Monocline does not influence the isopachs of the Monsal Dale Limestones. However, these isopachs are poorly constrained as there is no complete sequence of the Monsal Dale Limestones exposed elsewhere in the area of the intrashelf basin.

During deposition of the Eyam Limestones, the northern margin of the intrashelf basin was marked by Facies Association 3, interpreted as a southward dipping, distally steepened carbonate ramp (Fig. 10). Growth of the Longstone Edge Monocline caused southwards slumping of the nodular facies down the southern flank of the Longstone Edge Monocline.

5.2. The southern margin

The sedimentological evolution of the southern basin margin is poorly constrained owing to poor exposure; however, the facies distribution (Fig. 8) suggests that sedimentation occurred in two stages:

1. Detailed mapping around Locality 21 shows a lateral transition across the northern flank of the NW-SE trending Taddington-Bakewell Anticline (Fig. 1) from the pale facies, interpreted as shelf limestones, to the bioclastic wackestone/packstone and evenly bedded facies which are interpreted as basinal limestones. An interfingering of these facies over the northern flank of the Taddington — Bakewell Anticline also occurs (Fig. 8). This is interpreted as a result of periodic progradation and retreat of shelf conditions across the northern flank of the Taddington — Bakewell Anticline. N. J. D. Butcher (per. comm. 1987) has also recorded a thickening of part of the Monsal Dale Limestones down the northern flank of the Taddington — Bakewell Anticline. These facies relationships are interpreted to be a result of periodic growth of the Taddington — Bakewell Anticline which controlled the southern margin of the intrashelf basin.
2. Later, during the deposition of the Monsal Dale Limestones and the Eyam Limestones, the occurrence of facies association 1 overlying the northern flank of the Taddington — Bakewell Anticline shows that deeper-water conditions were established over the anticline. This implies that the Taddington — Bakewell Anticline no longer controlled the southern margin of the intrashelf basin, although there was a northward- or north-westward-dipping palaeoslope in this part of the basin, indicated by slumping in the Rosewood and Headstone laminites (Fig. 5). This change in sedimentation may reflect the development of an eastward-dipping carbonate ramp in the central part of the Derbyshire Dome described by Gutteridge (1983, 1984). This change in sedimentation, however, occurred during deposition of the Monsal Dale Limestones, whereas, according to Gutteridge (1983), the ramp developed at the beginning of deposition of the Eyam Limestones. Currie (1987) showed that the carbonate ramp in the central part of the Derbyshire Dome may have developed during deposition of the Monsal Dale Limestones, and that the occurrence of dark limestones, similar to the evenly bedded facies in Lathkill Dale some 3 km to the south of the Taddington — Bakewell Anticline shows that basinal conditions periodically extended to the south of this anticline.

