Sand volcano. Ross, near Loop Head.

SYNDEPOSITIONAL SLIDING AND SLUMPING IN THE WEST CLARE NAMURIAN BASIN, IRELAND

By
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FOREWORD

Almost half the surface area of Ireland is underlain by rocks of Carboniferous age but nowhere are these rocks better exposed than in County Clare. Over the years the area has therefore attracted the attention of many geologists. Professor Gill began his researches on the Namurian rocks of County Clare in the mid-1950s whilst holding the Chair of Geology and Mineralogy at Trinity College, Dublin. After. his appointment to the Chair of Petroleum Geology in Imperial College, London he frequently returned to Ireland, together with his research students, to continue work in County Clare and to initiate sedimentological studies elsewhere in Ireland. The paper which he published in association with Professor Ph. H. Kuenen of Leiden in 1958 on the sand volcanoes of County Clare was the earliest detailed description of this sedimentological phenomenon. This paper has made the County Clare coastline a classic area for the study of both sand volcanoes and their associated slump sheets. Professor Gill's field demonstration of the County Clare phenomena, to his students and to varied groups of Irish and overseas geologists, have always been marked for their enthusiasm and spirited discussions, as well as for the leader's detailed grasp of the subject.

The syndepositional slump features in County Clare are associated with an ancient deltaic environment and are clearly of significance as a model for petroleum geologists whose search for oil is often in similar but more recent deltaic deposits. It is with great pleasure that the Geological Survey of Ireland now publishes Professor Gill's monograph on the Clare slump sheets. It incorporates the results of more than twenty years investigation by Professor Gill and his research students, and must be regarded as the definitive study of his aspect of the geology of County Clare. The first part of the work comprises a classification and analysis of slump phenomena in general. This, together with the remarkable insights into slump generation afforded by the unique three-dimensional coastal exposures in County Clare, comprises a work of more than local importance which will be of value to all students of sedimentary rocks and to practising petroleum geologists.

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Syndepositional Sliding and Slumping in the West Clare Namurian Basin, Ireland

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INTRODUCTION

PREVIOUS WORK

Western County Clare is a tectonic basin of Namurian (E_1-R_2) shales, siltstones and sandstones, with a maximum thickness of about 1,525 m (5,000 feet), cut somewhere along its centre by the west-facing coastline (see Enclosure, Figure 1). The Namurian measures sit without angular discordance, and probably without great discontinuity, on Carboniferous Limestone (Clarke, 1966). The clastic Namurian sediments constitute an easterly prograding succession of euxinic, black, goniatite-bearing shales succeeded by parallel-bedded turbidite sandstones (E_1-R_2), and capped by a deltaic cyclothemic sequence (R_1-R_2). The area was mapped by the Irish Geological Survey between 1860 and 1862 under the direction of F. J. Foot (Foot, 1863), and published in Irish Geological Survey one inch to one mile sheets 122, 123, 131, 132, 140, 141 and 142.

At that time, the beds above the Carboniferous Limestone were considered to be Coal Measures. At a later date, Wheelton Hind (Hind, 1905) assigned them to the Pendleside Series on the evidence of the fossils collected, and indicated that there were no coal measures present. The beautiful goniatite faunas in the Clare Shale Formation at the base of the succession attracted the attention of Professor F. Hodson, who in 1954 established a detailed faunal and lithological succession in northwest County Clare (Hodson, 1954A), and in Foynes Island in the Shannon estuary, County Limerick (Hodson, 1954B). The author visited the area in 1953 with the prime objective of establishing the lithological and faunal succession, and recognised that the area presented a remarkable museum of syndepositional disturbances. Many geologists of extensive international experience, including Professors J. F. M. de Raaf, L. U. de Sitter, P. H. Kuenen, R. M. Shackleton, and R. Trumpy, have visited the area, and have voiced the opinion that these exposures probably represent the best models of this type of phenomena that have yet been described. The area was re-mapped in detail on the initial scale of 6 inches to the mile, and elaborated on various engineering scales by detailed maps of the slump structures. A small-scale reduction of this mapping project is presented in Figure 1 (Enclosure). A summary of progress was presented in 1958 (Brindley and Gill, 1958), and this account presented the overall stratigraphic picture, and recorded the evidence for extensive syn-sedimentary slumping in the sequence. Hodson and Lewarne (1961) presented a modified version of Hodson's (1954) succession and stratigraphy and applied it to the entire Namurian outcrop belt through counties Clare, Limerick, and parts of Cork. They found a variable thickness of the basal Clare Shale Formation (E_1-R_1), and indicated the possibility of a non-sequence between the shales and the limestones in northwest Clare. The extent of this non-sequence has been more recently contested on palaeontological grounds by Clarke (1966). The author, with Professor P. H. Kuenen (Gill and Kuenen, 1958), described the remarkable profusion of sand emissions with a volcano-like structure which occur on the top of syndepositional slides and slumps at many localities.

In 1969 Dr. M. H. Rider completed, under the author's direction, a study on the sediments of the West Clare Basin and compared them in detail with structures in sedimentary environments which he had studied and documented in the modern Mississippi Delta. This Ph.D. thesis (Rider, 1969) shows clearly that, particularly in the cyclothemic rocks, the slump movements and the sedimentation are inextricably genetically related to each other. This paper relies extensively on Dr. Rider's analysis of the sedimentology of the Clare sediments, and his contribution to the documentation and understanding of many slump structures is gratefully acknowledged. The material derived from his thesis is clearly acknowledged as such in the text and in the illustrations.*

Previous knowledge of the sedimentary basin was built up through various stratigraphic studies (Neville, 1958; Shepard-Thorne, 1958; Kelk, 1960; Shelford, 1963; Hopkins, 1964; Brennand, 1965; Morton, 1965; Hudson and Philcox, 1965 and Shelford, 1967). The present state of interpretation and recognition of the status of the structures has resulted from many years of detailed documentation, thought, and discussion with numerous colleagues and friends. Their contribution is gratefully acknowledged. In the preparation of the paper I am grateful to C. V. MacDermot for comments on the stratigraphy and for several of the plates, and to D. Naylor for help with the editing.

This paper is published with a view to establishing a range of models of syndepositional slides and slumps which can be seen in perfect exposure along the West Clare coast. These models, it is hoped, will help in the understanding and documentation of similar phenomena in less perfectly exposed regions elsewhere. It is thought that the structures have a well-defined sedimentological association, and are therefore likely to be encountered in similar stratigraphic situations.

TERMINOLOGY

To the author's knowledge, the first person to document and recognise syndepositional deformation of sedimentary rocks was Sir William Logan in his drawings of the Devonian limestones on the cliffs of Cape Gaspé in Canada in 1863. Arnold Heim's account of recent subaqueous gliding in Lake Zurich and other Swiss lakes, published in Zurich 1876, was summarised, and fossil examples from Miocene and Eocene slides and breccias were fully discussed, in Heim's paper of 1908. Hahn (1913) further describes gliding deformations from the Trenton Falls in North America. All these early accounts are fully illustrated in Grabau's "Principles of Stratigraphy" published in 1913, which is probably the best account of early writings in this field in the English language. The examples quoted by Grabau were all described as subaqueous gliding, and the first person to use the word "slump" in this connection appears to have been T. C. Brown later in 1913.

Some of the earliest writings in Britain on this topic go back to Professor P. F. Kendall (1922) who ascribes deformations in rocks associated with coal measures to "lurching" of sediments under the influence of seismic activity. Archie Lamont (1938) described and discussed a wide range of pene-contemporaneous deformations from eastern Ireland. He uses the term "slip-bedding" and "slumping", without precisely indicating what differences there are between the two terms. Hudson and Dunnington (1940) also use the term "slip-bedding", but do not explain why they prefer this term to any other. Professor O. T. Jones (1937, 1940) uncompromisingly uses the terms "slump" and "slumping" for a wide range of syndepositional disturbances, although he discusses the gliding mechanism, in many cases almost indicating that the terms sliding and slumping are for him the same thing.

There is thus a variety of descriptive work prior to 1940, which refers variously to slip-bedding and slump-bedding as applied to small-scale folds and other deformations in soft rocks. Sliding on over-steepened slopes, and the effects of seismic activity are held to be the main mechanisms of these deformations. The main confusion in terminology arises from the fact that to American geologists the term "slump" refers clearly to a small surface land-slip. As will be

* Since this paper went to press Dr. Rider has published 'Growth Faults in Carboniferous of Western Ireland', *Bull. Am. Assoc. Petrol. Geol.* 62, 2191-2213, to which it has not been possible to insert appropriate reference in the text.

apparent in this paper, there are many analogies between subaqueous and surface land-slips, so that the only modification of the American usage of the word is to emphasis the syndepositional association.

On the question of terminology, the author considered using the terms pene-contemporaneous and syndepositional slump, but decided to leave the word slump in its well-defined and clearly understood English usage, particularly since the papers of Professor O. T. Jones. Much of the original debate about Jones's structures arose from the question of their syndepositional status, since in his ground good exposures of top contacts are rare. Most of the structures described in the present work have unequivocal top contacts, beautifully exposed in clean rock sections at the coast, so that their syndepositional status is not in question.

Descriptions are extended to quite a number of structures which do not have clear top contacts, but in these cases, the morphology and orientation of the structures is clearly different from the post-Westphalian tectonism, so that there is no hesitation in ascribing them to a syn-depositional movement.

A CLASSIFICATION OF SYNDEPOSITIONAL DEFORMATION OF SEDIMENTS

INTRODUCTION

Elliott (1965) clearly saw the desirability of clearing up the confusion which has arisen in nomenclature, by erecting a classification. However, his subdivision of sediments into rheological categories such as quasi-liquid and quasi-solid, and the terms he used for translation slumps associated with these conditions, have not proved to be capable of application to the rocks of the West Clare Basin. A new comprehensive classification of syn-sedimentary deformations is proposed to fit the range of the phenomena which have been studied in the field, and which, to an important extent, control the terminology which is used. Although in general terms genetic classifications have their pitfalls, no alternative scheme seems to fit the analysis of the phenomena in such a clear way.

Syndepositional deformations are considered to be due to four types of movement:–

MOVEMENT OF AIR AND GAS

Air-heave, gas pits, and mud volcanoes.

STRUCTURES DUE TO MOVEMENT OF WATER ABOVE THE SEDIMENT LINE

(a) "Convolute bedding" (current convolutions of Dzulynski and Slaczka, 1965).

(b) Overturned or over-steepened current bedding due to current-induced deformations on underlying hydroplastic sediments.

(c) Anti-dunes which move the opposite way from normal current ripples.

STRUCTURES DUE TO WATER MOVEMENT WITHIN THE SEDIMENT LAYER

The majority of these deformations are due primarily to permeability barriers; either thin layers of clay in sand, or fining-upwards beds as commonly developed in turbidites. The profusion of structures of this type is a result of the balance of overburden pressure with pore water pressure. When these two pressures are equal, the sand medium has zero strength (quicksand) and therefore responds to very small stress differences.

A great variety of structures of this type are described by Selley *et al.* (1963) from the Torridonian of Skye and Raasay as "streamers", "cusps", or "point-up" structures. Good examples of these occur in the Namurian of County Clare.

STRUCTURES REFLECTING MOVEMENT OF THE SEDIMENT

Dominantly Vertical Movement: Cast Structures

Load Casts (Keunen, 1953)

These involve the local sinking of the sediment layer at points of irregular loading on a hydro-plastic underlayer. They may be sand in clay, or sand in sand. The irregular loading may be due to erosional features at the base of the bed as in the typical cases of load-cast beds of turbidites, or positive depositional features such as ripples.

These latter constitute the majority of features in the West Clare Namurian. The downward movement of the sediment load is accompanied by diapirism of the lower mobile layer, giving structures variously known as "flame" structures, etc.

Ball and Pillow Structures (Smith, 1916) (Pseudonodules of Macar (1951) are synonymous)

These are cast structures on scales up to original bed thickness of about 1.5 m. The top of the structure can always be clearly identified by a planar termination of the end of the sack-like downward protrusion, which frequently has preserved current marks on its base. The horizontal section of the sack may be circular or to a varying degree ellipsoidal, sometimes having extremely elongated ellipses. These objects are not related to any detectable difference in loading.

The downward movement is clearly not associated with any rotation in the horizontal plane. The long axes of elongated forms are invariably parallel in one bed. It may be that the separation of the bed may have been achieved by a seismic wave-front which was followed by vertical movement of load-cast type.

Raft and Shale Diapir Structures

These structures do not appear to have been described prior to the observations in County Clare. Clearly, some lateral movement is indicated in the separation of a sheet of flaggy sand rock up to 15 m thick into rafts probably over half a kilometre wide in visible continuity, but dominantly vertical movement is evident in the intervening diapiric shale mounds, which in some places obviously broke surface in analagous fashion to the Mississippi mud lumps. Drag structure on the edge of the rafts is also frequently displayed.

Structures Reflecting Dominantly Horizontal Movement: Slides

These are subdivided primarily according to the rheological state of the material involved, and secondarily according to the shape of the sliding body.

Structures in unconsolidated or semi-consolidated sediment

1. Sheet Slides

This category comprises disturbed sediments the real extent of which is great compared with their thickness. The thicker the sheet is, the greater is usually its lateral extent. It may have abruptly sheared or progressively attenuated margins.

Syndepositional slides of this category usually have an erosional top surface followed by emissions of sand, often forming volcano-like structures. These were originally described by Gill and Kuenen (1958), but have also been observed in the "coal measures" of other parts of Ireland (Nevill, 1957), the Silurian rocks of Denbighshire, described by Professor O. T. Jones, (1937, 1940) and in the Baggy Point beds of the Devonian of Devonshire (Goldring, 1971, p. 44).

Two categories of Sheet Slides are recognised, according to the thickness of beds affected by sliding; namely Thin Sheet Slides and Extensive Thick Sheet Slides. Thin Sheet Slides are divided into five sub-categories (a to e below). Thick Sheet Slides, which in County Clare are associated with non-cyclothemic rocks, comprise two types (f and g below).

a) Thin Sheet Slides: Thin Folded Sheets

The simplest folds are cylindrical and their axes are perpendicular to the local or basin slope.

b) Thin Sheet Slides: Sandstone Ball Sheets

The development of these sheets arises from dominantly three mechanisms:–

i) The balls and pillows mentioned above create a planar sheet of inhomogeneous rock which is often a plane of subsequent further movement. This leads to a modification of the balls and pillows into more rounded, irregular structures, although frequently their origin can be plainly diagnosed.

ii) A second type of sandstone ball is derived from advanced deformation of the cylindrical folds of the thin sheets. The folds became broken into segments and rolled, much like making balls from flour dough. This process can be seen well at the locality of the cylindrical folds at Kilkee, and in end-form at Freagh Point, County Clare (see Gill and Kuenen, 1958). Relatively thin sheets of rubbly sandstone occur in association with certain types of sheet slumps, but since this rubble is more correctly described as breccia, the rock is deduced as having been fairly well-consolidated at the time of its deformation.

iii) Balled-up pillow-like structures developed from sheets distended by gravity faults (horst and graben structure).

c) Thin Sheet Slides: Sheared Slab Sheets

The best documented case is that of Goleen, County Clare, where the bed has been imbricated by sheets dipping at an angle of 25° to 30°. This sheet passes over a short distance into a zone of isoclinal folds the axial planes of which dip in the same direction at only a slightly steeper angle.

d) Thin Sheet Slides: Chaotic Slab Sheets

These are essentially silty flagstones broken into polygonal slabs by fractures with partial re-slurrying of the edges, and a variable degree of rotation, presumably arising from flowage. Type: Atlantic Lodge Slump, County Clare.

e) Thin Sheet Slides: Sheets Distended by Gravity Faulting

These small faults frequently antipose each other, giving graben and horst style of deformation. Sometimes these small distension faults tend to converge towards the lower surface of the bed, and are rarely seen to affect beds more than about one metre thick. In the Clare examples they are almost always associated with major sliding or rafting of sediments in the bedded sandstone parts of the cyclothem, the deposition of which accompanied the development of major slump structures.

f) Extensive Thick Sheet Slides: The Fisherstreet Type

At the type locality the slide is more than 20 m thick, and can be seen in cross-section for a distance of 3 miles (4 kilometres), although it must be much wider seawards as the margins of the plate are divergent to the north northwest at an angle of 55°. The visible cross-section has an extraordinary similarity to a Kober orogen. The tectonic styles, with thrusts and nappe-like forms, have an internal fabric which simulates the style of metamorphic folds. Higher confining stress in relation to the low degree of sediment consolidation is obviously responsible for this deformation style. The Fisherstreet structure is a single slide phenomenon within a stratigraphic interval which is slumped throughout its known regional extent.

g) Extensive Thick Sheet Slides: The Ross Type

This has a dominantly imbricated rather than a folded style. It has a remarkably consistent thickness of about 6 m for a distance of 2 miles (3 kilometres), before disappearing under cover, and it consists of three layers:–

A top layer of sandstone rubble up to about one metre thick. This is the rubble of slides from an overlying bedded sandstone sequence in which the scarps marking the back wall of the slips are clearly seen.

The middle division consists dominantly of slabs of black, banded siltstone, which are affected by folds of varied orientation, and a host of micro-tectonic features. Other than this, the slabs are still in horizontal attitude. Between the slabs are masses, lenses, and stringers of balled-up sandstone, which are clearly emplaced by imbricate thrusts. These imbrications give a consistent directional pattern, as do the smaller scale imbrications on the third layer of the slide:

The basal rubble or carpet, which is derived from break-up of the floor of the sliding sheet as a whole. In its simplest form, the carpet has been dragged into conspicuous small-amplitude folds. It is important to note that in this, as in the Fisherstreet Slide, many folds connected with sliding are not simply related to the direction of mass transport of the slide, and it would be misleading if the orientation of axial planes were interpreted as being perpendicular to the basin slope.

2. Shearing Slides

a) Simple Rotational Shears

Rotational shears are the commonest deformation of the cyclothemic rocks of the Namurian of West Clare, and frequently involve the entire thickness of rocks in the cyclothem. They are intimately connected with the sedimentation of these deltaic rocks, the down-dip direction at the top of the slip being coincident with the direction of sediment transport. The geometry of the structure is precisely that of subaerial examples, the main differences being that the sediment is much less de-watered, and, being subaqueous, the depression at the foot wall of the shear is filled usually with sand/silt sediment which is not present outside the area of influence of the structure. In almost all respects, they simulate in remarkable detail the structures associated with larger scale faults in the Niger Delta and Louisiana. The best model in County Clare is the Donegal Point locality.

b) Complex Sheets of Rotational Shears

In these cases, the slip planes of the rotational shears are very closely spaced, and of more than one generation, giving a highly complex pattern. The shears converge on a basal plane which can be followed for a considerable distance. The structures immediately above the basal shear are, not surprisingly, chaotic. The model for this structure is at Spanish Point, County Clare. The thickness is controlled by the thickness of the cyclothem, perhaps up to 45 m.

c) Chaotic Sheared Sheets (unconsolidated or semi-consolidated sediment)

These are several thick sheets of sheared-up, mixed-up sediment, usually of silt and sand in which we have not been able to detect any structures interpretable from simpler models. They are up to 60 m thick; for example Freagh Point, County Clare.

Chaotic Slides, dominantly of fragmented consolidated rock

Although this paper is not concerned with phenomena of this kind these rocks form such large volumes in sedimentary columns in certain areas, that they must be included in the classification for the sake of completeness.

The author wishes to differentiate three categories of contrasted genesis and character:

i) *Reef Talus Slides:* These products of the reef environment are almost ubiquitous and are self explanatory.

ii) *Chaotic Basin Margin Conglomerates and Breccias:* These are commonly developed from over-steepened or faulted margins of sedimentary basins, but a considerable variety of environments is possible. One example known to the author occurs in the Lower Carboniferous of north County Dublin, Ireland, where at frequent intervals, slides of breccia and micro-breccia containing mixed rocks of divergent ages, derived from collapsed fault scarps on the basin margin were emplaced deep in the interior of the basin (Smyth, 1950). The variable interval of time whilst the materials lay on a sedimentary terrace explains the variable degree of rounding of the boulders.

iii) *Chaotic mountain front slides:* These rocks are spectacularly developed in the Appenines, where they have been called Olistostromes (Rigo, 1956) but are also extensively developed along the front of the Taurus Mountain complex, the Himalayan mountain front and many other analogous situations. The slides are greatly variable in size, contain a variation of rock component fragments, and the matrix ranges from clays to microbreccias to volcanics. Essentially, however, all these rocks originate from tectonically created slopes and often from thrust fronts.

In classifying the above range of syndepositional sediment deformation, we have arrived at the situation where not a single structure bears the name "slump". It is to be noted, however, that the use of the classification is limited to those structures whose genesis can be diagnosed. There are very many instances where limitation of exposure, and/or lack of diagnostic criteria, prevent such a determination being made. It is therefore convenient to have a "portmanteau" term for all these indeterminate deformations, thought to be of pene-contemporaneous origin. This seems to be the most appropriate use of the word "slump".

STRATIGRAPHY OF THE WEST CLARE BASIN

INTRODUCTION

The stratigraphy of the Namurian deposits of west County Clare is established from two main sections, one in the northwest margin of the basin from the Cliffs of Moher southward to Liscannor Bay, Rineen and Mal Bay, and the other from south Clare to the estuary of the Shannon.

Figure 2 shows an outline of the stratigraphy, and the remarkable differences between these two sections. Two groups are recognised in the Namurian, in ascending sequence the Shannon Group and the Central Clare Group.

SHANNON GROUP

The Shannon Group (Rider 1974) is essentially non-cyclothemic and consists of three sub-divisions:

1. The Clare Shale Formation which outcrops from northwest Clare along the eastern margin of the basin to the Shannon and, from its presence at the base of the Namurian in north Kerry and in the Marathon Doonbeg borehole, is inferred to underlie the whole basin.

2. The Ross Sandstone Formation, equivalent to the middle and upper part of the Clare Shale Formation, forms the lowest exposed beds in southwest Clare, to which area its outcrops are confined.
3. The Gull Island Formation directly overlies the Clare Shale Formation from Fisherstreet along the eastern margin of the basin to the Shannon Estuary, and rests on the Ross Sandstone Formation in southwest Clare.

Clare Shale Formation

The Clare Shale Formation overlies the Carboniferous Limestones in an abrupt facies change from shelf facies limestones to black mudstones. A northward thinning from over 215 m at the Shannon estuary to 12 m in northwest Clare has been recorded by Hodson and Lewarne (1961).

In northwest Clare zones $H_1 - R_1$ are condensed in black shales about 12 m thick with distinct goniatite-rich fossil bands (Hodson 1954a). Work in progress by the Geological Survey indicates that shales of H_1 age with a basal phosphorite rest with minor angular unconformity on limestone of middle Brigantian age, cherty mudstones of E age found at the base of the Clare Shales further to the southeast being absent through overlap and overstep. (MacDermot *pers. comm.*)

Ross Sandstone Formation

The general lithostratigraphic succession in the Ross Sandstone Formation is illustrated in Figure 3. The sandstone beds are characteristically parallel-bedded (Plate 1), the thicker units being commonly composite. Flutes and grooves are sometimes displayed, but more characteristically the bottom structures are "rill casts". Ripple marks are common at the top of each sedimentological unit, although convolutions are rare. Grading is seldom seen, but this is perhaps not unexpected in view of the unimodal nature of the sediment. Individual beds may be from 1 cm to 2 m in thickness, and commonly continue at constant thickness for large distances. The main exception to this rule is that of the cross-cutting surfaces, which will be explained later.

Shallow scour gouges 1.5 m to 9 m wide are indicative of erosion, and the gouges are mainly filled with shale. Sand-filled channels up to 10 m deep, and more than 100 m wide are indicative of occasional strong currents. Bands of black shale, 0.5 m to 1.5 m thick, with fossil goniatites and lamellibranchs (the former with pyritic nuclei) occur at three separate horizons in the section. Dark grey silts up to 6 m thick also occur between the sandstones. The sedimentological environment is diagnosed as mud-trough or pro-delta mud flat with invading turbidite sandstone. Interesting in this regard is the description of the Cloone Flagstones of County Kerry by Brennand (1965) which occupies a similar stratigraphic interval. Small sand volcanoes up to 5 cm in diameter occur on the upper surfaces of several turbidite beds.

Goniatite faunas from three black shale bands date the Ross Sandstone Formation as $H_2 - R_1$ (Rider 1974). The 12 m of Clare Shale Formation in the northwest have expanded over a distance of only 55 km to over 300 m of turbidites and siltstones at Loop Head.