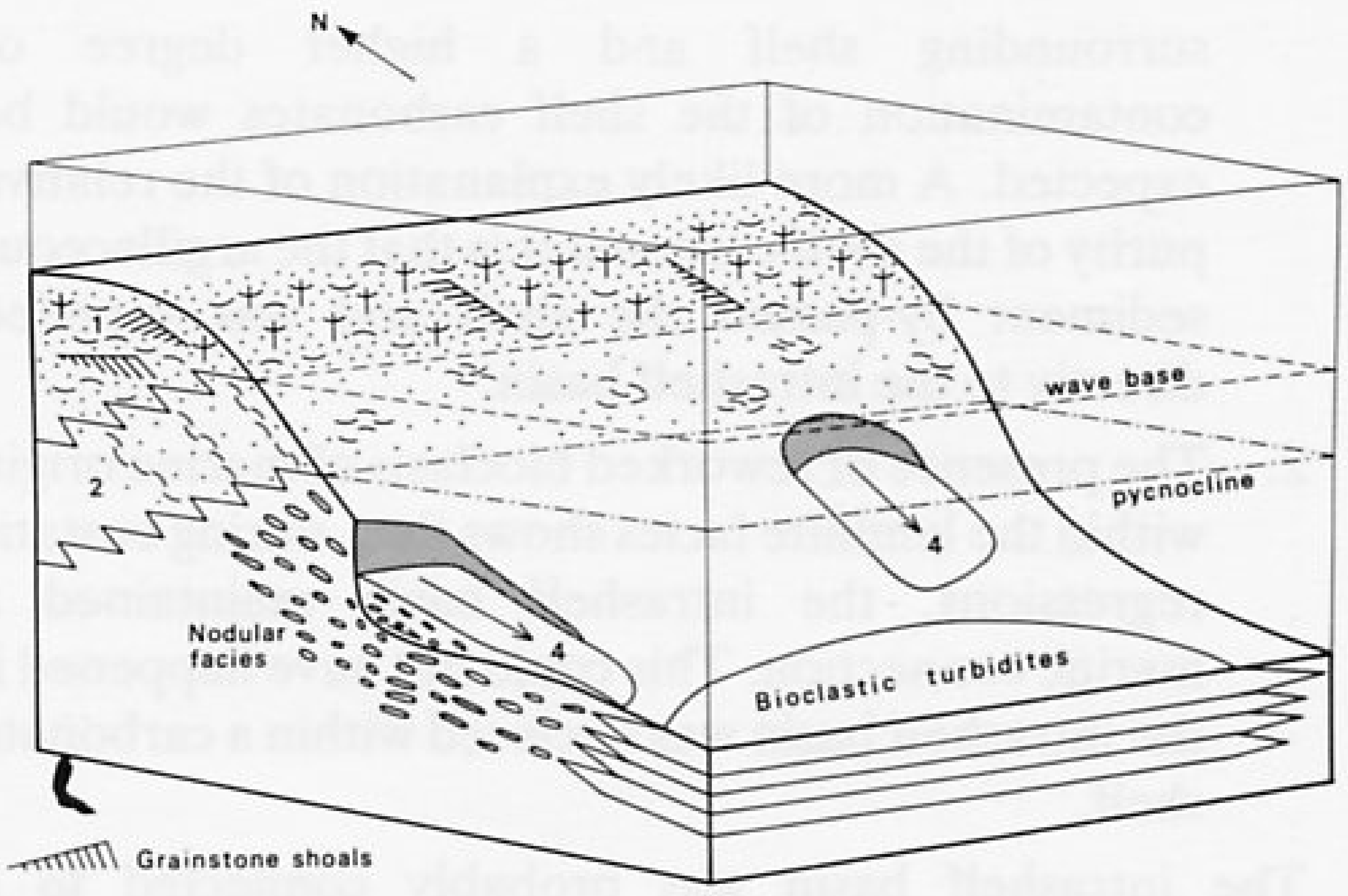

Fig. 10. Sedimentary model of southwards-dipping distally steepened carbonate ramp at the northern margin of the intrashelf basin when the Longstone Edge Monocline was active; 4. slumping of the nodular facies; symbols and abbreviations as in Figure 9.

5.3. The eastern margin

Gutteridge (1987) depicted this intrashelf basin as being entirely enclosed by carbonate shelf; however, the following lines of evidence suggest that the intrashelf basin may have been connected to a larger basinal area:

1. Basinal carbonates are interbedded with shales and are argillaceous, whereas the surrounding shelf carbonates are relatively pure, having insoluble residue contents of 1% to 5% (Cox & Bridge 1977). If the basin were enclosed, the argillaceous sediment would have been transported across the

surrounding shelf and a higher degree of contamination of the shelf carbonates would be expected. A more likely explanation of the relative purity of the shelf carbonates is that the argillaceous sediment by-passed the shelf and was supplied directly to the intrashelf basin.

2. The presence of reworked bioclasts of marine origin within the laminite facies shows that, during eustatic regressions, the intrashelf basin maintained a marine connection. This could not have happened if the intrashelf basin was enclosed within a carbonate shelf.

The intrashelf basin was probably connected to a southern extension of the Edale Basin via the eastern or south-eastern part of the outcrop. The N-S trending Edensor Anticline (Fig. 1) controlled the eastern margin of the intrashelf basin at least during the late Brigantian. Aitkenhead *et al.* (1978) mapped a facies change from basinal to shelf carbonates over the crest of the Edensor Anticline. A carbonate mud mound which grew by lateral accretion is present in the Eyam Limestones over the crest of the Edensor Anticline. This style of mud-mound growth developed in shallow water (Gutteridge 1983). Slumping of the Headstone Laminite indicates a westwards-dipping palaeoslope down the western flank of the Edensor Anticline (Fig. 5).

The stratigraphical relationship of the Dinantian limestones to the Longstone Mudstones and Namurian over the Edensor Anticline (Fig. 11) provides evidence of growth of the Edensor Anticline during the late Brigantian. The Eyam Limestones are overstepped locally and the Longstone Mudstones or Namurian shales rest on the Monsal Dale Limestones. This unconformity coincides with the axis of the Edensor Anticline and the Longstone Edge Monocline, indicating that both these features were active after deposition of the Eyam Limestones but before deposition of the Longstone Mudstones. A minor N-S trending anticline (Haddon Fields Anticline) lies on the southern extension of the Edensor Anticline (Fig. 11). A facies change in the Eyam Limestones associated with the Haddon Fields Anticline shows that this was also active during the late Brigantian (Gutteridge 1983).

6. BASIN DEVELOPMENT

6.1. Summary of sedimentation

1. The intrashelf basin was initiated at the beginning of the Brigantian by subsidence between the Taddington — Bakewell Anticline and the Longstone Edge Monocline. This produced a basin in which the Station Quarry Beds were deposited.
2. Growth of the Cressbrook Uplift caused erosion of the Station Quarry Beds and resulted in an angular unconformity at the base of the Monsal Dale Limestones which oversteps the Station Quarry Beds.
3. During deposition of the Monsal Dale Limestones, the Cressbrook Uplift formed an intrabasinal high upon which shallow, reworked sediments were deposited. This intrabasinal feature caused partial isolation of the western part of the intrashelf basin which was later infilled by progradation of surrounding shelf sediments. The Cressbrook Uplift subsequently formed the western margin of the intrashelf basin.
4. Sedimentation in the main part of the intrashelf basin was by deposition of periplatform carbonates and resedimented bioclastic sediments sourced from surrounding shelf areas and intrabasinal highs.