Gull Island Formation

Overlying the Ross Sandstone Formation is the Gull Island Formation. This is defined as the interval above the *Reticuloceras paucicrenulatum* band at the top of the Ross Sandstone Formation, to the base of the cyclothem containing the Tullig Sandstone Member. The nature of the upper contact in different parts of the West Clare Basin is illustrated by the graphic logs of Figure 4 which clearly reveal the point in the section at which the cyclothemic deposits commence. The dominant characteristics of the Gull Island Formation are that the formation is much thinner bedded, and has more silt than the Ross Formation below, the silt intervals being commonly associated with slumping.

The percentage of sandstone decreases upwards through the formation, the sandstones maintaining their turbidite characteristics, with sharp bases, frequent bottom structures, poor grading, and rippled top surfaces. In the lower third of the 400 m (1,300 ft) section, at the type locality, sandstones occur in 6 m to 10 m groups, some of which seem to have been deposited in channels 6 m to 10 m deep, and more than 1,000 m wide, the best exposed channel being visible in the Cliffs of Moher at the northern end of the section. In the upper two-thirds of the formation, groups of sandstones are rare, most beds being less than 1.5 m thick, with fluted bases.

The siltstones comprising most of the section are extensively slumped. Current directions, from sole marks and ripples, indicate a westerly provenance for the turbidites (Figure 2).

CENTRAL CLARE GROUP

This cyclothemic group was defined by Rider (1974). A typical cyclothem consists of a basal goniatite band in black shale, sometimes with a thin band of limestone, overlain by a thin shale which is followed by a thick, black, silty sequence, and topped by a sandy succession of variable thickness. The vertical order of these facies is considered to record a prograding lateral sequence, the facies juxtaposition being similar to that of the modern Mississippi Delta (Figure 5). Not only is the order of the coarsening-upwards sequence the same, but also the grain-size and sediment composition. The fossil-rich goniatite shales are regarded as being deposited in a marine shelf environment; the overlying black silts are regarded as pro-delta muds; the laminated sandstone beds are probably distal bar and distributary mouth deposits; the capping, thick sandstone development of the cycle is regarded as fluvial, and deposited in channels flowing over the delta surface.

The dominant lithostratigraphic marker horizons of the West Clare Namurian succession are associated with two ubiquitous cyclothems, firstly the Tullig Cyclothem, topped by the Tullig Sandstone Member, and secondly the Kilkee Cyclothem, topped by the Kilkee Sandstone Member (Figure 6). The Tullig Cyclothem does not have a marine horizon at its base but the change from turbidite conditions to the basal argillites of this cyclothem is well-defined.

The base of the Kilkee Cyclothem is the most conspicuous palaeontological marker in County Clare, i.e. the marine horizon with *Reticuloceras aff. stubblefieldi.* The sandstone members of these two cyclothems have almost everywhere a topographic expression, and form the basis of mapping the structure of County Clare (Figure 1). Overlying these two cyclothems is the Doonlicky Cyclothem which is similar to the previous two as a sedimentary model. The goniatite shale band at the base of this cyclothem contains *Reticuloceras cf. reticulatum ss.* (R_1). The rock successions of these cyclothems are clearly identified in Figure 7, and the location of the stratigraphic columns is provided in the map in Figure 8.

Associated with the major cyclothems, especially where the massive sandstone member is thin or not represented, are smaller cycles, frequently less than 30 m thick, which are diagnosed as being of two types:–

Sub-deltas, called Normal Small Cycles by Rider (1969)

Pro-grading Beaches, named Burrowed Small Cycles by Rider (1969)

Rocks stratigraphically higher than the Doonlicky Cyclothem have a very limited distribution in Central Clare. They occur on the coast, principally in the Mal Bay Syncline, between Spanish Point and the Cassino fault (Figure 1). Their cyclothemic character is still recognisable, but they do not form an important part of the studies here presented. They contain four bands of R_2 goniaties *(Reticuloceras bilingue* at the base, two close together about one third of the way up the sequence with *Reticuloceras wrighti,* and one with *Reticuloceras superblingue* near the top of the preserved succession (Rider 1974, Figure 2).

The detailed sedimentology of all the Clare rock successions will be described by Dr. Rider in separate publications, but a clear statement of the essential characters of the sedimentary model in each cyclothem is important to explain the slump phenomena to be described. The essential palaeocurrent data and facies distribution in the three main cyclothems in the succession are given in Figures 9, 10, and 11. It is important to note the contrasted directions of sediment transport at the top of these three sedimentary cycles which often allow each cyclothem to be identified without reference to the fossils at their bases. The maximum thickness observed for a cyclothem is about 120 m, with a maximum of say 30 m to 45 m for the delta sandstone finger at the top, the basal marine shale being rarely more than 15 m thick, leaving a dominant middle interval of dark, banded siltstone and laminated sandstone. It will be shown later that there is a close relationship between the scale of the cyclothem, and that of the slump structures associated with them.

TECTONICS

The Namurian succession outlined above was folded by the post-Westphalian Variscan orogenic movements. The intensity and morphology of the foldings are illustrated by the cross sections in Figure 1. These sections show:—

1) A southerly increase of fold intensity from the south-dipping monocline of north Clare to the Shannon Estuary.
2) The major antiforms and synforms have superimposed upon them a second order frill of folds, probably reflecting a high strain rate. The general axial trend of the folds is 10° to 15° north of east, which is not far from the trend of the axis of the Shannon trough of deposition.
3) Faulting is common in two directions:—
   a) *Strike faulting.* The Cassino Fault at the southern margin of Mal Bay is a northwards thrust structure with a displacement of over 150 m.
   b) *Dip faults.* These are associated with sub-parallel northerly-trending gash-vein sets, and are very common, particularly near the Shannon Estuary.

The folding was followed by strong thermal effects, which see the thin coals devolatolised to high-grade anthracites. The sand and siltstone rocks are orthoquartzites, which ring under the hammer, and as a result of long marine erosion, many of the structures here described have been beautifully etched but are not easily eroded. Quartz segregation is common, much of it in joints and gash-veins, but also in cavities created by slump-breccias and slump faults. The latest tectonic movements have created slickensides along these quartz-filled planar elements, and these greatly complicate the process of differentiating between slump movements and later tectonic movements. The essential understanding of this difference comes from the recognition not only of the top contacts, following erosion of slumped masses, but also of the essential rheological state of the materials at the time of their deformation. The low level of consolidation of sediments down to a depth of 60 m, almost invariably gave the range of slump structures a totally different deformation style and morphology from those due to a later tectonic influence, which involved well-consolidated rocks. These criteria alone should serve to identify syndepositional slumps in isolated exposures, or in cores, where top contracts are not seen.

SLUMP STRUCTURES IN THE WEST CLARE BASIN

To document all the slump structures of the West Clare Basin would be to produce a long repetitious document, since possibly between 30 and 40 per cent of the entire succession is affected in this way. Neither is it feasible to publish all the field records which have been made during the past 20 years of intermittent study. A careful selection of structures has been made in order best to illustrate the genetic classification outlined in the earlier chapter. The first two categories of the classification (Structures produced by: Movement of air and gas; Movement of water above the sediment layer) will not be further discussed as they are not represented in the sedimentary environment of West Clare.

STRUCTURES DUE TO WATER MOVEMENT WITHIN THE SEDIMENT LAYER

Thick sandstones of the on-delta parts of the cyclothems in County Clare sometimes show obvious modifications of the original sedimentary structure, the bedding being wavy and sometimes more severely contorted. This is due to a "quick" condition of the sand, brought about by loading of water-logged sand, above a low permeability film by which the overburden and pore-water pressures become equal. What causes the precise shape of the contortions is not quite clear. The best examples of these are to be seen in the upper layers of the Tullig Sandstone at Killard, on the southeast limb of the syncline, slump locality G, Figure 1. As can be seen from the section in Figure 13, the base of the disturbed rock is cross-cutting the major bedding planes. Why this structure should be confined to the top of the 60 m thick sandstone member is not clear. The author originally thought that release of pore-water following closer packing of the sand grains after seismic shock might explain this "sloppy" top, but the similarity of the structures to those described by Selley *et al.* (1963) is obvious (see Plates 2 and 3).

STRUCTURES REFLECTING MOVEMENT OF THE SEDIMENT

Structures due dominantly to vertical movement

Load Casts (Keunen, 1953)

These structures are common in the West Clare Namurian succession, with two sedimentary associations:–

(a) Load cast bottom structures of turbidite sandstone in the Ross Sandstone and the Gull Island Formations.

(b) Load casting of ripples on many scales is a common feature of the cyclothemic rocks. An interesting study of progressive development of these structures can be seen in the face of the Black Rock on the foreshore of Mal Bay in measures of R_2 age. On the seaward side the entire interval is unaffected, and consists of rippled sands in silt and clay. Over a distance of 40 m, load casting of ripples progressively develops through the stage seen in Plate 4 to the conditions seen in Plates 5 and 6, where much-deformed small ripples are seen to float in a matrix of slurried silt.

Ball and Pillow Structures (Plates 7 and 8)

Structures of this type are conspicuously developed in the cyclothemic formations, particularly in the banded sandstone facies in the upper parts of the cyclothems, but they are rarely to be seen in thick beds of sandstone. They function excellently as local marker horizons, and often permit precise measurement of the extent of vertical slump movement. Beds so affected are commonly up to 1.5 m thick, often of rippled flagstone. Sometimes the bed is affected over the entire visible lateral range of exposure up to 300 m but frequently it passes from a balled-up condition into normal bedding within the exposures. The mode of formation can be clearly seen in Plate 7 as a typical sack-like downwards protrusion of the sand bed, accompanied by upward diapirism of the underlying clay or silt. The best examples, which are easy of access for study, may be seen in the foreshore at Quilty (Locality I, opposite the church, see Figure 14). At Quilty, a bed of rippled flagstones 1 m thick passes from normal to balled-up condition. Adjacent pillows frequently preserve the ripples at the base of the bed (Plate 8), with a common current direction still preserved. Other excellent examples on a scale which is similar can be seen at Blind Horse's Cave, Killard (see Figure 15 and Plate 15). Beds 7 and 8 on this figure show many variations in shape and degree of development of the pillow. The rounded ball shapes to the right, in Bed 8, have lost their sack-like shape, but the sack structure can still be clearly diagnosed. Many variations of the same structure can be seen in the eastern parts of the Atlantic Slump, Spanish Point (Figure 16).

Most of the examples quoted above present sack-like forms. In some localities, the horizontal

Table 1

CLASSIFICATION OF SYNDEPOSITIONAL DEFORMATION OF SEDIMENTS

TYPE OF MOVEMENT	STRUCTURES
MOVEMENT OF AIR OR GAS	**Air heave, gas pits, mud volcanoes**
MOVEMENT OF WATER ABOVE SEDIMENT LINE	**Convolute bedding, oversteepened current bedding etc.**
WATER MOVEMENT WITHIN SEDIMENT LAYER	**'Point up' structures etc.**
MOVEMENT OF THE SEDIMENT **Dominantly Vertical Movement**	**Load casts** **Ball and pillow structures** **Raft and shale diapir structures**
MOVEMENT OF THE SEDIMENT **Dominantly Horizontal Movement**	**Unconsolidated – semi consolidated sediments** *Sheet Slides:* *(a) Thin sheet slides : thin folded sheets* *(b) Thin sheet slides : sandstone ball sheets* *(c) Thin sheet slides : sheared slab sheets* *(d) Thin sheet slides : chaotic slab sheets* *(e) Thin sheet slides : distended by gravity* *(f) Extensive thick slides : Fisherstreet type* *(g) Extensive thick sheet slides : Ross type* *Shearing Slides* *(a) Simple rotational Shears* *(b) Complex sheets of rotational shears* *(c) Chaotic sheared sheets* **Chaotic slides of fragmented consolidated rock** *Reef talus slides* *Chaotic Basin margin conglomerates and breccias* *Chaotic mountain front slides*

section becomes elliptical, and the structure may be loosely described as pillow-shaped. A bed of these structures can be seen in the amphitheatre near the Diamond Rocks, Kilkee, and on the foreshore at Quilty (Locality I) a few hundred metres to the south of the section represented on Figure 14. In both these localities, the pillows are isolated by a metre or two of silt or shale, and the axes of the pillows are sub-parallel. With regard to the origin of ball and pillow structures Smith (1916) suggested the influence of gravity. Whilst this is obviously true of the later parts of the development, the triggering mechanism of the original movement is not established. Selley *(pers. comm.)* has produced similar structures in a beaker by tapping the side of the beaker with a rubber hammer. The author is therefore inclined to think that the separation of the bed comes from an earthquake shock. This might explain the pillow structures as being parallel to the wave-front. These structures are quite plainly not produced by any difference in sediment load due to original sedimentation processes. Once the bed has been separated, however, the deformation proceeds much as a load cast.

Raft and Shale Diapir Structures

These remarkable structures are to be seen at several points around Moore Bay, the bay on which the town of Kilkee stands. A map of the area is given in Figure 18, and the strata affected constitute a coarsening-upwards cycle some 25 m to 30 m thick, which is only one third of the thickness of the thick progradational sequences in the basin (Plates 9 and 10). Plate 9 shows the typical appearance of the rafts in the coarser, flaggy sandstones (about 8 m thick locally) at the top of the cycle which have separated and become intruded by a brecciated shale diapir which cuts right through to the top of the cyclothem. At these localities, the intruding silty shale is a breccia with fragments up to 8 cm in the longer dimension, with a thin film of re-slurried matrix. Only 5 m of the silty shale is involved; the lower part is undeformed. This characteristic form of flow breccia is illustrated in Plate 11 from Killard. In other instances the shale is in much larger blocks, but still seems to have achieved its present position by flow.

A detailed study of the area has shown that the movements are not restricted to one episode at the end of the cyclothem, but that they affect bed thickness, and create the features of minor slump structures, seemingly throughout the deposition of the sandstone part of the cyclothem. This part of the sequence is affected by sheet slump features, sandstone ball beds, and conspicuous discontinuities, which are regarded as accompanying the diapirism.

At the Pollack Holes, Locality C3, Figure 18, Rider (1969) mapped four discontinuities with three distinct phases of thin sheet slumping, see Figure 19. The beds below Discontinuity 3 are dominantly affected by ball and pillow structures, which have a preferred orientation perpendicular to the margin of the diapir immediately to the west. The beds immediately above Discontinuity 2, and above Discontinuity 1, show remarkable sheet extension by steep minor faults of rotational shear aspect. The intensity of these small sheet unconformities and slumping increases towards the points of diapirism. There is a fine exposure of thin sheet movement by cylindrical folding in the cliff face at Locality D2, Figure 18 on the north side of Moore Bay, Kilkee (Figure 20). These sheets of "tablecloth" folds will be described in a later section, but it is interesting to note that the direction of the movement – towards the southeast (150°)—could well be reflecting the slope from the adjacent diapir to the north.

The cause of the separation of the sandstones into rafts is not known. An obvious recent analogy to these ancient diapirs can be found in the shale diapirs or mud lumps of the modern Mississippi Delta (Morgan *et al.*, 1968). There, the diapirism can be shown to result from rapidly deposited sand at the distributary mouth, weighing heavily on the lighter pro-delta silts and clays beneath. It would seem therefore that this is an effect of progradation. The author is of the opinion that progradation and unequal loading are probably the causal factor of the Namurian diapirs of Kilkee, but exposure limitations, and the advanced stage of many of the deformations, make this difficult to prove.

Structures due dominantly to horizontal movement (slides) in unconsolidated or semi-consolidated sediments

1. Sheet Slides

(a) Thin Sheet Slides: Thin Folded Sheets

The best exposures of these are to be seen in the cliff faces on the north side of Moore Bay (Locality D2, Figure 18). The structures are shown in block diagram form in Figure 20. The lowest sheet, with a maximum thickness of about 1.5 m, is the clearest demonstration of "tablecloth" sliding in the area. Three parts of the sheet can be clearly diagnosed.

At the southern end, near the sea, the cylindrical folding passes without discontinuity into normal bedded flagstones, underlain by somewhat sheared silt. The belt of cylindrical folds extends for about 15 m northwards, the underlying silt being squeezed up into the anticlinal cores. The eroded top of this sheet gives the synclinal element a cross sectional shape rather like the ball and pillow structures described above. This emphasises the need when classifying slumped rocks not to base too many conclusions on poor exposures.

To the north of the cylindrical folds are unfolded flagstones, somewhat tilted northwards, the bedding being cut off at a basal plane. The interval of flagstones progressively diminishes until it is represented by about 30 cm of sheared sand and silt.

This sheet illustrates the fundamental principles which the author has followed to resolve movement directions in slides of all kinds. Firstly, the passage of duplicated rock into unduplicated rock without structural discontinuity indicates that this is the distal end of the slide. More important, perhaps, is the diagnosis of the diastem, which indicates the proximal end, or the hole from which the material of the slide came.

The two overlying slumped beds at Locality D2 are also of interest. The lower of these beds shows sections of cylindrical folds (extreme northern end of the diagram, Figure 20) but these pass quickly into a thicker bed of completely balled-up flagstones. The top bed depicted at the northern end of the sheet shows a layer less than 30 cm thick, increasing in thickness southwards, of beautiful cylindrical folds, with an axial trend similar to that of the lowest layer. Further south, these cylinders show numerous deformations, and their axes become distinctly bent and broken. Finally, the sheet passes into a zone of chaotic sandstone balls, which are clearly the result of the break-up and rolling of the original cylinders. In this zone, the two top sheets merge, and cannot be differentiated. These sandstone balls are illustrated in Plate 12. The cylindrical folds in all three sheets at this locality have a common orientation, and the direction of sliding is to the southeast (150°).

Thin folded sheets in more muddy sediments show a wide variety of more complex deformations representing much less viscous materials. The core in Plate 13 shows miniature nappe-like forms. The external relationships of this specimen are not established, and although the local sense of movement is quite clear, it is not known whether this is of any use in indicating basin slope. A discussion of this point will be deferred to a later section.

b) Thin Sheet Slides: Sandstone Ball Sheets

Sheets of sandstone balls on all scales of bed thickness, up to a maximum of about 3 m are so common that they appear on almost every figure in this paper, and may be said to be the commonest obvious "slump" deformation of the entire succession. They are more common in the cyclothemic rocks than in the non-cyclothemic ones. In the earlier drawings of many of the figures, the desirability of diagnosing the mode of origin of the sandstone balls was not appreciated, but in later work, clear examples were recorded, showing four quite different types of development.

The first and commonest form concerns the further deformation by sliding of ball and pillow structures, and, perhaps, on a small scale, of load casts. Clear gradations from one structure to the other can be seen at Blind Horse's Cave (Figure 15) and at Quilty (Figure 14). Even when this transition cannot be seen, a cross section of the ball frequently permits remnants of the sack-like structure to be diagnosed. In other examples, the clear parallelism of a large part of the ball surface with original bedding, as revealed by the bent ripple-marked surfaces (e.g. Quilty, Plate 8), leads to the consideration that these ball sheets arise from break-down of lateral continuity in the strength of the bed, which renders it a preferred plane for further sliding, being as it were, a layer of ball bearings. In the Quilty case (Figure 14) the mass of balls can hardly be regarded as a sheet by the original definition, since it is to be seen in a two-dimensional section over a distance of 35 m to be a shallow, channel-like form. The downcutting motion of the masses of pillows is best interpreted as a series of rotational shears, creating hollows which were current-infilled from the right by a mixture of sand and rubble. It is therefore considered that this slump should come under the category of a rotational shear.

The development of the second type of sandstone ball bed has been discussed from Locality D2 in Moore Bay, Kilkee (Figure 20). Here it was deduced that the sandstone balls arose from the break-up and rolling of cylindrical folds. This is probably a unique exposure, but the product of the mechanism is a sandstone ball, rolled up like a piece of dough, in which the structure is quite chaotic, and no antecedent structure can be found. A bed of such balls has been figured underlying a sheet of sand volcanoes (Figure 21) at Freagh Point (Gill and Kuenen 1958, Figure 2).

Balled-up structures derived from sheet distension by gravity form the third category. In many localities in the cyclothemic succession, sandstone beds infilling hollows created by slumping are distended by systems of small faults resembling rotational shears. When seen in profile, the bottom part of each small fault is seen to curve towards the basal plane. At the top of the bed, the faults are typically steep, as seen at Spanish Point (Figure 55) where they form a characteristic graben and horst relationship. Systems of this kind have been referred to in the map of the Pollack Holes, Moore Bay (Figure 19) above. This style of deformation is seen very clearly in the Donegal Point Slump, Figure 22, where the small faults are sub-parallel to the major rotational shear faults. A large variety of structures of this type can be seen in the somewhat flat-lying top of the Kilkee Cyclothem, between Freagh Point and Green Island (Figure 24). For a long time, the author was doubtful about the status of these structures *vis-à-vis* the tectonic situation, but is now firmly of the opinion that they are syndepositional structures.

At two key localities, the separation into graben and horst can be seen to be followed by a further development. At the Atlantic Slump, Mal Bay (Figures 16, 17), the graben and horst blocks are seen to be progressively deformed until they assume large ball and pillow-like masses in a siltstone matrix. At Spanish Point (Locality K north, Figure 55), a similar development of sandstone balls from graben and horst structure can also be demonstrated.

The fourth category of sandstone balls comprises those which are derived from the collapse of sand volcanoes. These structures were first observed by Gill and Kuenen (1958). Figure 21 shows that the sheet of sand volcanoes loses diagnostic structure to the left of the diagram, and immediately below it are curiously contorted sandstone balls. In one of the balls remnants of the sand volcano structure can be diagnosed. Also, in the upper layers of sandstone slumps of many kinds, particularly in those associated with the Ross Slide, it has been recognised that the partial re-slurrying below the sand volcano layer results in sediment incapable of supporting the structures. In consequence the layer then collapses into the substructure with various degrees of deformation, very frequently breaking up into separate fragments, and ultimately forms sandstone balls whose origin is then totally obscure. This type of collapse would appear to be the most common origin for balled structures having a diameter of more than 1 m. Deformation of this sort can also be seen at the Baggy Point Slump in the Devonian of Devonshire, England.

c) Thin Sheet Slides: Sheared Slab Sheets

At Goleen (Locality B, Figure 1), a 30 m thick section illustrating a wide variety of slumped rocks is well exposed on a broad, bare-rock platform. This is the locality of the largest sand volcanoes in County Clare, figured by Gill and Kuenen (1956; Figures 3, 4 and 5). The beds have been mapped on a scale of 1:50, and a vertical section projected approximately on the line XY (Figures 25 and 26). In the upper part of this section are beds of fine sandstone, 30 cm to 1 m thick, which have a very distinctive structure typified by Plate 16, which is of the thicker parts of Bed No. 2 (Figure 26). This bed consists of sheared slabs with silty partings. Two sets of shears are clearly visible in Plate 16, those on the right dipping at 30° and those on the left at about 40°. As the bed thins out towards the west, the slab structure disappears, and finally only a thin layer about 15 cm thick of structureless sand is to be seen. The structure of this bed is regarded as the product of imbrication of a sheet of sand by shear sliding from west to east, with consequent thickening of the bed.

Bed No. 3 (Figure 26) shows a similar structure in the eastern half of the diagram. It then becomes a thin, structureless bed for about 30 m. Over the next 30 m the bed thickens and has a somewhat chaotic structure with large sandstone balls. There then follows a zone of isoclinal folds with the axial planes dipping at similar angles to the sheared slabs, and with an eroded upper surface. This passes without discontinuity into undeformed sandstone. From basic reasoning, this passage from folded to unfolded rock ought to indicate the distal end of a slide, but in this case, the geometry is clearly indicative of a movement from left to right, and it would suggest that the apparently undeformed sand flowed in "quick" condition. Bed No. 6 (Figure 26) is a mixture of deformation styles, each indicating a movement from west to east.

d) Thin Sheet Slides: Chaotic Slab Sheets

The best examples of this type of sheet are to be seen at the Atlantic Slump (Figures 16 and 17). The basal member of this is a flaggy siltstone, probably about 3 m thick, although the base is not seen in the central part of the section. This distinctive style of deformation begins where the bed appears to have been broken up into polygonal or oval slabs, separated by vertical zones which superficially appear to have been re-slurried (Plate 18). Many of these zones cross each other, showing that they belong to several generations. The impression of a broken crust with fluid under-layer oozing from the cracks proves to be erroneous on closer inspection, in that the cracks consist of multiple shears separated by thin walls of silt, with original bedding intact. The only re-alignment of clay minerals is along the shears.