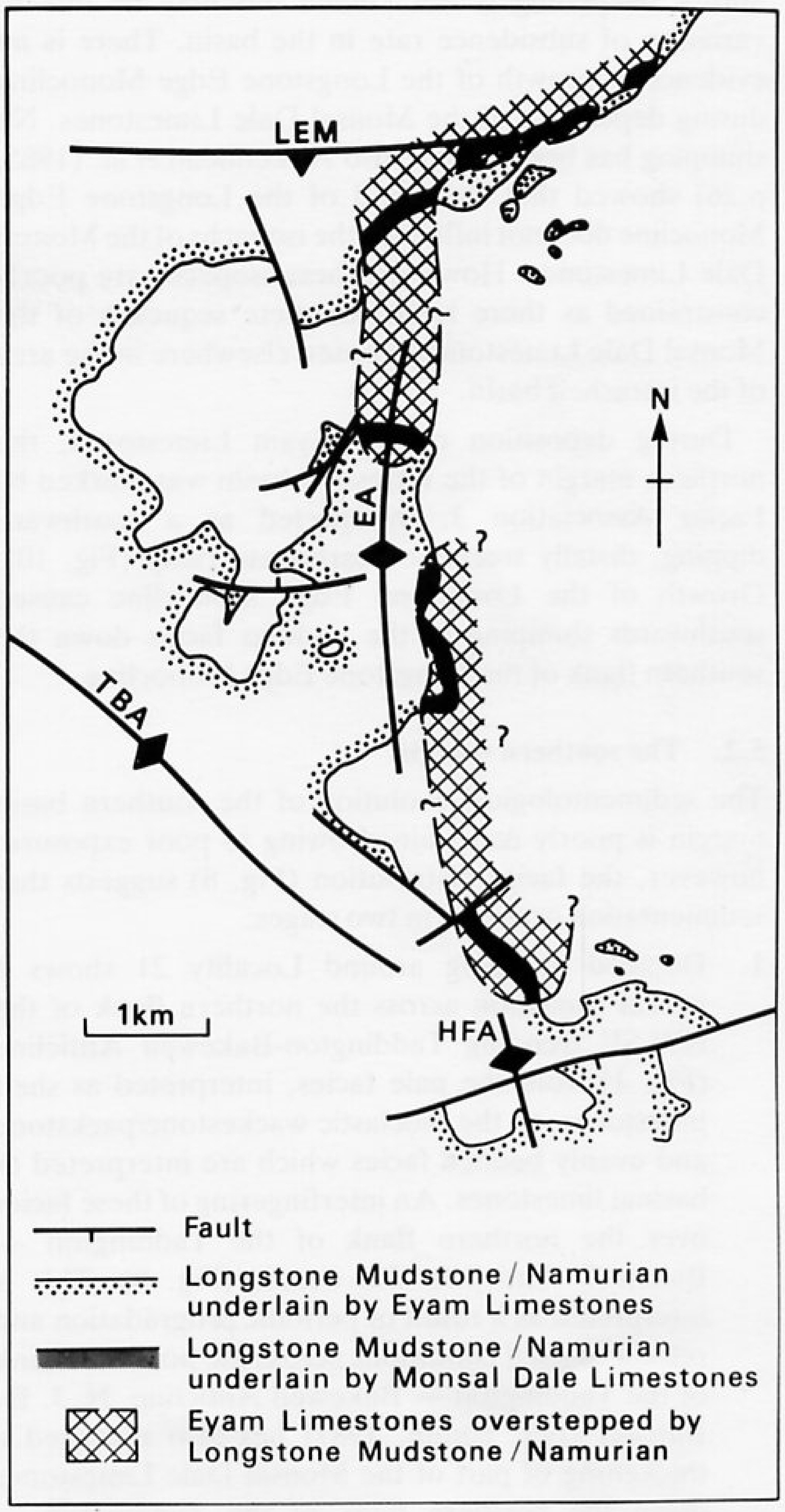

Fig. 11. Stratigraphical relationships between the Dinantian limestones and the Longstone Mudstones and Namurian shales at the eastern margin of the intrashelf basin: EA. Edensor Anticline; HFA: Haddon Fields Anticline; LEM. Longstone Edge Monocline; TBA. Taddington — Bakewell Anticline.

Conditions within the basin were anoxic or marginally oxic. Peritidal sedimentation developed within the basin four times owing to eustatic regressions.

5. During deposition of the Monsal Dale Limestones, the northern basin margin was formed by a southward dipping carbonate ramp which prograded southwards, probably in response to variations of subsidence rate within the basin. There is no evidence of growth of the Longstone Edge Monocline during deposition of the Monsal Dale Limestones. During the deposition of the Eyam Limestones the Longstone Edge Monocline was active and the northern basin margin was formed by a southward dipping, distally steepened carbonate ramp.
6. The southern basin margin was controlled by the Taddington — Bakewell Anticline during the early Brigantian. This margin migrated southwards during the Brigantian, probably as a response to the development of an eastward dipping carbonate ramp in the central part of the Derbyshire Dome.
7. During the early Brigantian, the intrashelf basin was connected to a southern extension of the Edale Basin to the east. Later in the Brigantian the eastern margin of the intrashelf basin was controlled by the Edensor Anticline.

6.2. Structural setting of the intrashelf basin.

This study has suggested a way of resolving some of the apparent contradictions between sedimentological and geophysical data that were pointed out by Gutteridge (1987).