From east to west (A-B, Figure 17), the slabs become progressively tilted until at the western limit they lie at chaotic angles. Here there is more re-slurrying of the matrix, but essentially it consists of a system of shears. No bending of the slabs has been observed at this locality, indicating a considerable degree of consolidation at a shallow depth of burial.

Another common style of deformation in silts is that illustrated at the western end of the map in Figure 17 (Bed 6). It consists of a system of flat-lying, gently curved shear planes, which break up the mass into flat, lentical forms, perhaps up to 6 m in maximum dimension.

It is interesting that these deformation styles are confined to dominantly silty rocks, and thus to the lower third of each cyclothem. Rider (1969) refers to them as dish slumps. The absence of deformation in the siltstone slabs, in spite of the extensive geometric re-arrangement, suggests that these rocks have broken up by tensional shears. Where broken block fabrics have been seen with the ends of the blocks deformed, they are invariably in more sandy facies than at the Atlantic Slump. Truly balled-up structures appear to be confined to sandstones.

e) Thin Sheets: Distended by Gravity Faulting

In the cyclothemic rocks of County Clare, minor faulting with a graben and horst arrangement is a widespread small-scale, but nevertheless conspicuous, feature. The faults are normal, with 30-60 cm or so of throw, display straight faces for 10 m to 30 m and are spaced from 1 m to 3 m apart. Since the bedded sandstones occur in the upper parts of the cyclothem, the minor fault structures are synchronous with larger scale slumping, and are thought to indicate mild distension of the beds as a secondary feature accompanying the major disturbances.

The most frequent association is with the large rotational shears, whose trend they almost perfectly parallel. They presumably represent extension of the shear as it slips away from the fault face. Another association is with shale diapirism, as shown at the Pollack Holes (Figure 19). They are spectacularly seen in front of the major rotational shear face at the Donegal Point Slump (Figure 22). A reference has already been made to the very extensive development of this structural style between Freagh Point and Green Island (Figure 24). Here, Rider (1969) mapped and studied in detail the 50 m of the exposed part of the Kilkee cyclothem, which shows a downward increase in intensity of slumping with a dominant component of closely-spaced, sub-parallel, major slumped faults. The sub-parallel aspect of the minor horst and graven fault features, and the complete array of other minor slump features clearly confirm the contemporaneous status of these structures.

f) Extensive Thick Sheet Slides: The Fisherstreet Type

The Fisherstreet Slide is the largest and most instructive slide of its type that the author has seen. It occurs in the lower part of the Gull Island Formation, above the basal member termed the Ribbed Beds by Hodson and Lewarne (1961). From the author's measurements at Fisherstreet the slumped beds are at least 20 m thick. The base is well seen inland at the top of a waterfall in the Cronagort stream (grid ref. R 077966) where sandy silts with a well-developed flow cleavage rest on 9.5 m undeformed siltstones and distal turbidite sandstones of the Ribbed Beds (MacDermot *pers. comm.*).

Essentially, the lithology at Fisherstreet is one of dark grey micaceous siltstone, with frequent paler, finer, sand laminae, and some fine sandstone beds up to 30 cm thick. There is a strong suggestion that this sandy silt section has been sliding on the underlying Ribbed Beds. The Fisherstreet Slide can be studied in a combination of cliff and rock platform exposures over a distance of 4 kilometres at Locality M (Figure 1), easily accessible from the track going southwards from the village of Fisherstreet towards the Cliffs of Moher. Further south, the southerly dip results in the disappearance of the horizon below the higher succession of the Cliffs of Moher. The description of the Fisherstreet Slide will be dealt with as follows:

Overall tectonic form and tectonic style.
Features of internal deformation and fabrics.
A discussion of overall kinematics and dynamics.

Overall Tectonic Form and Tectonic Style

The generalised map and section of the Fisherstreet Slide is shown in Figure 27. In the section, the structures are shown as if viewed from the southeast, which better enables the integration of plan and cross-section. The detailed cross-sections in Figures 28 and 29, along lines CDE and AB, are shown as they are seen looking at the cliff. Essentially, the sheet consists of a northern two-thirds, characterised by thrusts dipping southwards at 20° to 40°. Based on these thrusts is a varied system of recumbent, nappe-like folds. The southern third of the section has a dominant component of northerly-dipping thrusts at much the same angle. There is an angular difference of about 55°, however, between the axial trend of the southern structures which are approximately west northwest trending, and those at the northern end which trend almost due north. The resemblance to a Kober orogen in this sheet about 3.5 kilometres wide, and probably about 20 m thick, is truly remarkable. Indeed, the eroded top, followed by flat-lying Cronagort Sandstone would truly impress any observer as at least a significant stage of orogeny (Plate 19). It is obvious from the appearance of the sections that the fold style simulates that seen in metamorphic rocks, with a high degree of flank-thinning and hinge-thickening. This is seen at its best in Plate 20. In section CED, Figure 28, it is significant to note that axial plane cleavage develops in some of the recumbent folds. The straightforward nappe-like structures have obviously become modified in the upper layers of the slide by two features. Re-folded folds on the right of section CDE have a geometry which strongly suggests that they have been refolded by sliding in the opposite direction to the primary thrusts. This is further corroborated by the folds on the left of section CDE, which have northerly dipping axial planes. It would appear that the outward thrusts formed unstable topographic highs which subsequently slid backwards under the influence of gravity. From section CDE southwards, there are narrow zones of high intensity thrusting and folding, separated by more flat-lying strata. These flat-lying beds, however, have strong internal deformations, which will be discussed later. The thrusts are clean-cut structures (Plate 21), and divide the section into separate structural compartments which have much contrasted geometry. Figure 30 is a detailed map of the southern margin of the slide. The section on Figure 29 is on a somewhat exaggerated vertical scale, which necessitates drawing the shear planes at a steeper angle than that at which they occur. Essentially, the planes are at an angle of 35° or less. At the left of Figure 29, there is a system of south-dipping thrusts with accompanying recumbent folds. At the southern end of the section, in the lower layers, there is a complete change of directions, and, for a short distance, some thrust faults dip northwards, their recumbent axial planes dipping in the same direction.

The fold closure (Plate 22) against the largest of these northerly-dipping thrusts is depicted about half way across the section in Figure 30. The axis of this structure is at an angle of 30° to the surface trace of the thrust plane, and indicates a movement southeastwards. This evidence of an easterly component in the movement picture is important in any consideration of the kinematics of the slide. At the southern end of the section, the slide has a distinctive upper tectonic unit in which the folds, as discernable from the thicker sandstone beds, have been much broken up, and lie everywhere in recumbent attitude, as seen in Plate 23. The cross-section in Figure 29 strongly suggests that this upper layer has been formed by southwards gliding from a topographic high, now seen in the eroded, truncated primary structures further north. It is interesting to note that wherever the axial planes in this slide are truly horizontal, either in Figures 28 or 29, there is reason to infer gliding tectonics, thus supporting De Sitter's (1956) contention about such structures in mountain systems.

Features of Internal Deformation and Fabrics

One of the conspicuous features of the Fisherstreet Slide is the host of smaller scale structures, all of which give the rocks a strong lineation sub-parallel to the axial trend of the folds, but they are, as will be seen, very complex, and of varied origin. One unusual feature is depicted in Plate 24, in what superficially looks like a folded bed. In fact, as can be seen, the base of the bed is not affected, but the upper surface has been deeply grooved, presumably by differential movement between the bed and the bed overlying it. There is clearly no fold deformation of this bed, and the strong lineation is due to the intersection of the grooved surface and the fine laminations of the sandy silts. This grooving can be readily diagnosed, particularly on flat-lying surfaces, but can also be discerned even when it occurs with a variety of fabrics of slump origin. Phenomena which will be familiar to metamorphic geologists are responsible for the strongly lineated surfaces seen in Plate 25. These are figured in block diagram form in Figures 31 and 32, which show the distinctive separation of the sandy beds into mullion-like forms. There is a suspicion, however, that the bed separation is not simply pull-apart in the AC plane. Figure 32 in particular, suggests that the separation was accompanied by some form of grooving in the B direction, and the author suspects that there has been considerable movement in the apparent B direction between elements of fold cores. Further evidence of differential movement in this B direction is shown by the twisted fold core figured in Figure 33, the apparent drag effect of which would indicate movement in an easterly direction. Most of the sections of rounded sand bodies show in Figure 29 have this mullion-like aspect.

On a smaller scale, to be seen on hand specimens, are conspicuous B lineations due to the intersection of micro-scale folds, thrusts, and fold frills, which are simply a scaled-down model of the major structures. The specimen shown in Figure 34 was collected at about the middle of section CDE (Figure 29). Small scale fold frills can be seen even where larger structures are not visible, as in Figure 35. In the examples mentioned above, where grooving can be diagnosed, it is parallel to the axial trend of folds on all scales. There are many localities, however, in which sets of fold frills intersect the grooving direction (Figure 36), the angular difference being commonly about 45°. Two sets of frills can be seen in the same specimen (Figure 36), but in many cases in a 1.5 m section of flat-lying sediments, the trace of fold frills is never consistent from one surface to the next, and their trend can often been seen to change within the area of a surface only a metre or so in dimension. The effect of these small scale fabrics is clearly seen in thin section.

Finally, in the flat-lying beds, sets of micro-faults are frequently densely represented on the scale seen in Figure 37. These faults are of normal geometry, and indicate a distension of the beds in the B direction.

Discussion of Overall Kinematics and Dynamics

Summarising the description detailed above, we may state that the Fisherstreet Slide is a plate of sediment probably up to 20 m thick, which displays outward systems of thrusts and recumbent folds which in general indicate a stress normal to the trend of the structures. The stress must have been high in relation to the viscosity of the sediments. This is in contrast to many other slump structures in County Clare, which frequently show clean fractures in materials which appear to have had reasonable cohesion and consolidation even at shallow depth (Figures 16 and 17). The structures at Fisherstreet are of highly fluid style, closely resembling many kinds of metamorphic structures. There are indications, however, that the structures in the slide are not due simply to stress normal to the folds and thrusts.

Firstly, there is the almost universal grooving effect, particularly discernable in the flat-lying beds, which is indicative of bed-over-bed movement in a direction sub-parallel to the B axis of the major structures. This southeasterly movement of the entire plate is evidenced by the fold lying obliquely to the southerly boundary thrust, (Figure 30, Plate 22) which would permit movement only in that direction.

Secondly, there is the core of the fold (Figure 33) which is hooked by a drag effect, again indicating a southeasterly component of movement. The cause of the relatively high confining stress is to be seen in a shorewards projection of the converging boundary thrusts of the slide, which would meet at a point some three kilometres from the coast section. The energy for the model must come from the plate itself, and is envisaged as depicted in Figure 38. It should be emphasised that beds corresponding to this horizon are known to be extensively slumped in the same style over a wide area inland and that the Fisherstreet Slide shown in Figure 38 represents a large slide within this extensive deformed sheet.

The main mass of the slide, extending for an unknown distance seawards, moved in a south-easterly direction and flowed outwards, resulting in local stress resolution along its edges. The outwards thrusting appears to have resulted in formation of topographic highs, which, on relaxation of strain, formed gliding structures moving in the opposite direction to the primary thrusting.

Probably the most important lesson to be learnt from the Fisherstreet Slide is that the fold axes of slump folds of this type do not in this case, and probably do not in many other cases, lie perpendicular to the basin slope. A simple relationship of this type has frequently been invoked in the literature, but, the Fisherstreet example clearly shows that determination of basin slope from slump folds can only safely be made when the kinematics and dynamics of the folds are clearly understood. Obviously, therefore, it is essential, wherever possible, to differentiate between folds of compressional origin with a metamorphic style, and those of free gravitational, gliding style, which are clearly directly controlled by local basin slope.

The contemporaneous nature of structures on the scale of the Fisherstreet Slide has frequently been strongly challenged and distinguishing criteria are required. A conclusion which may be drawn from the Fisherstreet examples is that where normal sedimentary rock shows metamorphic structural styles then the structures are pre-lithification, and therefore very probably contemporeous.

g) Extensive Thick Sheet Slides: The Ross Type

The Ross Slide is exposed in an intermittent series of bare-rock platform exposures for about three kilometres along the north coast of the Loop Head Peninsula, including the well-known tourist spot of the Natural Bridges of Ross. This is Locality A (Figure 1) on the general map of Clare, and a detailed map appears in Figure 39. The Ross Slide is one of several slumped horizons which occur within the Ross Sandstone Formation whose general stratigraphy has been outlined on page 10.

In the Loop Head Peninsula there are probably up to 360 m of bedded sandstones between fossil bands with *Homoceratoides prereticulatus, Reticuloceras circumplicatile,* and *Reticuloceras paucicrenulatum* horizons. This interval is the clear equivalent of a few metres of black shales in North Clare, as described by Hodson (1954A) and by Hodson and Lewarne (1961). The horizon of the Ross Slide is thus somewhat lower than that of the Fisherstreet Slide. The lithologies are typically parallel bedded sandstones with beds up to 2 m in thickness, often composite units, alternating with dark grey to black silts and bands of black marine shale 0.5-1.5 m thick. Detailed maps are shown in Figures 39 and 40, a stratigraphic section in Figure 3, and detailed cross-sections of the slide are given in Figures 41, 42 and 43. The slide is overlain and underlain by parallel bedded sandstones and generally consists of three distinct units:–

Top Layer of Sandstone Rubble, often with sub-angular fragments

The overlying bedded sandstone unit has remarkable slip scars, to be described later, which clearly indicate the subaqueous sliding origin of the top rubble layer (though not from slides visible in the present section). As indicated by the sections, the top rubble layer is not everywhere present.

Black Silt

This is up to 6 m thick and frequently has a deceptively simple appearance in which the stratification appears to be intact. Detailed examination reveals intense small scale deformation in the form of micro-faults, as shown in plan in Figure 41. Several shear planes can be seen near the base of the silt and it is deduced that basal sliding occurred along a marine black shale horizon which contains nodules with goniatites. The occurrence of sandstone blocks of various sizes embedded in the shale and silt layer, as shown in Figure 40, reveal the extent of the deformation of the silt layer. In several places this simple parallel bedded arrangement is replaced by an imbricated structure of much-contorted sandstone and silt, as exemplified by Section F-G, Figure 41, and at Locality A5, Figure 39. The imbrication on Section F-G clearly indicates a movement approximately north northwest-south southeast. Locality A5 shows the remarkable structure seen in Plate 26 and Figure 44. This gives a clear kinematic picture of the bulldozing action of the slide, rucking up the bedded sandstone at the base, and folding it into the very clear locked structure of Figure 44. Here, this local movement resolution is in a southwesterly direction. Peripheral to these imbricated bulldozing masses, the silts are folded after the manner seen in Section C-D (Figure 41). These particular structures have a north northwest-south southeast axial trend, and are regarded as bow-wave structures surrounding the imbricated sandstone masses. In the imbricated zone, extremely complex flow folds are common.

Basal Rubble or Carpet

This shows an interesting range of deformations and it is clear that the base of the slide itself cuts across several beds of the underlying sandstone group. The common appearance is that shown in Plate 27 which shows the small-scale folds with a trend west southwest-east northeast, clearly colinear with the trend of the major imbrications on Section F-G (Figure 41). Elsewhere, the basal rubble is scraped off and moulded into the cores of fold structures and imbrications reflecting the tectonics of the overlying siltstones. Clear examples of such a moulded core are seen in Plate 28, and others are shown in Sections C-D and F-G in Figure 41. Thin sandstone dykes are often visible in the basal sandstone beds (Figure 43). The linear elements of the Ross Slide are summarised in Figures 42 and 43.

The imbrications and fold cores clearly show a direction of movement from north northwest to south southeast. Sandstone dykes, micro-faults and folds in the silt layers do not have such clear linear components. Evidence in support of the north-south movement is available from the remarkable slip scars in the bedded sandstones which overlie the Ross Slide. Figure 46 is a drawing of the section at Locality A8 (Figure 39) with the tectonic element removed. This reveals a series of southerly facing slip scars with a succession of sandstone beds deposited abutting against the slope. These slip scars can be traced for a considerable distance in the cliff. The rock platform at Locality A1 (Figure 39) shows, in the lower part of this section, a southerly facing slip scar with attendant basal breccia and an overlying sand bed which has obviously undergone some movement. The top bed at this locality suggests a sheet-like movement, with imbric thrusts and recumbent folds also with the same orientation as other imbricated structures at Ross.

The intense folding at Ross happens to have a closely similar B-axis orientation to the slump imbrication. Due to the tectonic later phase of folding, there has been much slickensiding of quartz impregnations which were emplaced along all planar elements and cavities. It has taken much time and documentation to differentiate these two distinct structural components. However, the conclusion is that the Ross Slide has a similar direction of movement to the Fisherstreet Slide, although the morphology of the structures is quite different.

2. Shearing slides (growth faults) in the Central Clare Group

As has been stated in the section on stratigraphy, there is an abrupt change in the pattern of stratigraphic development in County Clare at the top of the Gull Island Formation; from here upwards the succession is characterised by cyclothemic deposits. This change is clearly recorded in the columnar sections in Figure 7, where the most important cyclothems are named. These are the Tullig Cyclothem, the Kilkee Cyclothem, and the Doonlicky Cyclothem. Rider (1969) recorded five cyclothems in all.

Each cyclothem consists of a basal black marine shale with a rich, but restricted, marine fauna. The fauna was controlled by euxinic conditions represented by black pyritic mud and comprises nectonic goniatites. Succeeding this relatively small interval there is a large thickness of banded silts, again dominantly dark grey in colour, and rich in organic matter. The silts are succeeded by a laminated sandstone facies diagnosed as distal bar and distributary mouth deposits. Commonly, the cycle terminates with a channel sandstone of varying thickness regarded as being an on-delta fluvial deposit. The thickness of the cyclothem varies from about 100 m to about 30 m for the sub-deltas, and in each area they are regarded as prograding units.

Early studies by the author showed that below clear unconformities were faults in the sandstone facies at the top of the cyclothem which had the typical geometry of gravitational faulting. It was also recognised from a very early stage that moderately thick sands are preserved on the downthrow side of these faults which are not present outside the area of fault influence and that current transport of the sands is almost invariably in the down-dip direction of the faults.

Rider (1969) was the first person to recognise that each distinctive lithological part of the cyclothem had its characteristic slump geometry, namely steep normal faults in the top sandy part, flow folds in the middle part and flat imbricate shears near the base.

The Clare coast has cliffs rarely more than 30 m high. In the northern part of the county, the dip is very flat and therefore, except for the thinnest cyclothems, one cannot see the top and the base of the cyclothem in continuous section. Further south, where the fold amplitudes are greater, one can see truncated sections of slump forms, but without any degree of horizontal continuity of exposure. Therefore, for the geometrical considerations of slump forms in the thick cyclothems there remained considerable elements of doubt about the overall integrated geometrical forms. Rider (1969) was able to recognise at Donegal Point (Figure 22), in a 35 m thick sub-delta of the Doonlicky Cyclothem, the clue to the integrated geometrical form and dynamics of the slumps in the cyclothemic rocks. Using this small model it has been possible to interpret the thicker cyclothems of West Clare, and to compare them with the well-documented, much larger scale analogues in Louisiana, Texas and Nigeria. It seems, in fact, that the scale of the model reflects the depth of water into which deltas are built. The larger the scale of the model the more difficult it is to integrate the top and the bottom. Indeed, in Louisiana and Nigeria, the basal parts of the model are still largely in the realms of conjecture. The mini-models of County Clare will therefore be of some interest to those who are contemplating the large scale events of Louisiana and Nigeria.

The dominant geometrical aspect in all these models is that of a fault which starts at the surface at a fairly steep angle, becoming progressively flatter, sometimes almost disappearing as a structure sub-parallel to bedding and at other times with a rising toe structure.

The name "growth fault" given to these faults in Louisiana relates to a purely chronological status as a syndepositional structure. This syndepositional aspect is simply related to the fact that it is a subaqueous structure which is connected with the pattern of deposition and in which the down-dip hollow created by the fault is most frequently infilled with sand. The word "growth fault" seems to have been first used in the literature by Ocamb (1963); but it seems probable that the concept was in fairly wide circulation before this time. The more precise definition of these synchronous faults was admirably clarified by Meyerhoff *et al.* (1967). The fascinating feature of Rider's (1969) work in County Clare was that he evolved diagrams like Figures 47 and 48 without being familiar with their larger-scale counterparts in the petroleum provinces of Louisiana and Nigeria. These generalised models are almost unbelievably like their better-established counterparts in various deltaic environments. A further point is that the megascopic models in Louisiana and Nigeria were thought to have some connection with the undoubted existence of evaporites at the base of the succession. In the small models from County Clare, it can be demonstrated that evaporites are not necessary to the development of these structures.

The structures selected for description are in the order of increasing structural intensity and can be regarded as a morphological series. An attempt is made to show that developments at specific stratigraphic intervals, i.e. specific cyclothems, establish the connection between the geometry of the structures and the sedimentological and other features of palaeogeography.

a) Simple Rotational Shears

Example 1: Donegal Point Complex

Although perhaps not the simplest of the Clare models of shearing slide structures, this is nevertheless the locality at which the essential features of this type of development can best be examined and integrated. Because of the small scale of the cyclothem it is one of the few places in which the total structural geometry can be clearly demonstrated. A detailed map and section with directional elements of structures and palaeocurrents are given in Figure 22. The tectonic dip is 10° to 15° to the southwest. The cliffs on the south side of the promontory are a little over 15 m high. The top part of the cyclothem therefore is well exposed along the southern cliffs and the base is exposed in Smugglers' Cove at Bealnalicka, where the basal marine band rests on the top sandstone member of the underlying cycle. The lithologies are black marine shales at the base succeeded by siltstones, laminated silty sands and, near the top, foresetted sandstones.

At the top of this section it can be seen that there are four distinctive south-southeast dipping major faults and many minor structures with the same orientation. The most northerly fault (Major Fault 1, Figure 22), with its attendant small faults, dips at an angle of about 30°. This dip is seen to flatten within the exposure range. The infill of sediment, sand volcanoes, and other features at this exposure make it one of the most important localities in County Clare for demonstrating the syndepositional nature of these structures.

The second important feature of this fault is that the thickness of foresetted sandstone on the downthrow side greatly exceeds the thickness on the upthrow side and the foresets are almost precisely in the down-dip direction of the faults.

Major Slump Fault 2 (Figure 22) is somewhat steeper in its upper part than the first fault but flattens visibly within the exposure range. It has 10 m of sandstone on the down-dip side compared with approximately 5 m on the up-throw side. The distensional graben and horst structures here do not parallel the major fault which may possibly be the lateral termination of an arcuate shear. This may account for the fact that the no back-tilt into the fault is apparent in this line of section.

The next fault in the section clearly dips in the opposite direction to the minor distension structures. This arrangement of opposite facing faults is seen in many localities in County Clare, including the Carrowmore Point (Locality H) described below, and there is still some doubt about the interpretation of some of them. In the case of Major Slump Fault 2, small scale imbrications at the base of the opposite-dipping termination make it seem more than probable that this is the toe of a rotational shearing slide.

Major Fault 3, dipping southeastwards, is also rather flat in its upper part but again preserves fairly thick sandstone on the down-dip side. It is thought that the opposite dipping faults which terminate this structure are also probably toe structures. With regard to the flatness of the structures, Merki (1972) has emphasised the role of compaction in changing the geometrical shape of such structures but it would seem that this would apply almost exclusively to the basal, shaly parts of the cyclothem, as will be seen in later descriptions. Many similar structures in Clare have quite steep angles in the upper part of the structure.

With regard to the oil trapping structures in major deltas, it will be seen that the structure on the upthrow side of Major Fault 2 has a close resemblance to a roll-over anticline, and that the structure bounded by Major Fault No. 3 and the toe of Major Fault No. 2, resembles the "back to back" structure figured by Merki *(op. cit.)*.

It has been possible to examine the structure of the basal part of this cyclothem in detail at one locality only. This is at the foot of the path down into Smugglers' Cove at Bealnalicka (map on Figure 22). The northern cliffs of the rest of Donegal Point are vertical and inaccessible. At this locality, however, the marine band at the base of the cyclothem is repeated three times by imbricate shears which rise to about the level of the first silty part of the section. They are projected on to the section B-C of Figure 22.

Elsewhere in the cliff, flow folds are evident, but no details can be resolved. To summarise, the evidence from Donegal Point would suggest that:

(i) The whole mass of the cyclothem has probably sheared at the base.

(ii) There exists a complex system of rotational shears, whose limits are bounded by major gravity fault systems. These are plainly synsedimentary and have many subsidiary distension fault systems with graben and horst type structures which frequently parallel the major fault systems. There is also a great abundance of small scale micro-tectonic features. The faults which dip in the opposite direction to the major ones described are thought to be toe structures at the rising ends of the rotational shears.

(iii) The systems of rotational shears give rise to geometrical shapes very similar to structures in Louisiana and Nigeria which have been called roll-over anticlines and "back to back" structures.

(iv) The syndepositional status of the faults is clearly established by:—

the nature of the top contracts:

by the development of thick sand bodies on the downthrow side of the major faults, correspondingly either very thin or completely absent on the up-throw side:

by the numerous sand emissions from the major fault planes and from sheets of sand which clearly point to the unconsolidated nature of the sediment during formation of the structures.

Example 2: Slumps below the Tullig Sandstone

A spectacular and instructive group of slump structures occurs in the area west of Doonbeg in the Killard/Pulleen stretch of the coast (Locality F, Figure 1). They occur at a horizon about

100 m below the base of the Tullig Sandstone in the uppermost part of the Gull Island Formation. In this area the interval is almost entirely represented by striped black and grey siltstones, and normally there are no interbedded sheets of sand. Sandstone, however, occurs in appreciable thickness on the downthrow side of the slump faults. A general map (Figure 49) shows the localities where slump faults occur. In the Blind Horses Cave area, Locality F, (Figure 49) a slump fault can be studied over the greatest horizontal distance (almost one kilometre) of any fault of this type. The rocks are steeply folded tectonically but are beautifully exposed.

Locality F1 (Figure 49) Blind Horse's Cave: The section in Figure 15 shows a typical and clear difference in the lithological succession on either side of the slump fault. The main fault can be followed with absolute clarity from the top where it is fairly steep, down to the lower part of the section where it is much flatter. On the downthrown side are flagstones and sandstones with siltstone beds. The sandstones and flagstones frequently show ball and pillow structure and the beds in the upper part have an appreciable dip into the fault. The major fault is paralleled by a similar rotational shear in the enclosing grey siltstones which is difficult to follow, on account of uniform lithology. It is also noticeable that flat lying, probably gravity, slides developed from the top of the fault face, and produced hollows which were later filled with sand. There are several bands of imbrication along the wall of the major fault indicating a failure by shearing of the fault wall. It would appear that this fault has had a complex history which cannot be totally elucidated in this section. The sandstone channel fill, the fault face and the imbricated structure of the foot wall of the fault are seen in Plate 29.

Locality F2 (Figures 49 and 50), Blind Horse's Cave West: At this point Rider (1969) has described two further slump faults of somewhat similar geometry to that of Locality F1 and probably at a closely similar horizon. They have a somewhat flatter aspect but have essentially the same features. On the downthrow side are sandstones, sometimes tilted into the fault, and frequently affected by ball slumping. The thickness of sandstone varies between 0 and 15 m and is nowhere to be seen either laterally or above, where the beds are dominantly siltstones, sometimes sandy silts. At Locality F1, the enclosing siltstones are folded and faulted in a very complex manner, and it is considered that these structures are synchronous with the major faulting. Sandstones on the downthrow side at Locality F2 display a great variety of small scale structures; distension horst and graben structures are very common particularly close to the steep-dipping fault planes which they parallel. Micro-faulting is common in the lower flat-lying parts of the fault planes, and large sand volcanoes occur along the top surface of the down-faulted sandstones. It must be emphasised that the overlying thick siltstones are completely undisturbed at localities F1, F2 and F3.

The directional elements of Localities F1 and F2 are given in Figure 50. It will be seen that the foresets and micro-foresets again display this remarkable derivation from the up-thrown part of the fault. At this locality the sediment continued to be derived from the same direction after the slumping had ceased. In the flaggy type of deposits, palaeocurrents for asymmetric and symmetric ripples show a direction parallel to the face of the main faults. It could be that the hollow on the down-thrown side of the fault caused deviations from time to time of the delta current system, but this factor is not clearly understood.

Locality F3 (Figures 49 and 50), Pulleen: The section is about 150 m long and exposes about 35 m of beds affected by fault slumping (Figure 51). There are two systems of faults, one involving the lower 15 m of the section (the base is not seen), and the other affecting the top 20 m. The two systems are somewhat mixed near their contact. As at Killard, the faults delineate abrupt facies changes. In the lower system, the rocks below the faults and enclosing them are banded siltstones which are tilted, faulted, and slightly folded. The faults clearly contain the development of appreciable sandstones on their downthrow side. The sands have large and small

ments and the succeeding cyclothemic patterns of deposition. In the non-cyclothemic slumps there scale cross-lamination, with some siltstone intercalations. The beds are completely affected by horst and graben structures. There is no appreciable difference between the fault directions and the palaeocurrents in the upper and lower parts of the systems. There are, however, two sets of cross- ... it seems that the southeast trending set was formed later than that

was a movement of large plates, with complex folding, shearing and imbrication, consistently southwards towards the axis of the trough of deposition, as indicated by the marked increase in stratigraphic thickness in this direction. The slopes responsible for this movement are therefore regarded as slopes of tectonic origin. In contrast, the slopes in the cyclothemic deposits are regarded as being depositional slopes of the fingers of delta channels, and the tectonic slope seems to have had no influence whatever.

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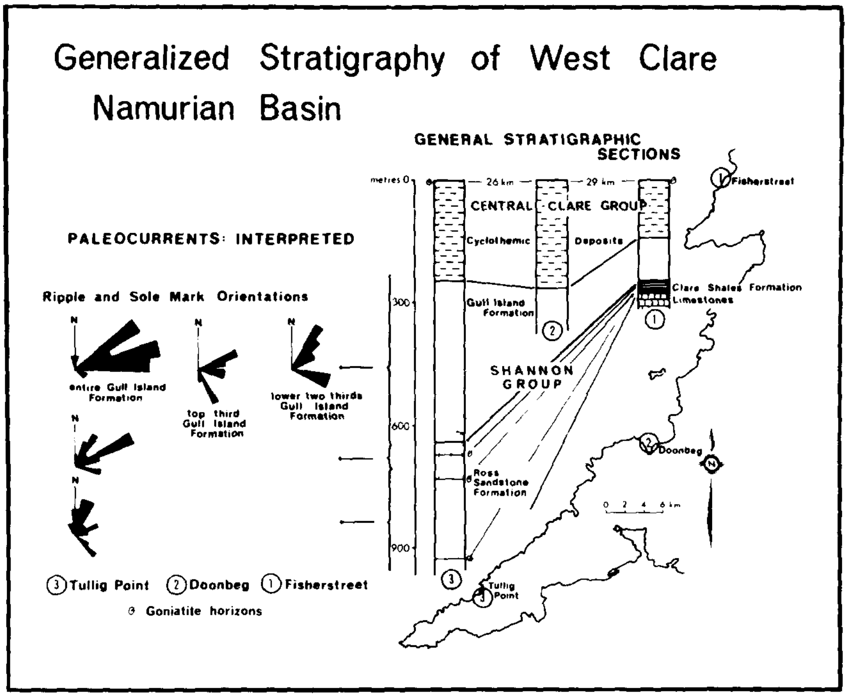

Figure 2. Generalised Stratigraphy of West Clare Namurian. (After Rider, 1969).

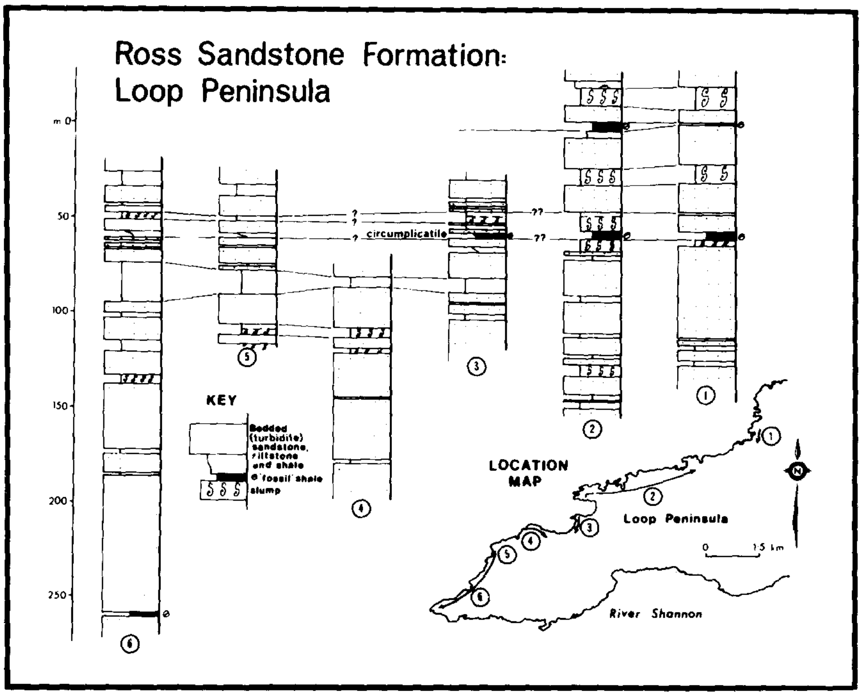

Figure 3. Ross Sandstone Formation: Loop Peninsula. (After Rider, 1969).

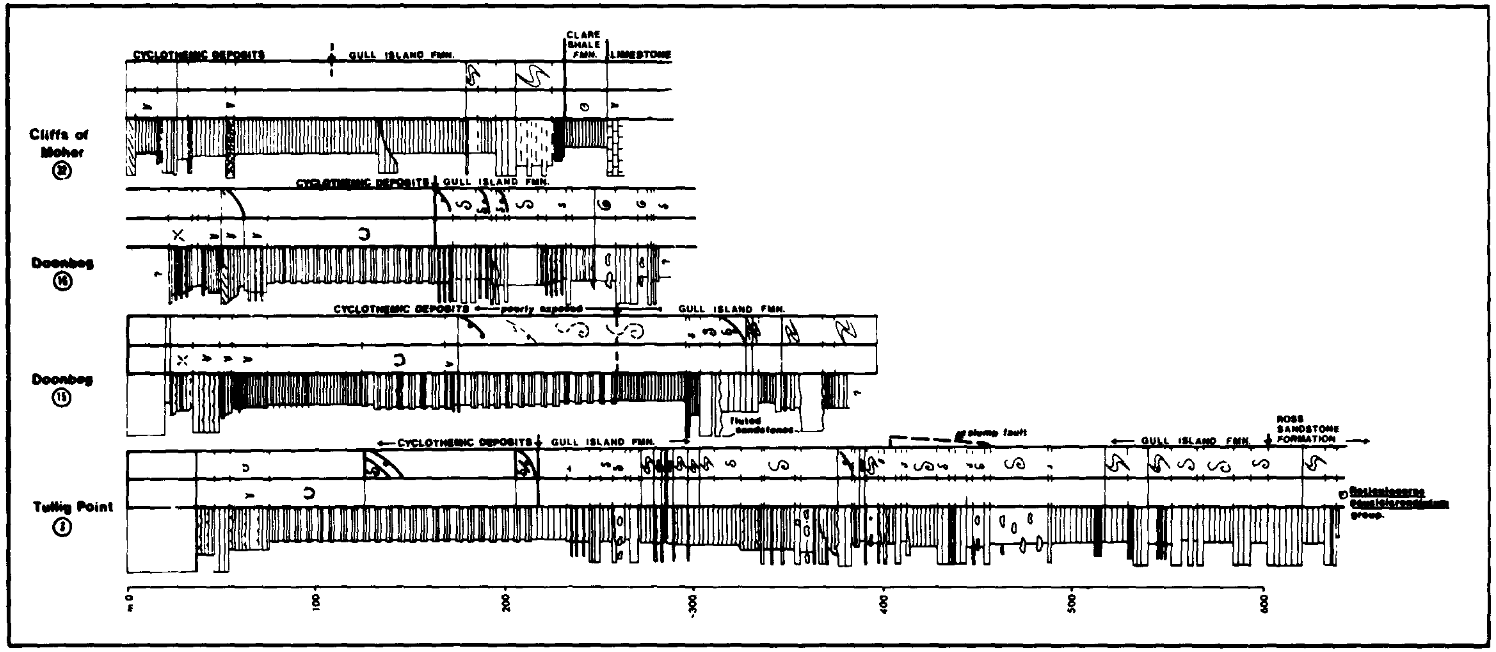

Figure 4. *Sections through the Gull Island Formation.* (After Rider, 1969). For locations see Figure 8. Legend as Figure 7.

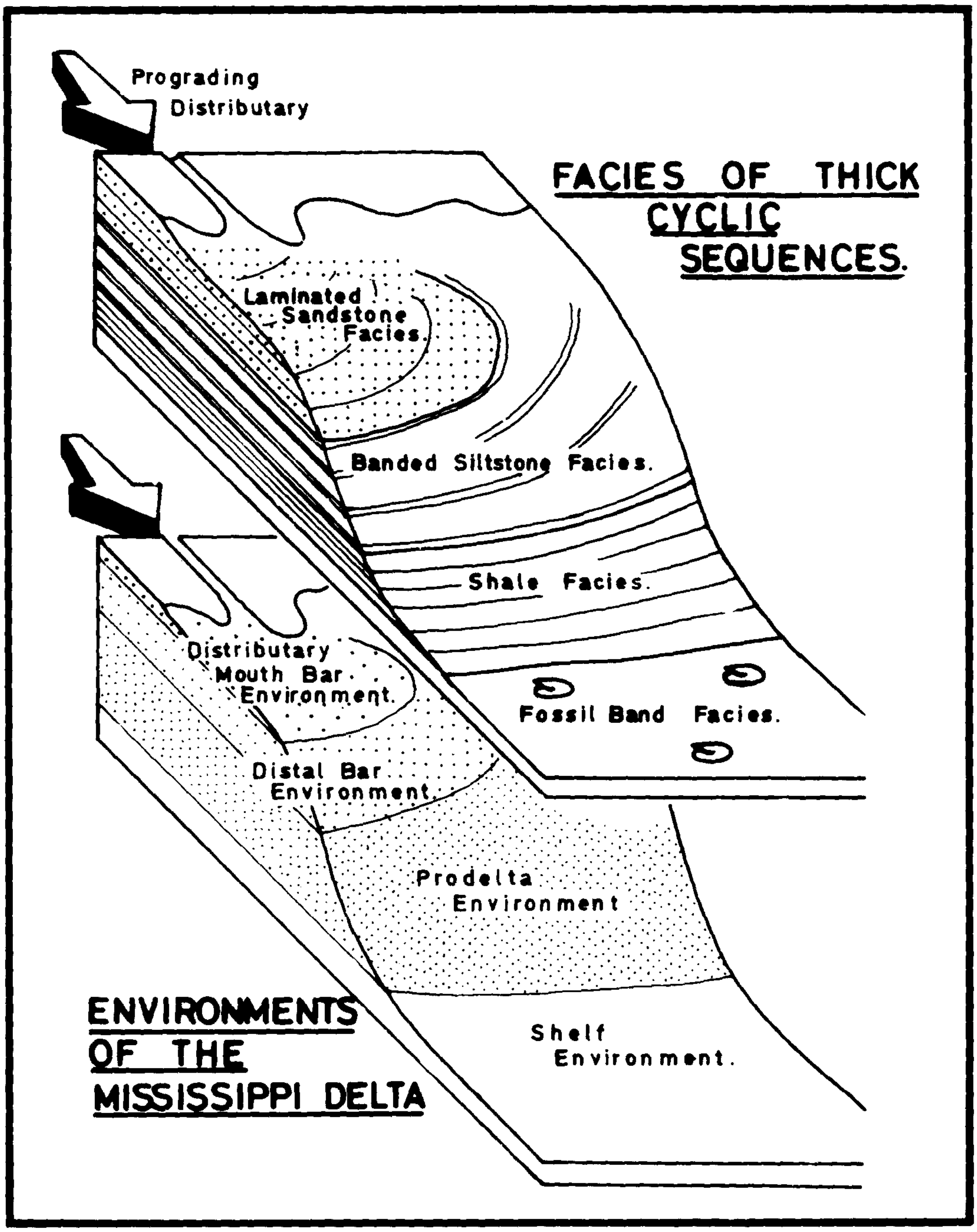

Figure 5. Facies of thick cyclic sequences: Environments of the Mississippi Delta. (After Rider, 1969).

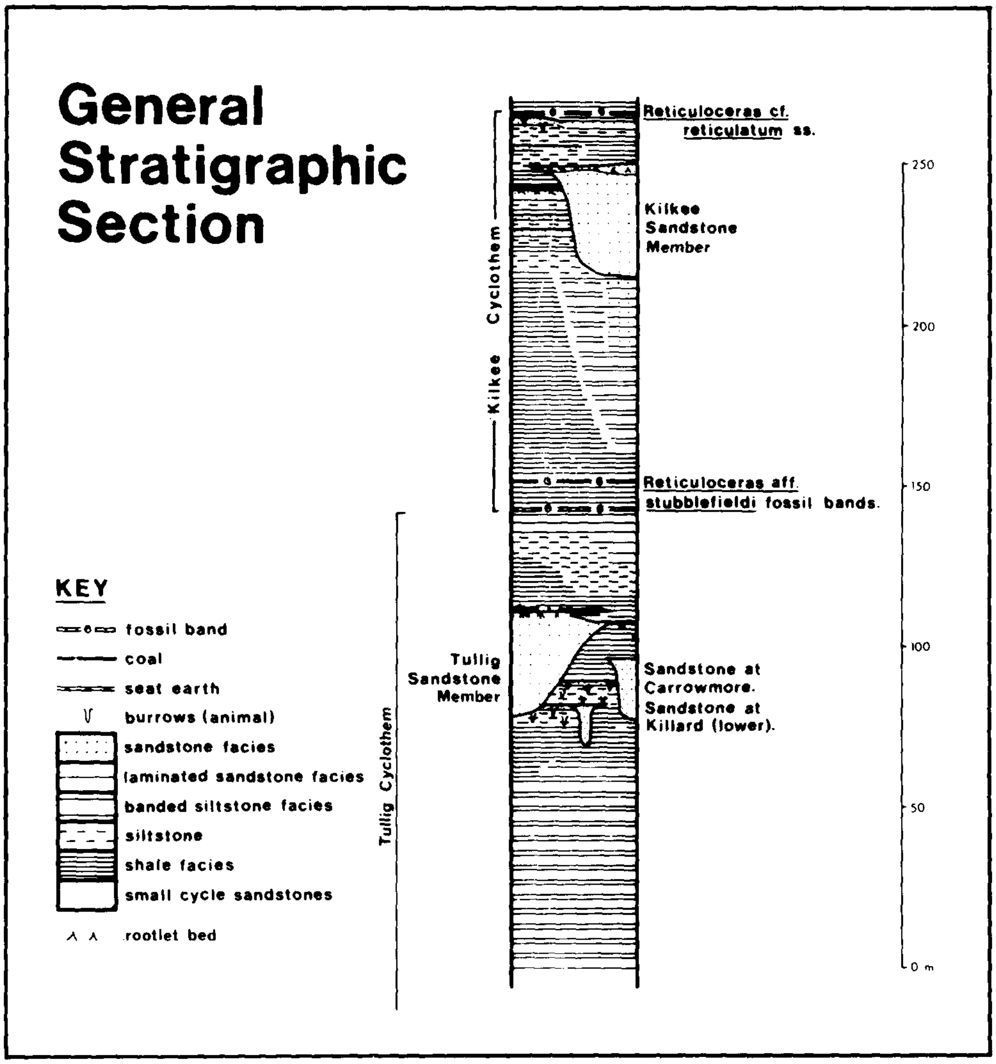

Figure 6. General Stratigraphic Section 1 Kilkee and Tullig Cyclothems. (After Rider, 1969).

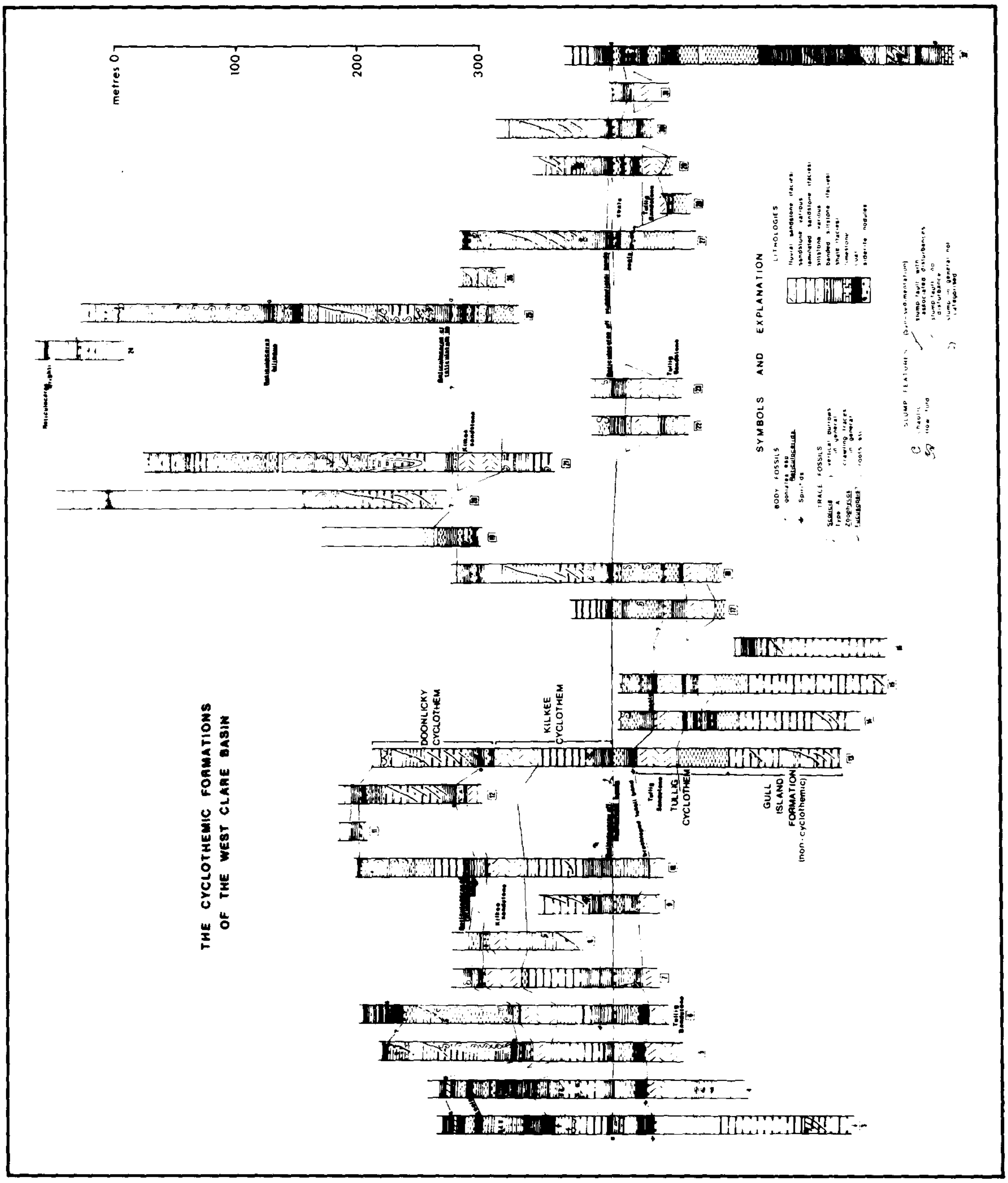

Figure 7. The cyclothemic formations of the West Clare Basin. (After Rider, 1969).