The basement structure underlying the central part of the Derbyshire Dome was inferred by Gutteridge (1987, fig. 8) to be either a fault terrace with progressive northward downthrow, or a tilt block. Bouguer gravity data in the central part of the Derbyshire Dome shows a northward declining gradient (Fig. 12) which is incompatible with the tilt-block model which requires a southward dipping basement surface. Sedimentological evidence suggests that the southern margin of the intrashelf basin was controlled initially by the Taddington — Bakewell Anticline, but later migrated southwards following the development of a carbonate ramp in the central part of the Derbyshire Dome. This is explained by suggesting that the southern basin margin was originally controlled by a fault in a similar position to the Bakewell Fault of Smith *et al.* (1985) which underlies the Taddington — Bakewell Anticline. Further extension across the basin resulted in the footwall collapse (Gibbs 1984) of the southern basement block, which in turn caused the basin margin to migrate southwards, and produced the fault terrace in the basement extending southwards to the Bonsall Fault.

The results of this study can also place constraints on the basement structure at the eastern side of the Derbyshire Dome. There is a negative Bouguer gravity anomaly (the Bakewell Low of Aitkenhead *et al.* 1985) to the east of the intrashelf basin (Fig. 12). Gravity modelling across this feature by A. G. Lee (pers. comm. 1987) suggests there is a major N-S trending basement fault (Ashford Fault) with easterly downthrow (Fig. 12). The Edensor Anticline overlies the Ashford Fault and it is suggested that this anticline formed as a result of inversion (slip reversal) of the Ashford Fault during the late Brigantian. Inversion of contemporary structures during the late Brigantian has been described from other Dinantian basins in northern England and North Wales by Gawthorpe *et al.* (this volume).