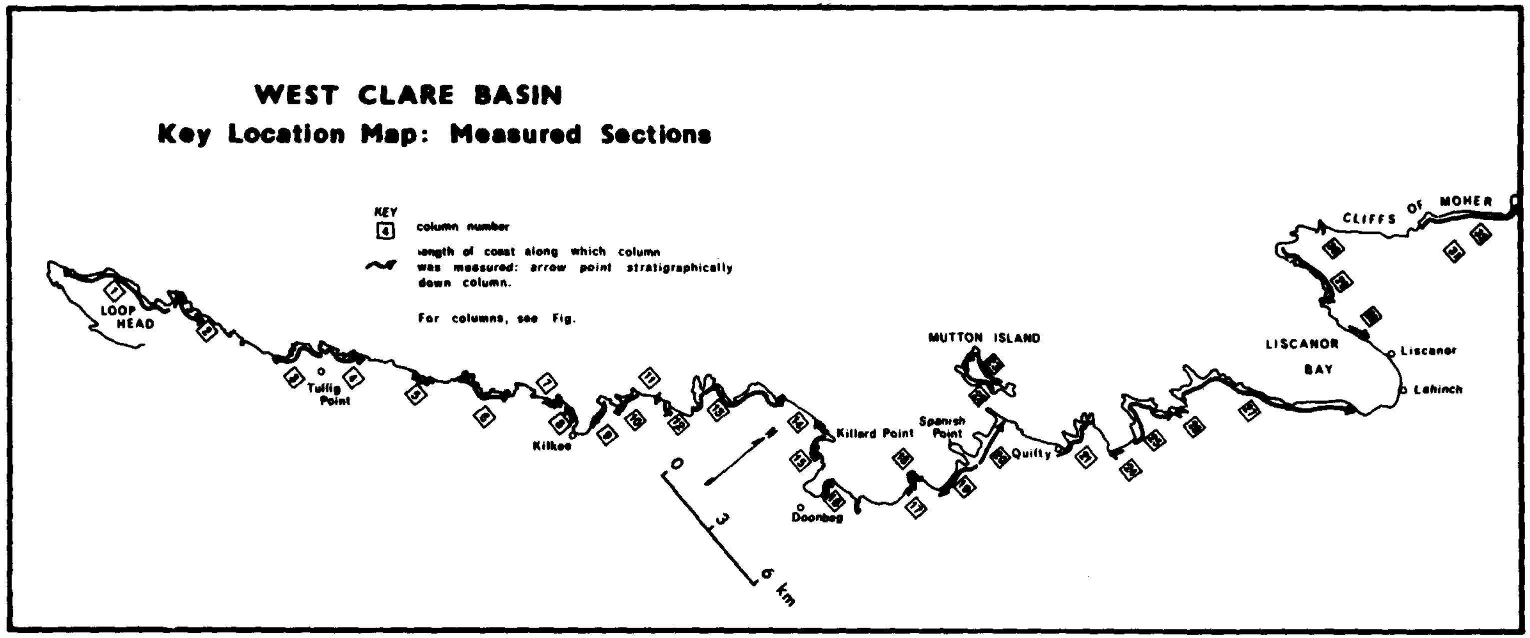

Figure 8. West Clare Basin, Key Location Map. (After Rider, 1969).

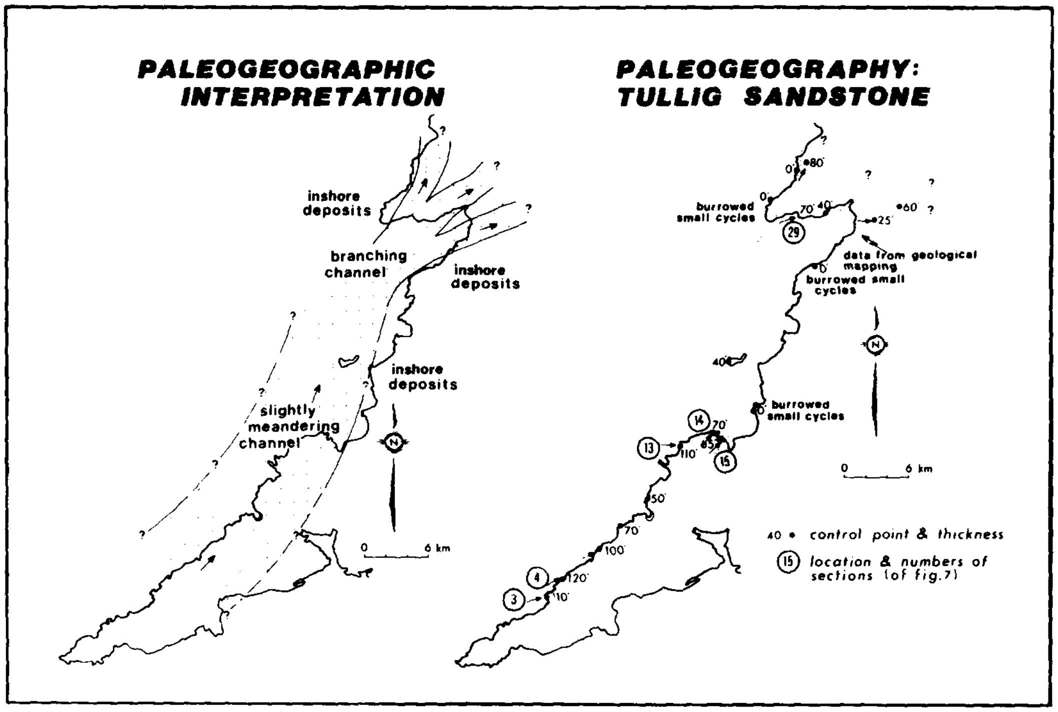

Figure 9. Tullig Sandstone. (After Rider, 1969).

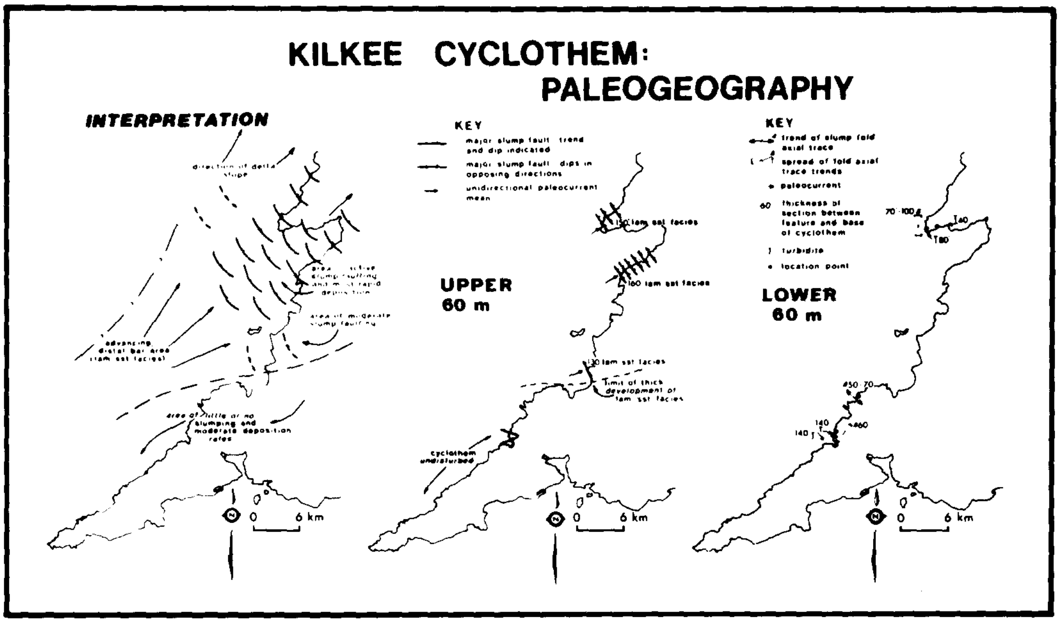

Figure 10. Kilkee Cyclothem: palaeogeography. (After Rider 1969).

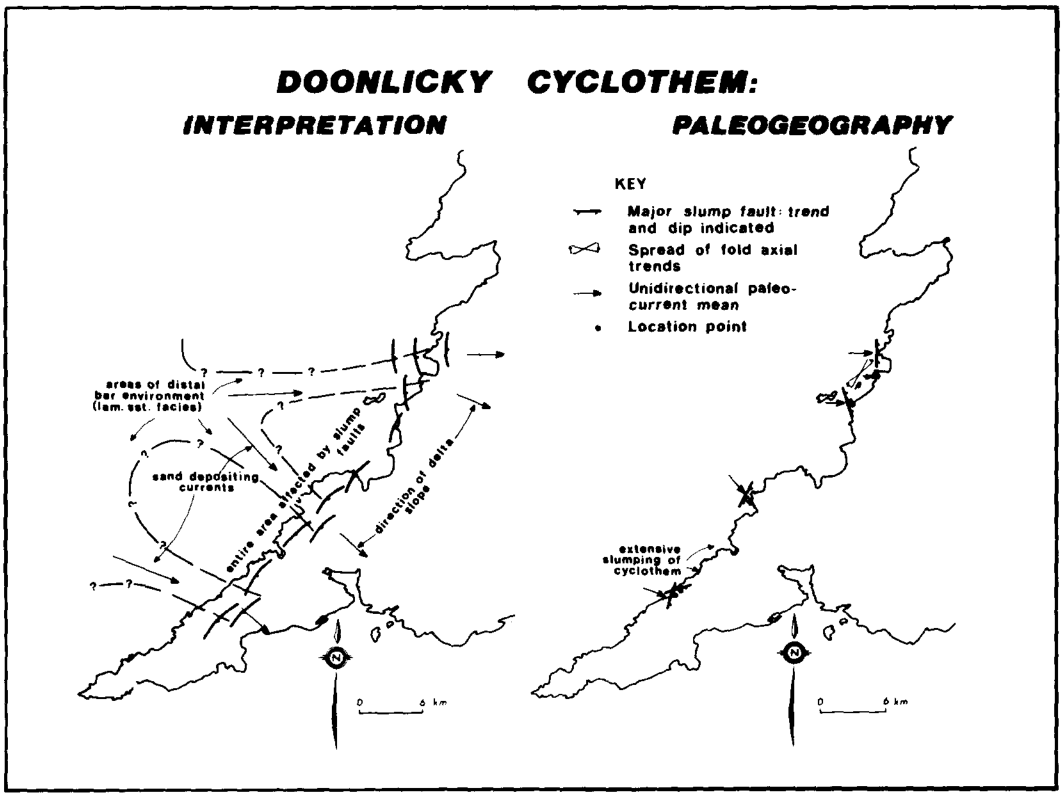

Figure 11. Doonlicky Cyclothem: interpretation, palaeogeography. (After Rider, 1969).

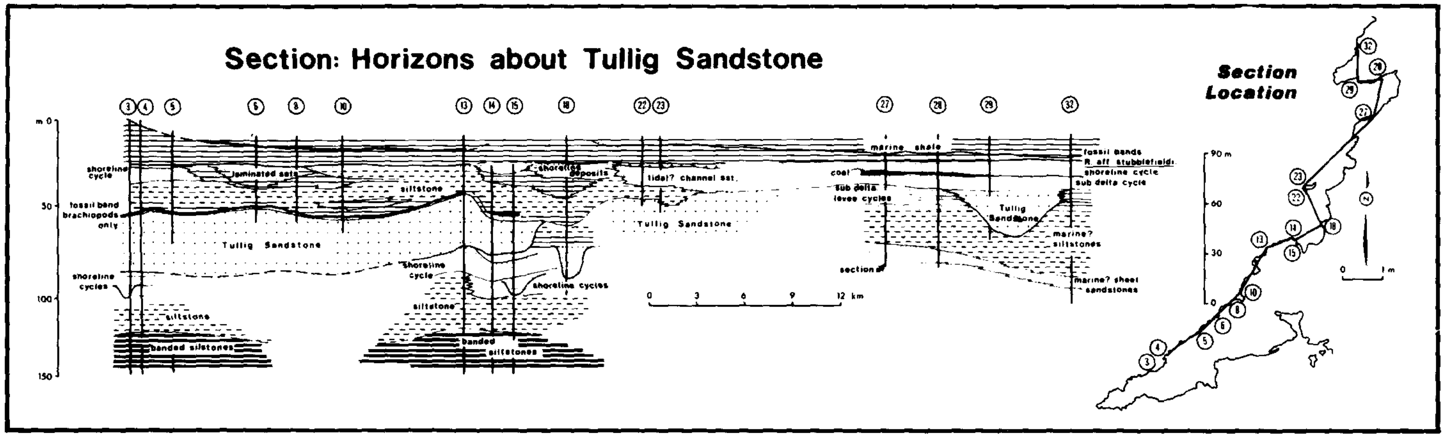

Figure 12. Sections: horizons about Tullig Sandstone. (After Rider, 1969).

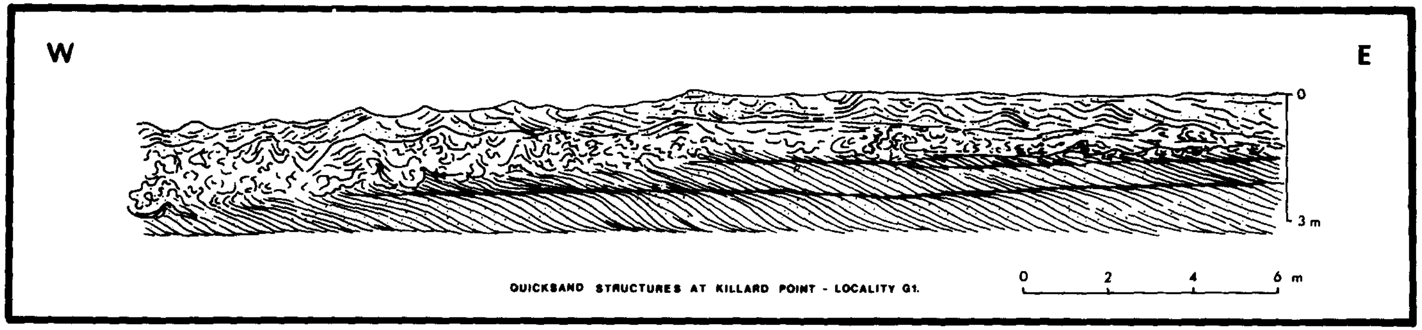

Figure 13. Quicksand structures at Killard Point. Locality G1, Figure 49 (true scale).

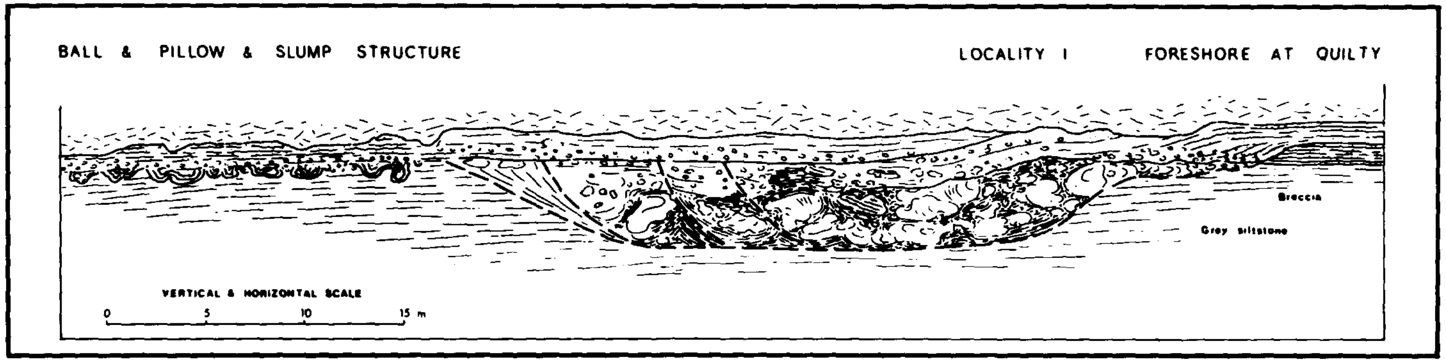

Figure 14. Ball and pillow and slump structures: Forshore at Quilty. Locality I, Figure 1.

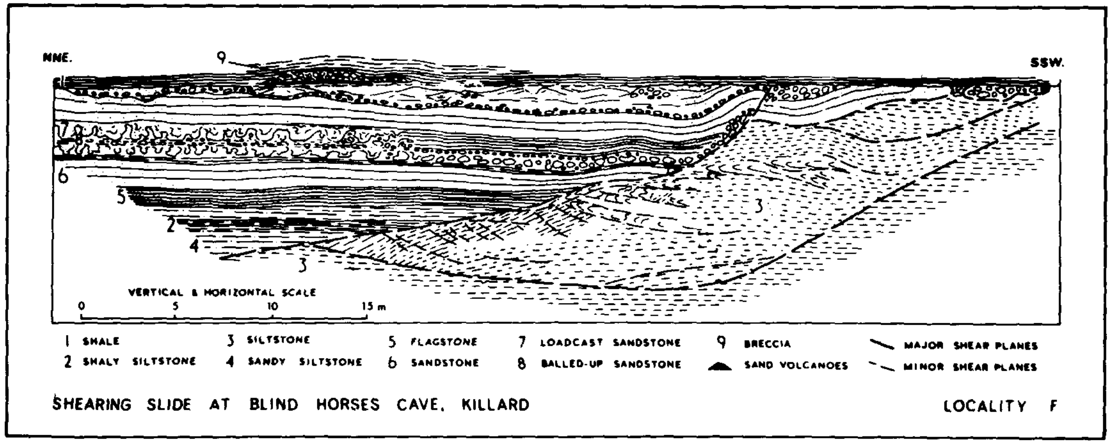

Figure 15. Shearing slide at Blind Horse's Cave, Killard. Locality F1, Figure 49.

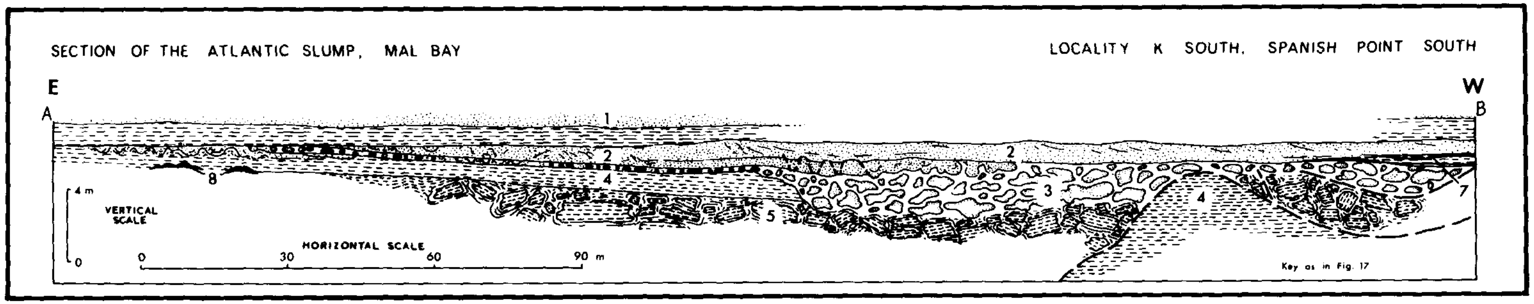

Figure 16. Section of the Atlantic Slump: Mal Bay. Locality K South, Figure 1.

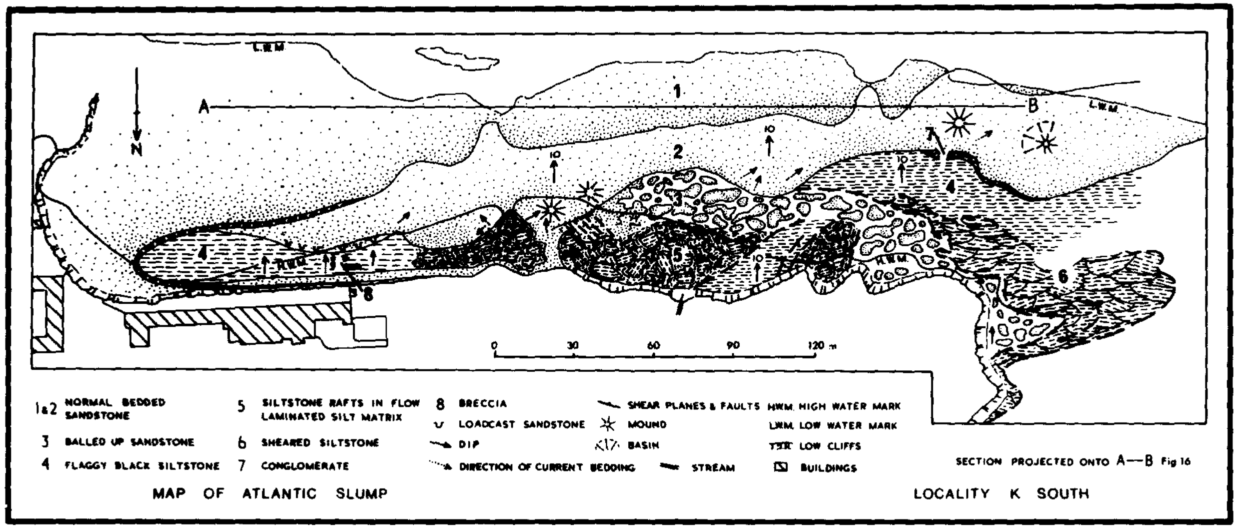

Figure 17. Map of Atlantic Slump: Locality K South, Figure 1.

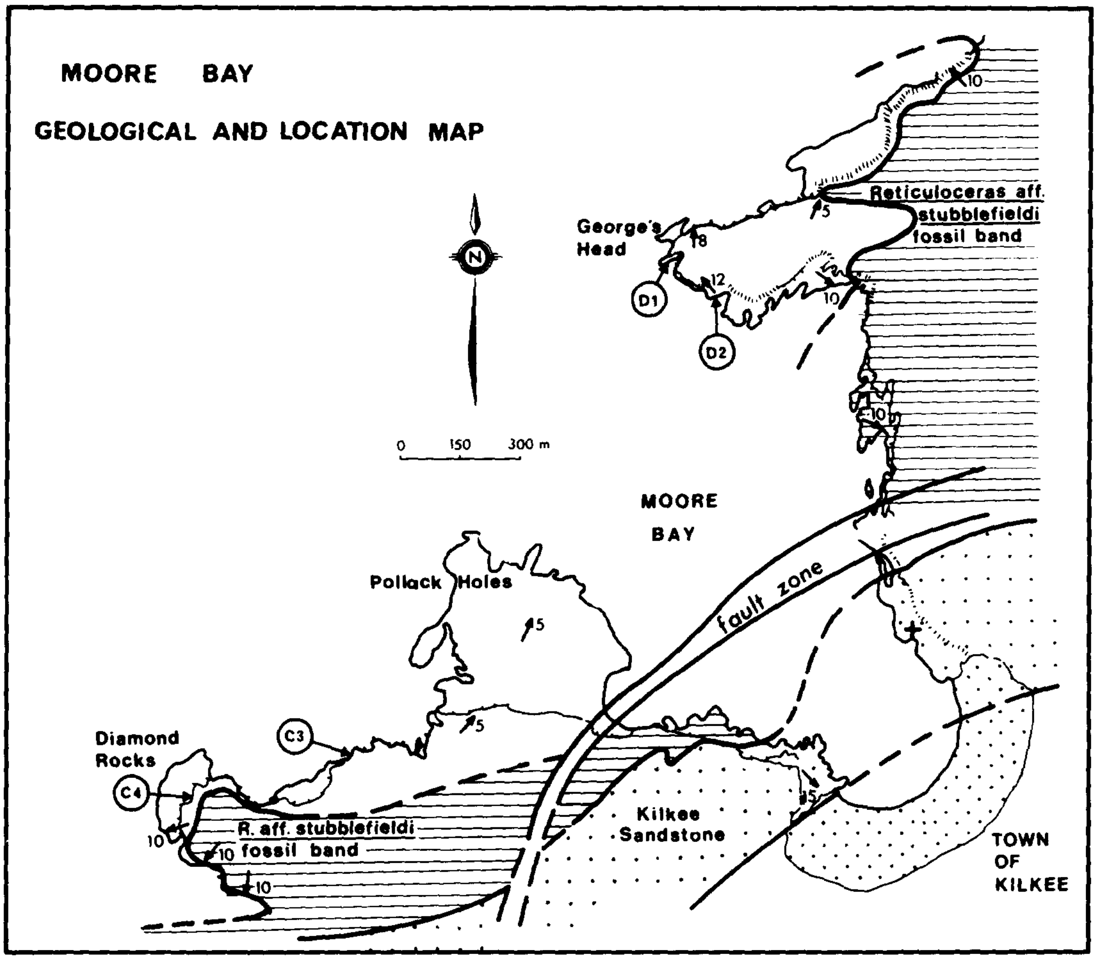

Figure 18. Moore Bay: geological and location Map. (After Rider, 1969).

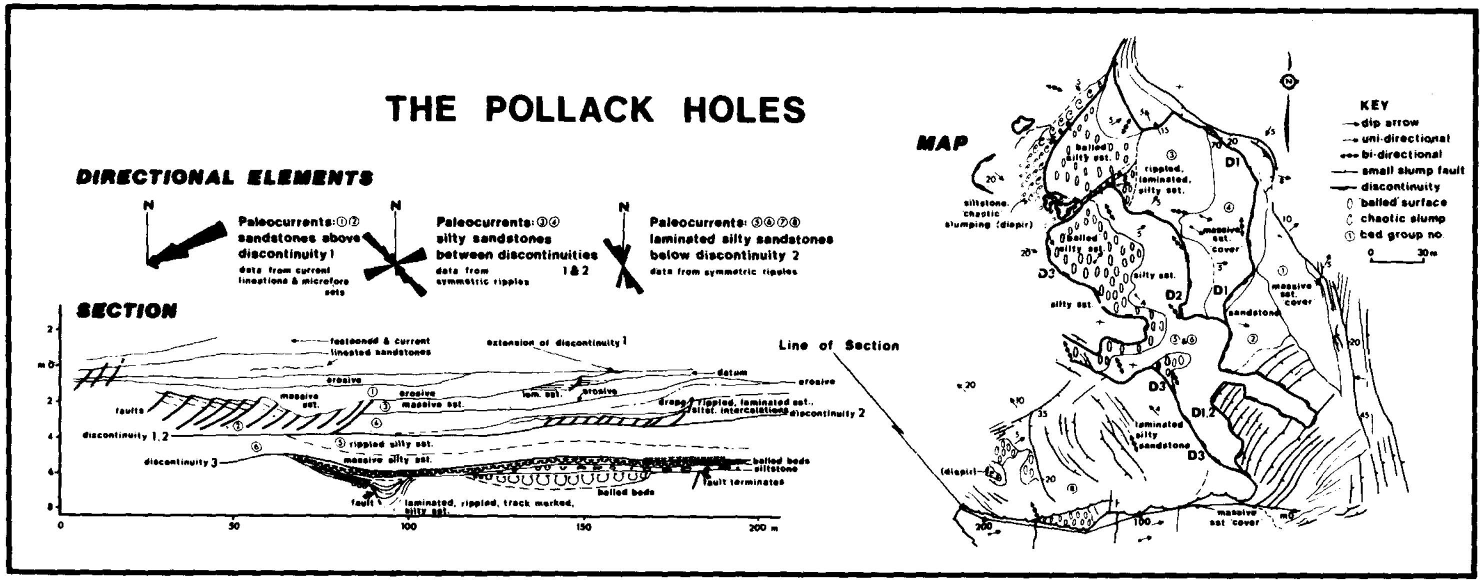

Figure 19. The Pollack Holes. (After Rider, 1969). Figure 18 for location.

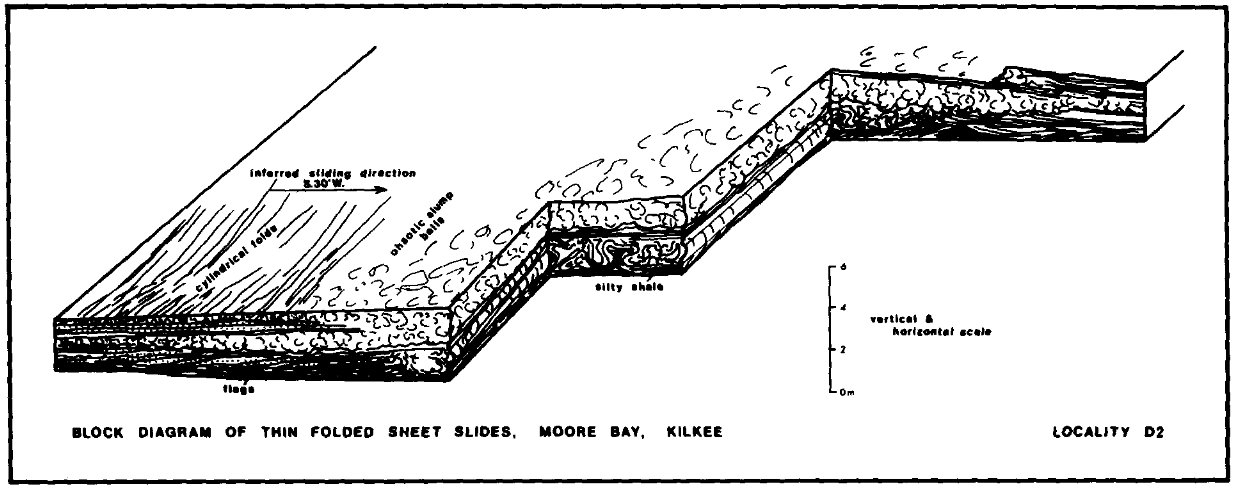

Figure 20. Block diagram of thin folded sheet slides: Moore Bay. Locality D2, Figure 18.

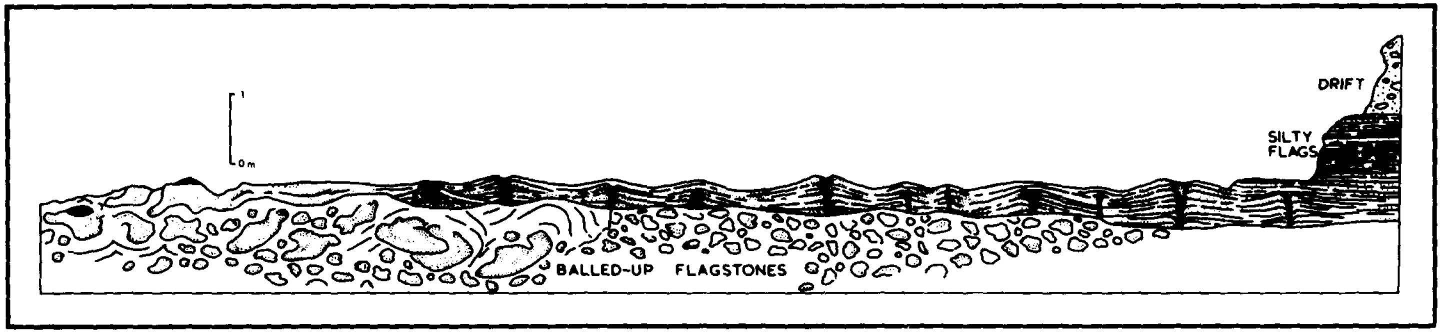

Figure 21. Sand volcano sheet at Freagh Point (Gill & Kuenen, 1958, Figure 2, reproduced by kind permission of the Geological Society of London).

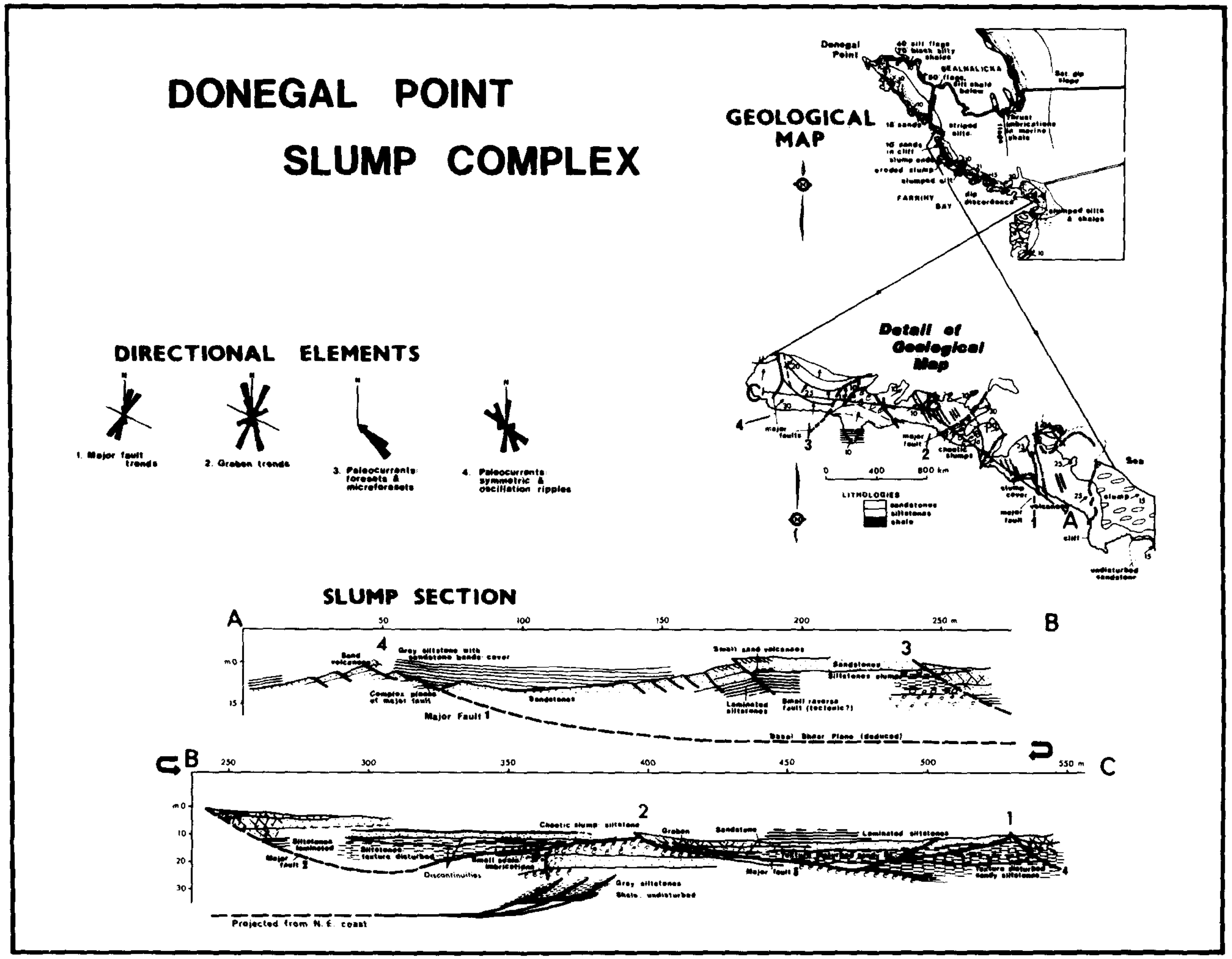

Figure 22. Donegal Point slump complex. (After Rider, 1969). Locality E, Figure 1.

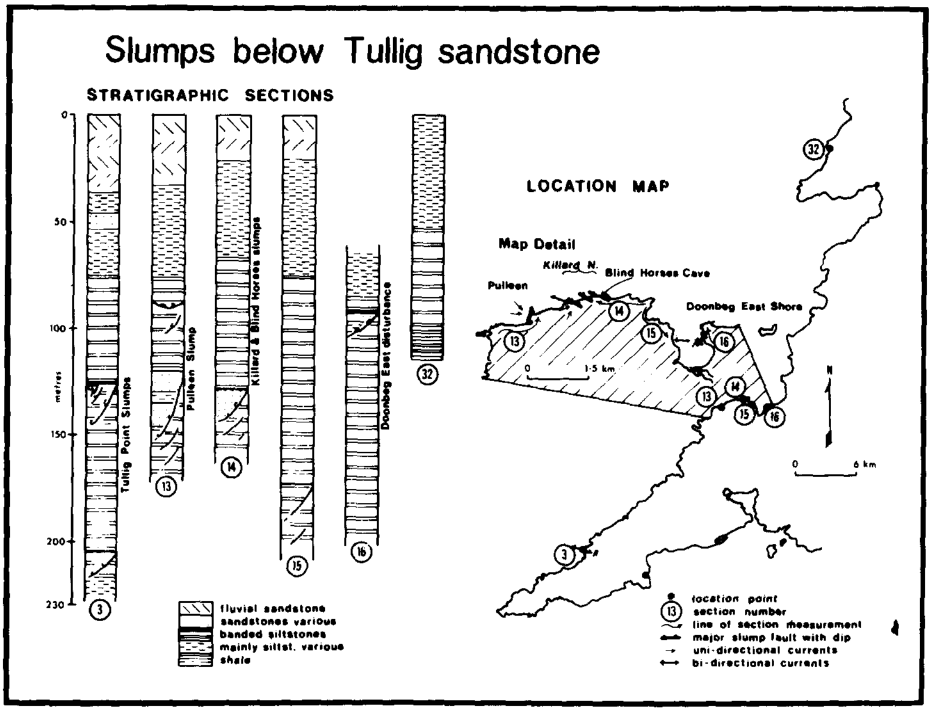

Figure 23. Slumps below Tullig Sandstone. (After Rider, 1969).

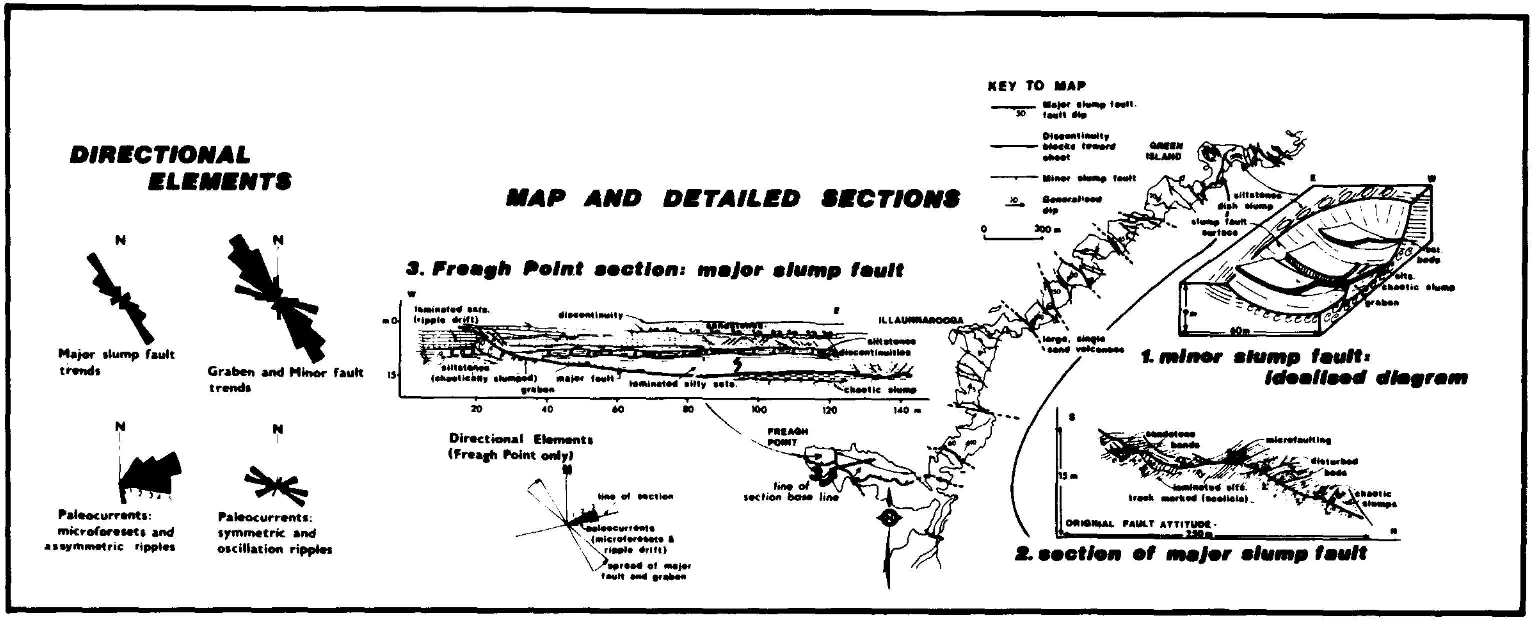

Figure 24. Map and detailed sections northeast of Freagh Point (Locality L, Figure 1). (After Rider, 1969).

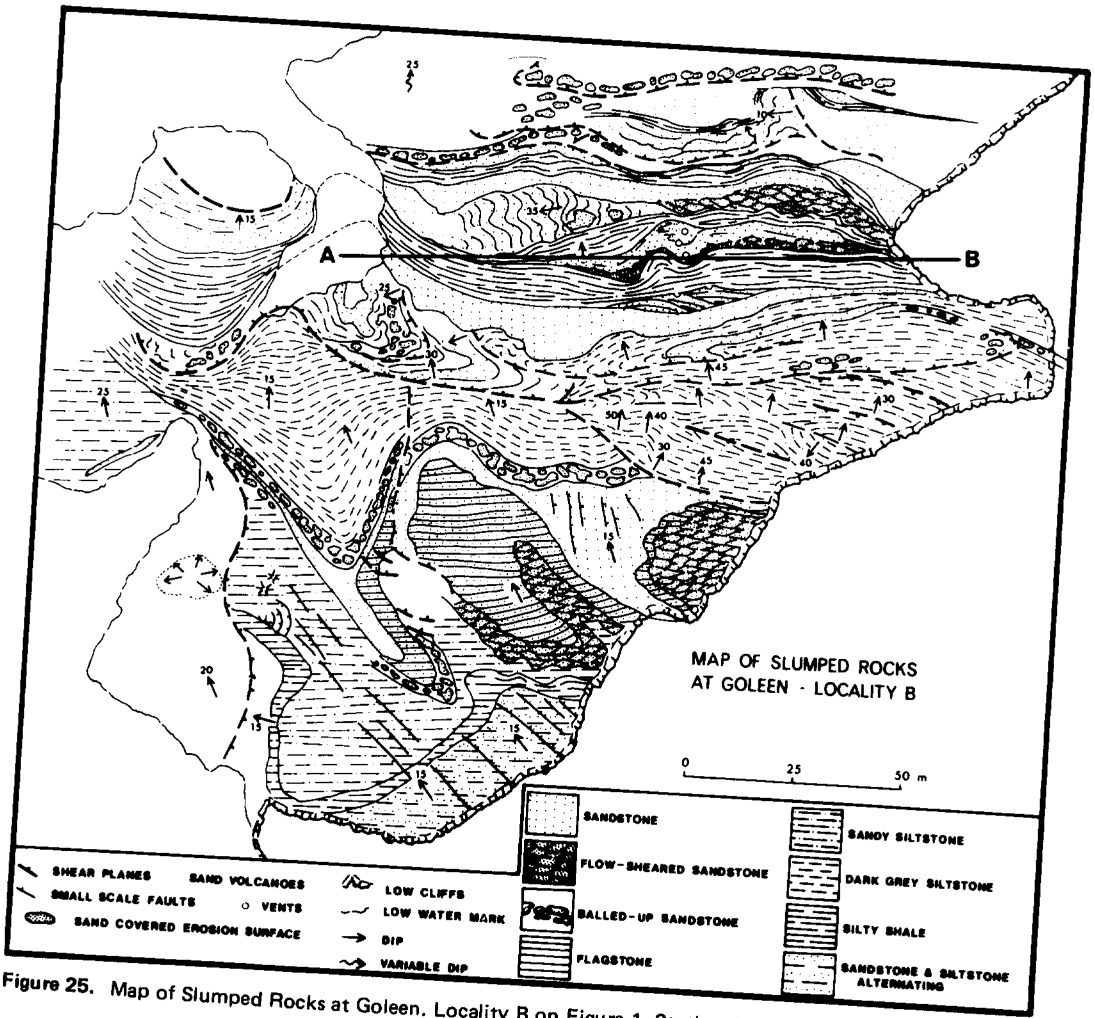

Figure 25. Map of Slumped Rocks at Goleen. Locality B on Figure 1. Section A-B shown on Figure 26.

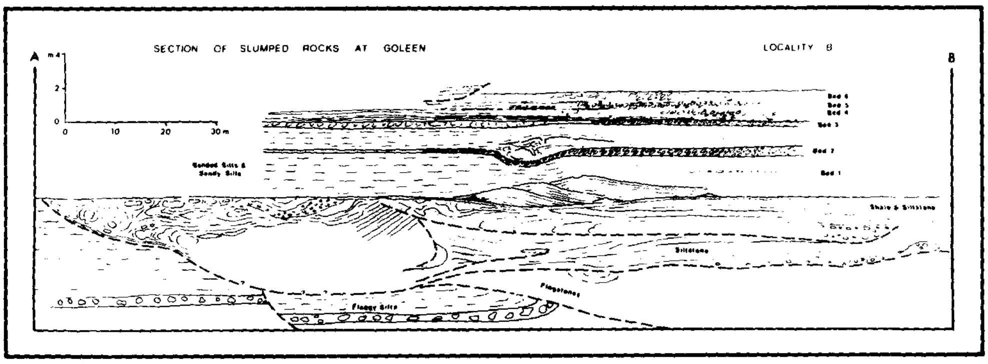

Figure 26. Section A-B of slumped rocks at Goleen. Locality B on Figure 1. See Figure 25 for section location.

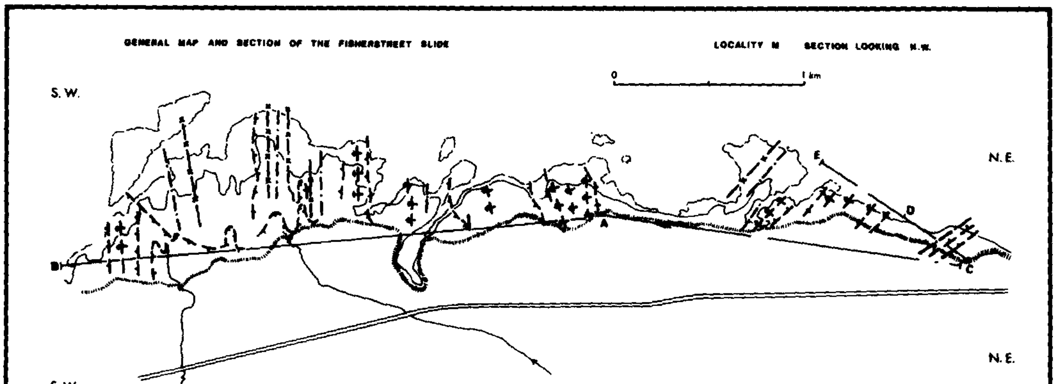

Figure 27. General map and section of Fisherstreet Slide. Locality M on Figure 1.

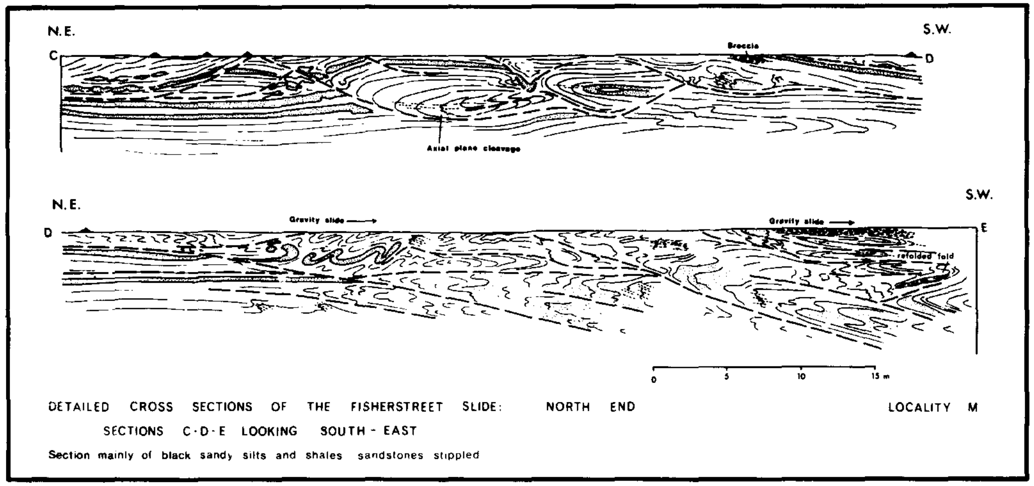

Figure 28. Detailed cross-section of Fisherstreet Slide. Locality M on Figure 1. For section location see Figure 27.

Figure 29. Detailed cross-section of Fisherstreet Slide. Locality M, Figure 1. For section location see Figure 27.