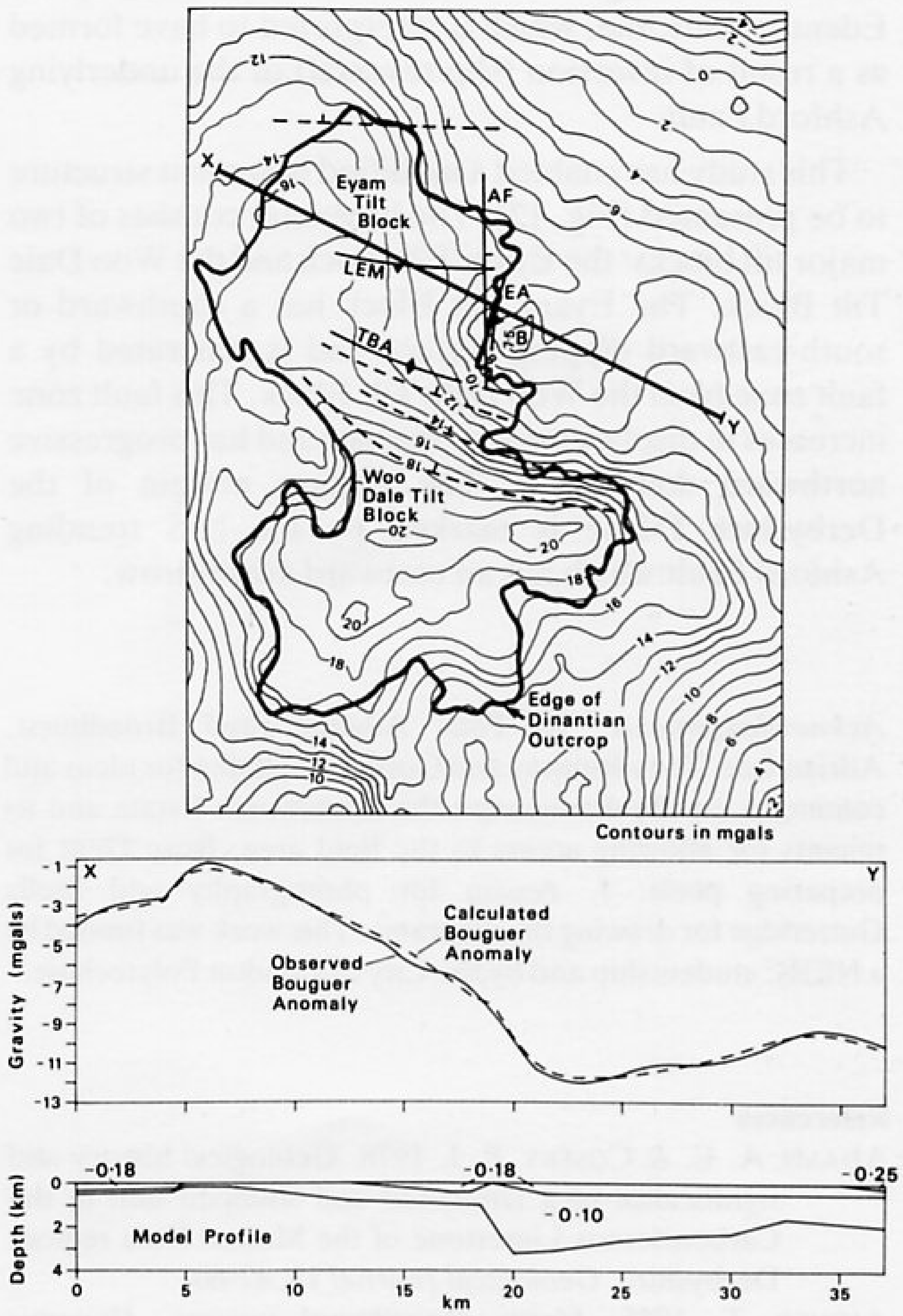

Fig 12. Basement structure determined by gravity modelling across intrashelf basin; line of model is shown on Bouguer Anomaly map. Densities reduced to a basement density of 2.8gcm^{-3} (Maroof 1976). Density contrasts as follows: Dinantian carbonates –0.1gcm^{-3}, Namurian and mud-dominated Dinantian –0.18gcm^{-3}, Westphalian –0.25gcm^{-3}. Geologically controlled regional of 18 mgal and gradient of 0.06 mgal per km subtracted from observed Bouguer anomaly. Revised pre-Dinantian structure is shown on Bouguer gravity map: B. Bakewell Gravity Low; AF. Ashford Fault; LEM. Longstone Edge Monocline; EA. Edensor Anticline; TBA. Taddington — Bakewell Anticline.

7. CONCLUSIONS

An intrashelf basin developed over part of the Derbyshire carbonate platform at the Asbian/Brigantian boundary due to extension which caused reactivation of basement faults. The southern margin of the intrashelf basin was determined initially by the growth of the Taddington — Bakewell Anticline. Further extension across the basin caused footwall collapse of the underlying basement fault block which was expressed at the surface by southward migration of the basin margin and the development of a carbonate ramp in the central part of the Derbyshire Dome. The northern margin of the basin was initially an homoclinal ramp which later became a distally steepened ramp with the development of the Longstone Edge Monocline. The eastern margin of the basin was controlled by the Edensor Anticline, which is interpreted to have formed as a result of inversion (slip reversal) of the underlying Ashford Fault.

This study has enabled a modified basement structure to be presented (Fig. 12). The basement consists of two major tilt blocks: the Eyam Tilt Block and the Woo Dale Tilt Block. The Eyam Tilt Block has a southward or south-eastward dipping surface and is separated by a fault zone from the Woo Dale Tilt Block. This fault zone increases in displacement to the east and has progressive northward downthrow. The eastern margin of the Derbyshire Dome is marked by the N-S trending Ashford Fault which has an eastward downthrow.

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P. GUTTERIDGE, Ph.D.
School of Earth Sciences
Thames Polytechnic
Walburgh House
Bigland Street
London E1 2NG

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