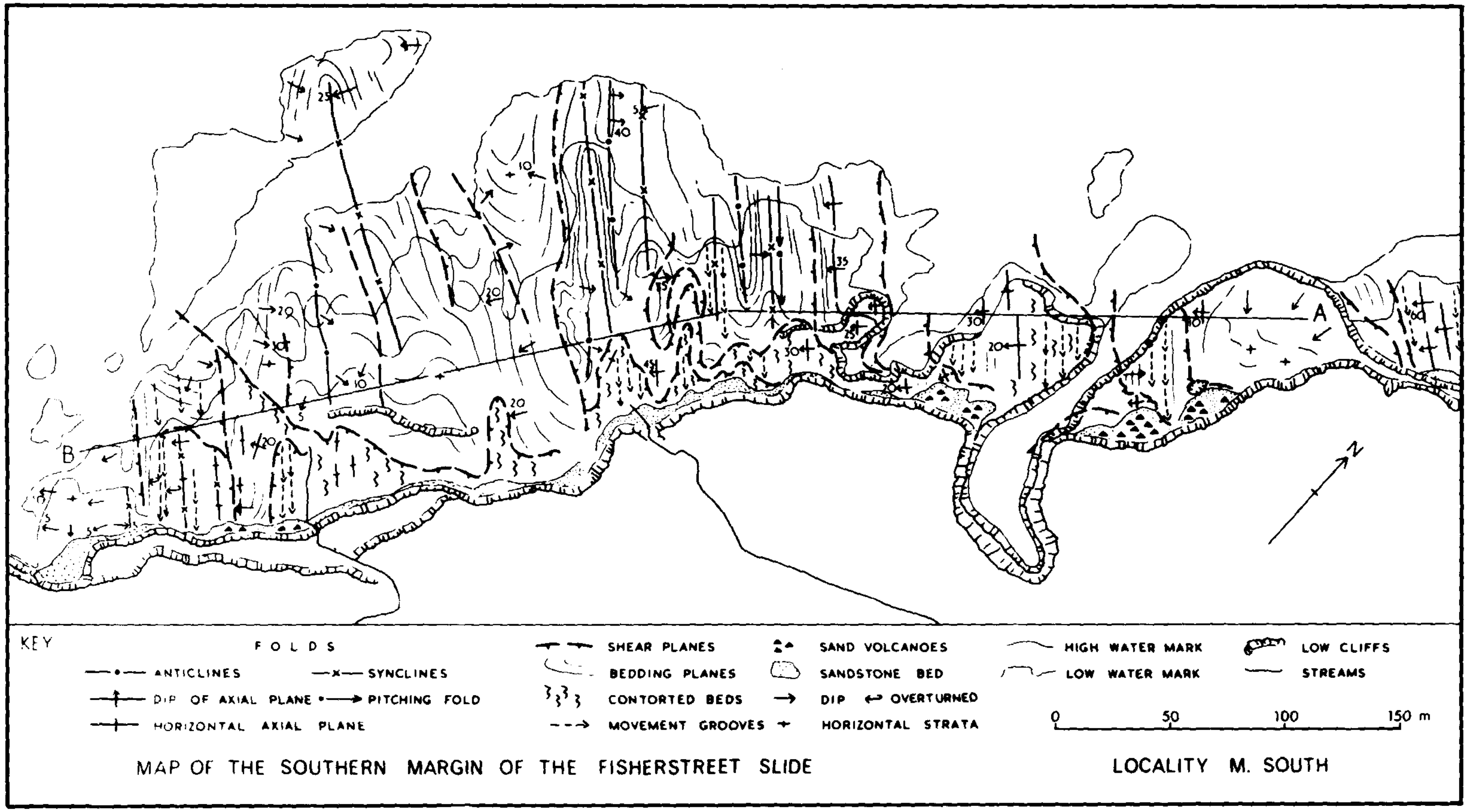

Figure 30. Map of southern margin of Fisherstreet Slide. Locality M, Figure 1.

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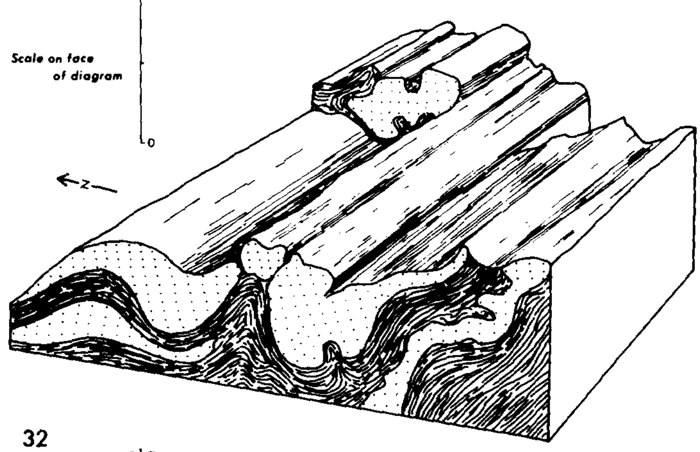

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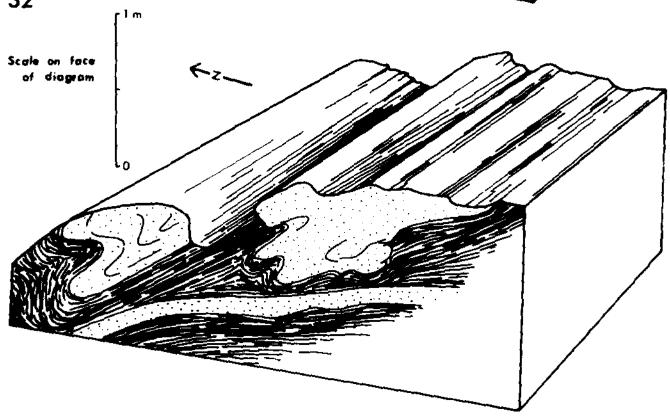

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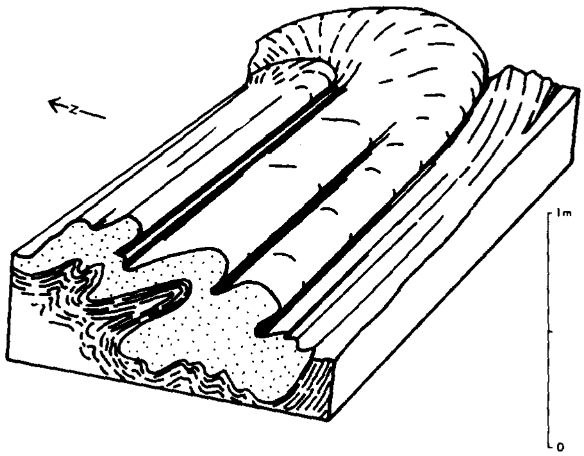

Figure 31. Fold Mullions in the Fisherstreet Slide. Locality M, Figure 1.

Figure 32. Fold Mullions in the Fisherstreet Slide. Locality M, Figure 1.

Figure 33. Twisted Fold Core. Fisherstreet Slide. Locality M, Figure 1.

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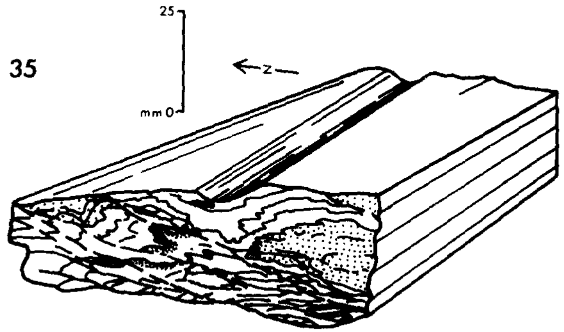

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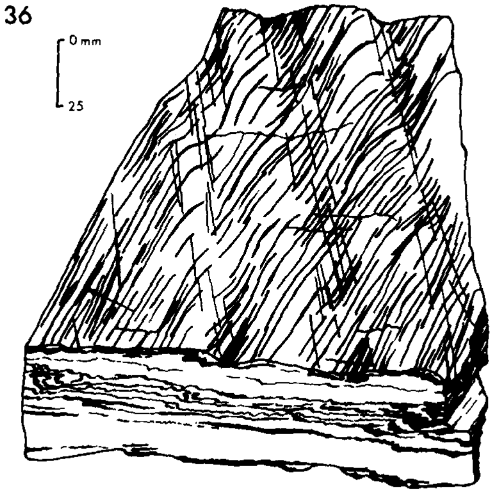

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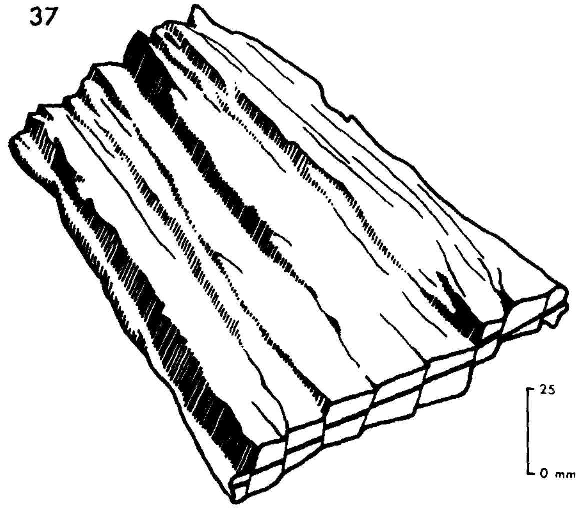

Figure 34. Fold frill: Fisherstreet. Locality M, Figure 1.

Figure 35. Small scale fold frill: Fisherstreet. Locality M, Figure 1.

Figure 36. Fold frill intersecting grooving: Fisherstreet. Locality M, Figure 1.

Figure 37. Sets of microfaults. Locality M, Figure 1.

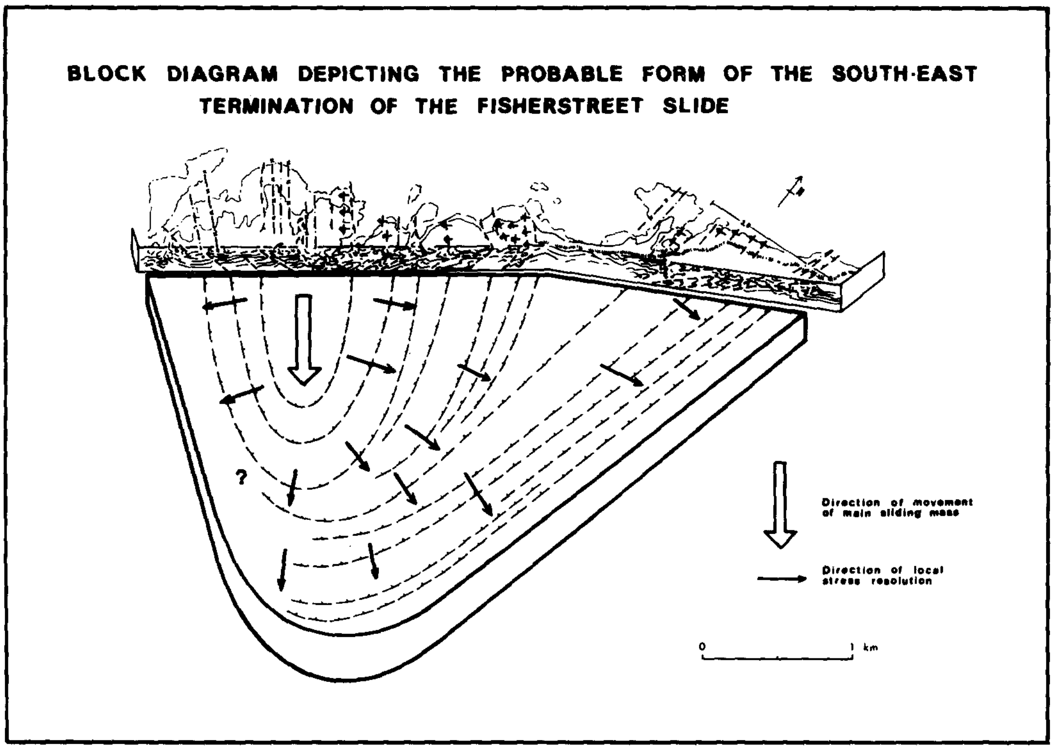

Figure 38. Block diagram depicting the probable form of the southeast termination of the Fisherstreet Slide.

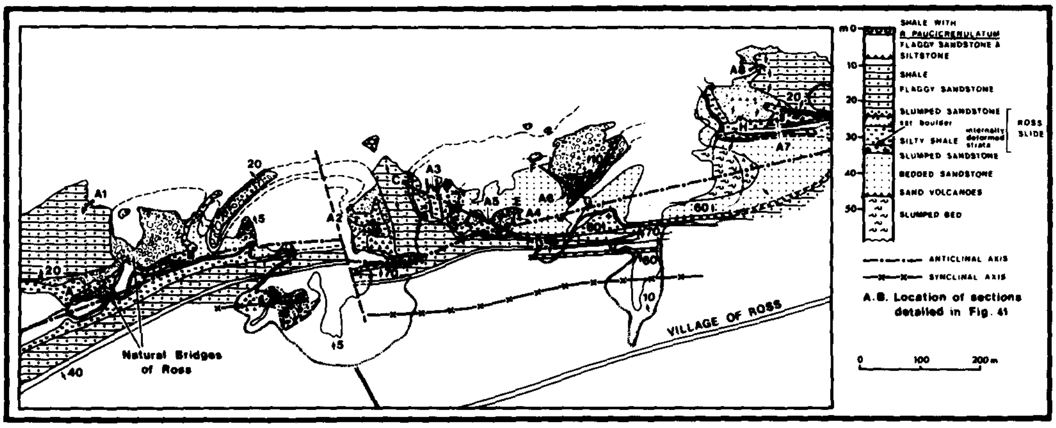

Figure 39. Map of Slump Sheet at Ross. Locality A, Figure 1.

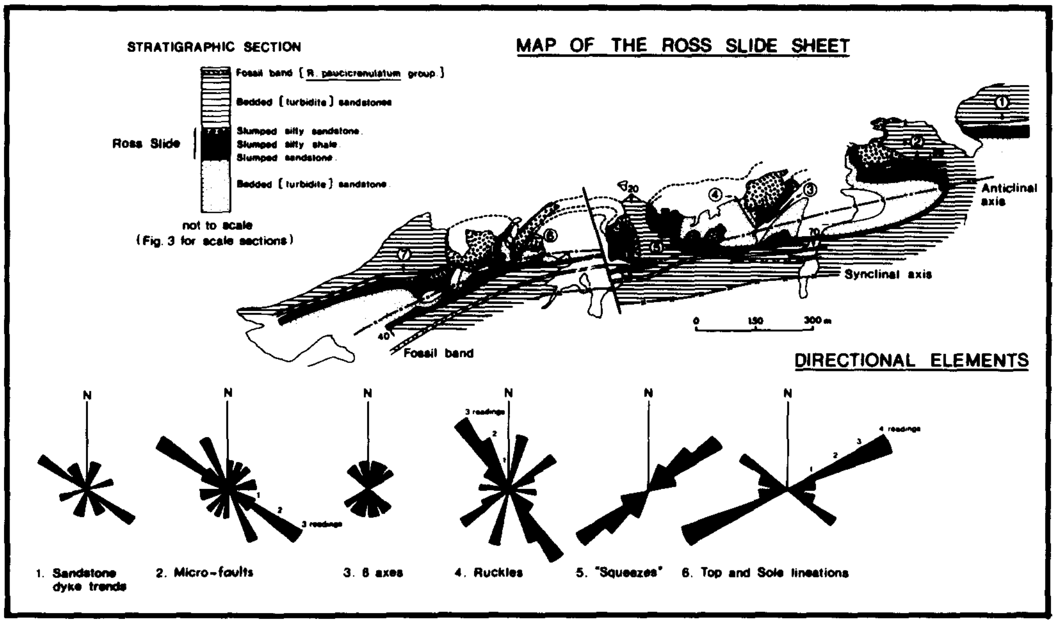

Figure 40. Map of the Ross Slide Sheet with section locations for Figures 42 and 43.

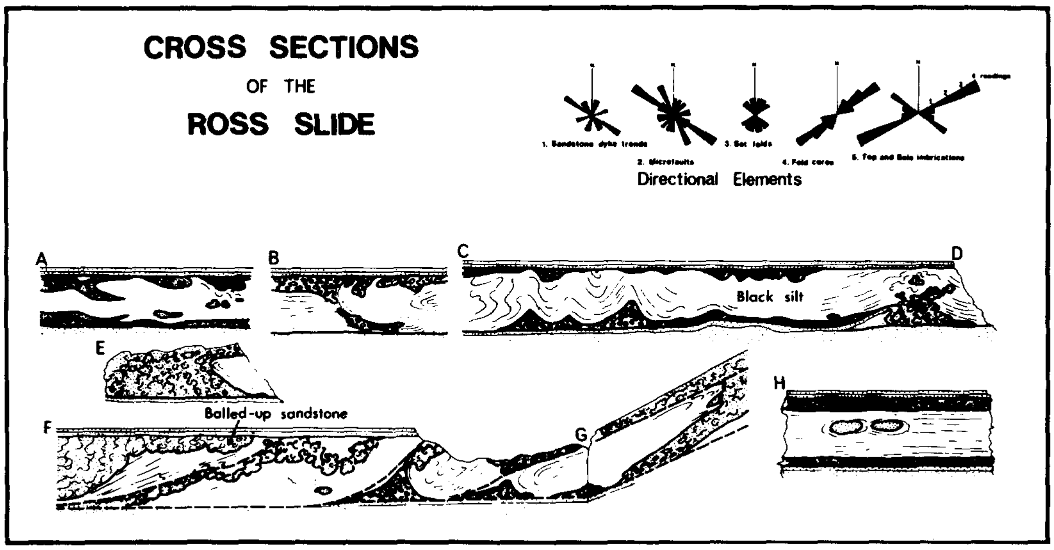

Figure 41. Cross Sections of the Ross Slide. Locality A, Figure 1. For locations see Figure 39.

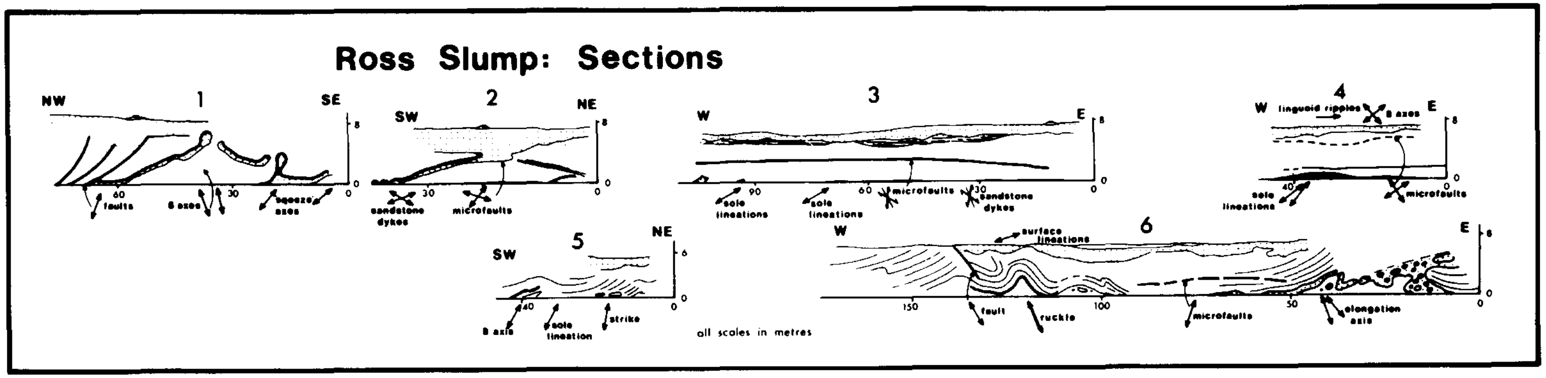

Figure 42. Ross Slump sections 1-6. (After Rider, 1969). For locations see Figure 40.

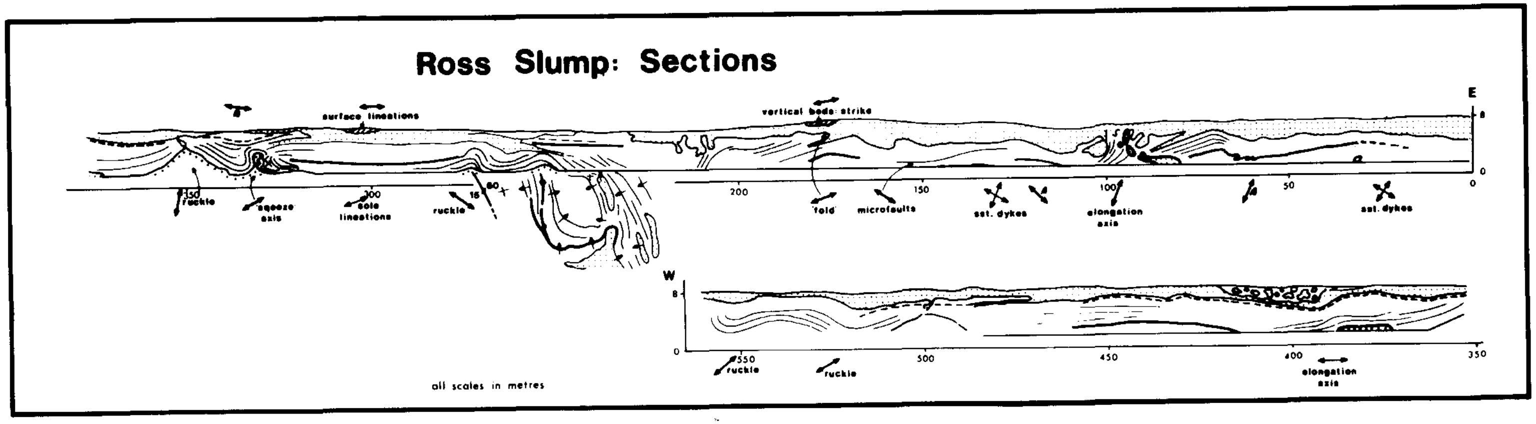

Figure 43. Ross Slump section 7. (After Rider, 1969). For location see Figure 40.

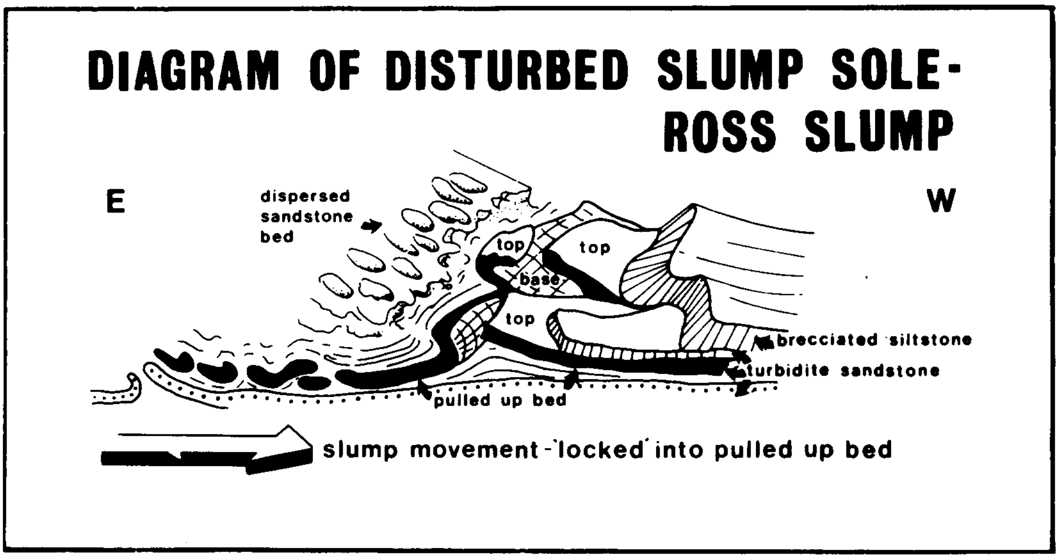

Figure 44. Disturbed Slump 'sole'. (After Rider, 1969). Locality A5, Figure 40.

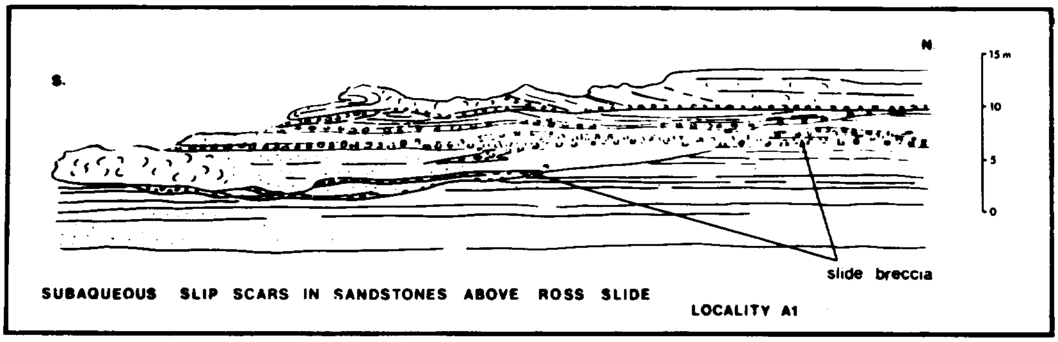

Figure 45. Subaqueous slip scars. Locality A1, Figure 39.

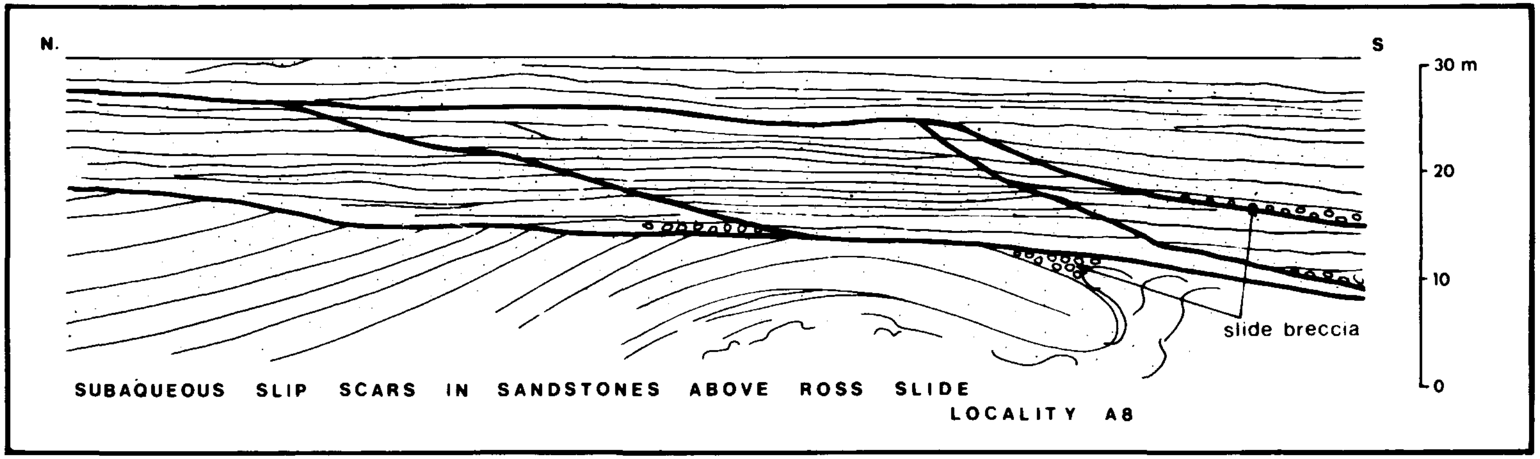

Figure 46. Subaqueous slip scars. Locality A8, Figure 39.

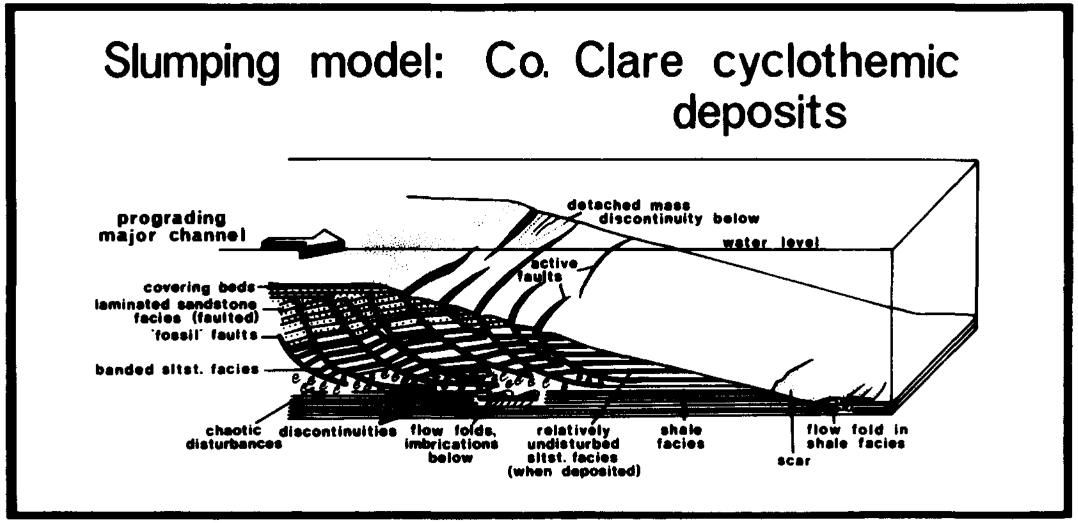

Figure 47. Slumping model Co. Clare cyclothemic deposits. (After Rider, 1969).

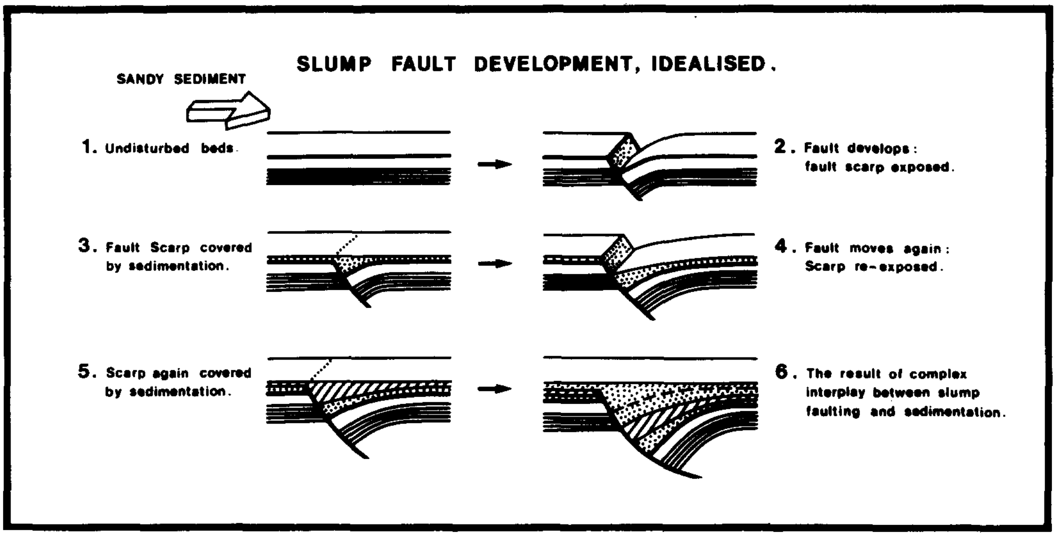

Figure 48. Slump fault development idealised. (After Rider, 1969).

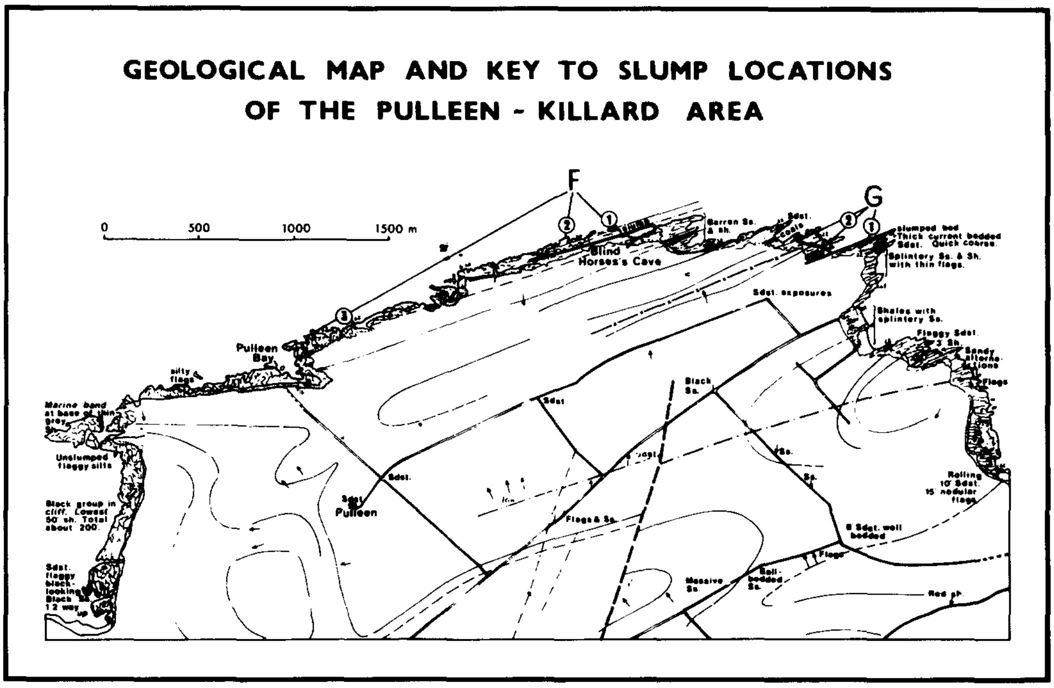

Figure 49. Geological map and key to slump locations of the Pulleen-Killard area. Localities F and G, Figure 1.

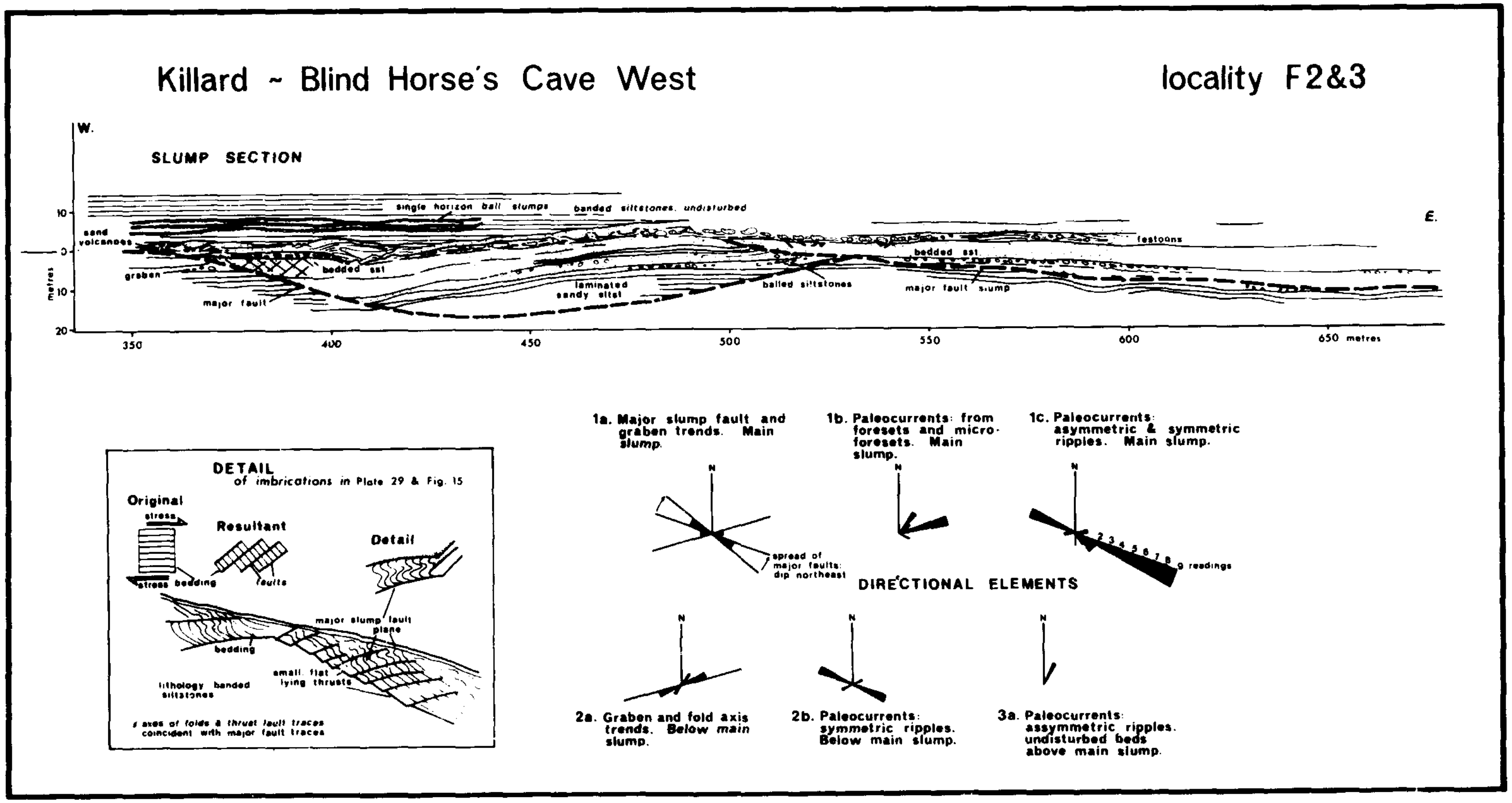

Figure 50. Killard: Blind Horse's Cave West. (After Rider, 1969). Localities F2 and F3 on Figure 49.

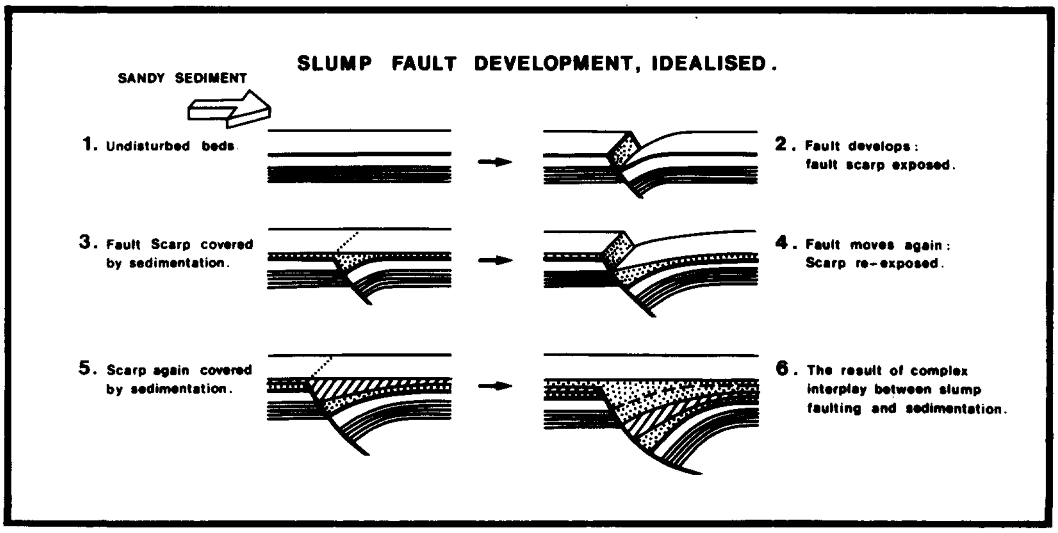

Figure 48. Slump fault development idealised. (After Rider, 1969).

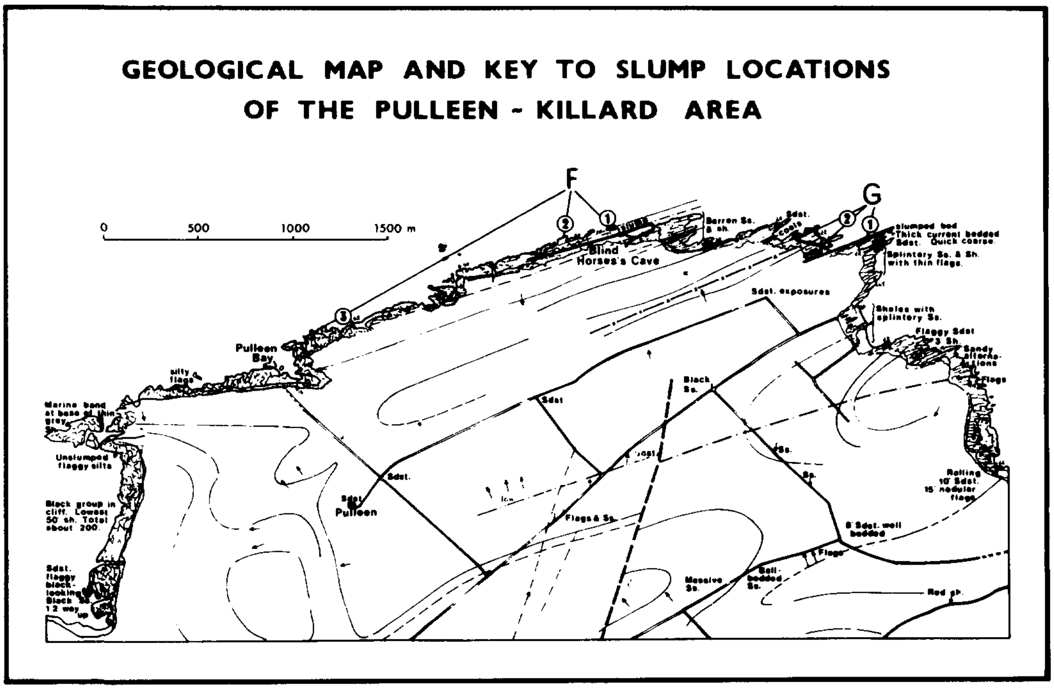

Figure 49. Geological map and key to slump locations of the Pulleen-Killard area. Localities F and G, Figure 1.

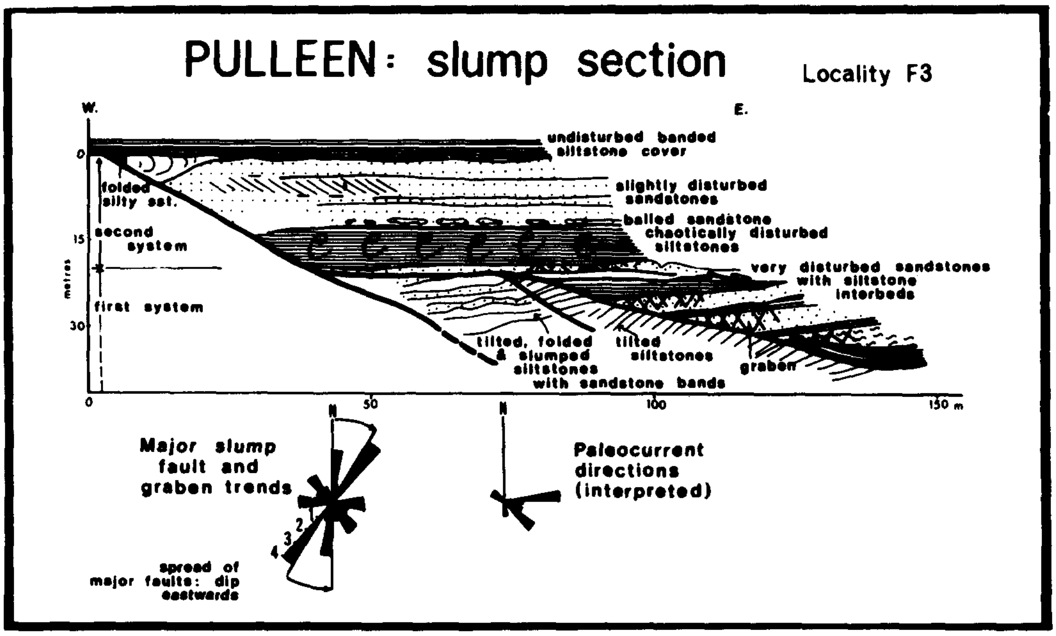

Figure 51. Pulleen Slump Section. (After Rider, 1969). Locality F3, Figure 49.

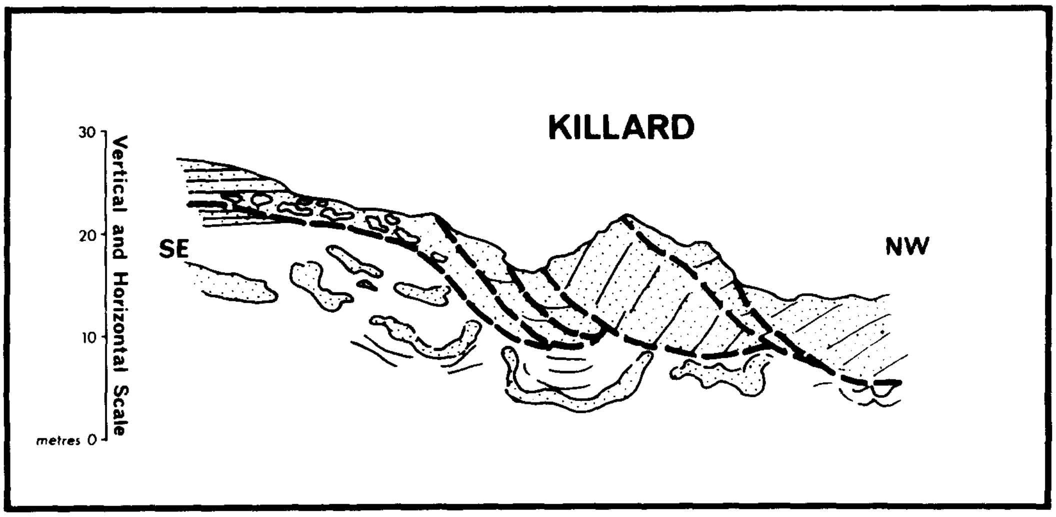

Figure 52. Killard Section. Locality G2, Figure 49.

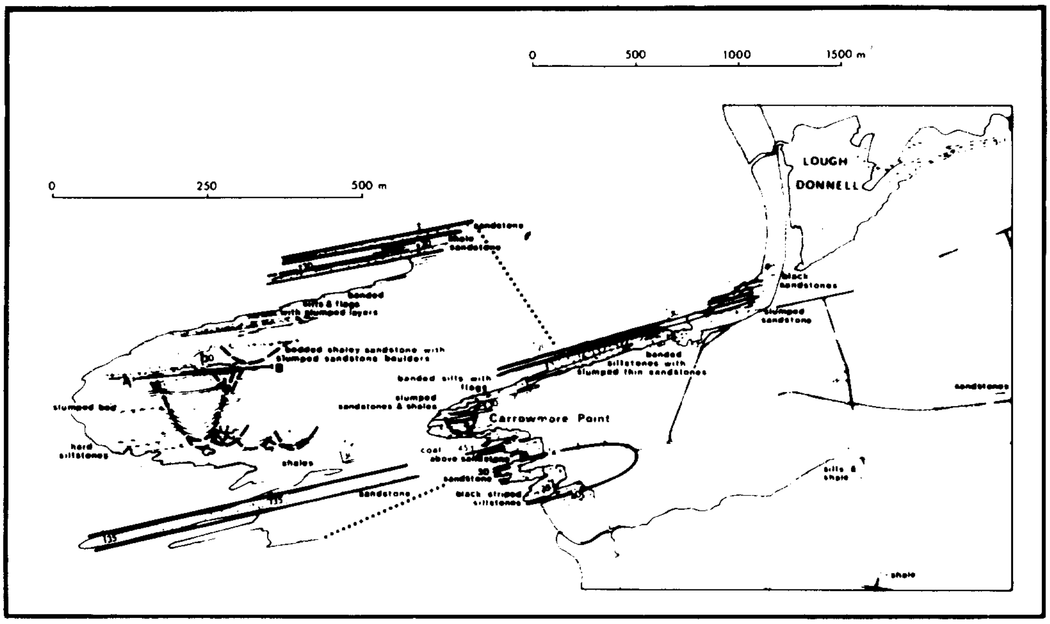

Figure 53. Carrowmore Point. Locality H, Figure 1.

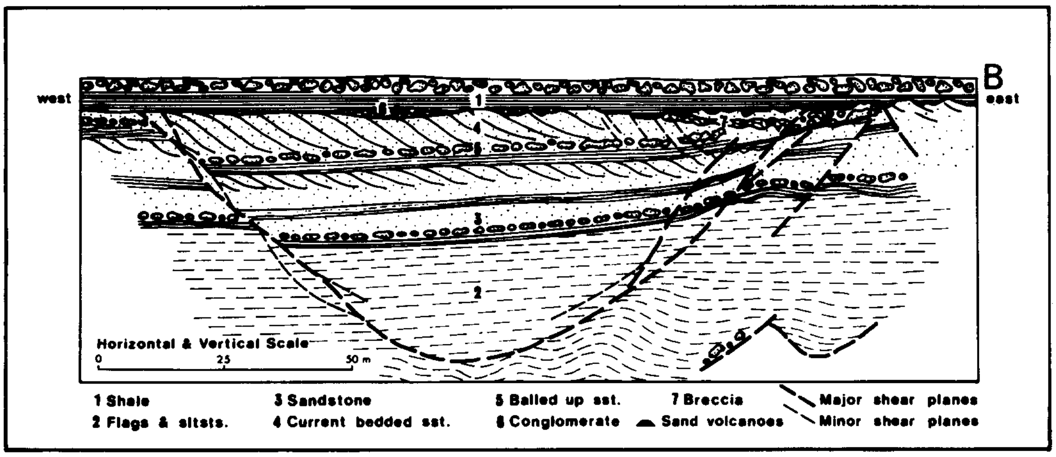

Figure 54. Cross-section of Carrowmore Point. Locality H, Figure 1. Line of section at west of Figure 53.

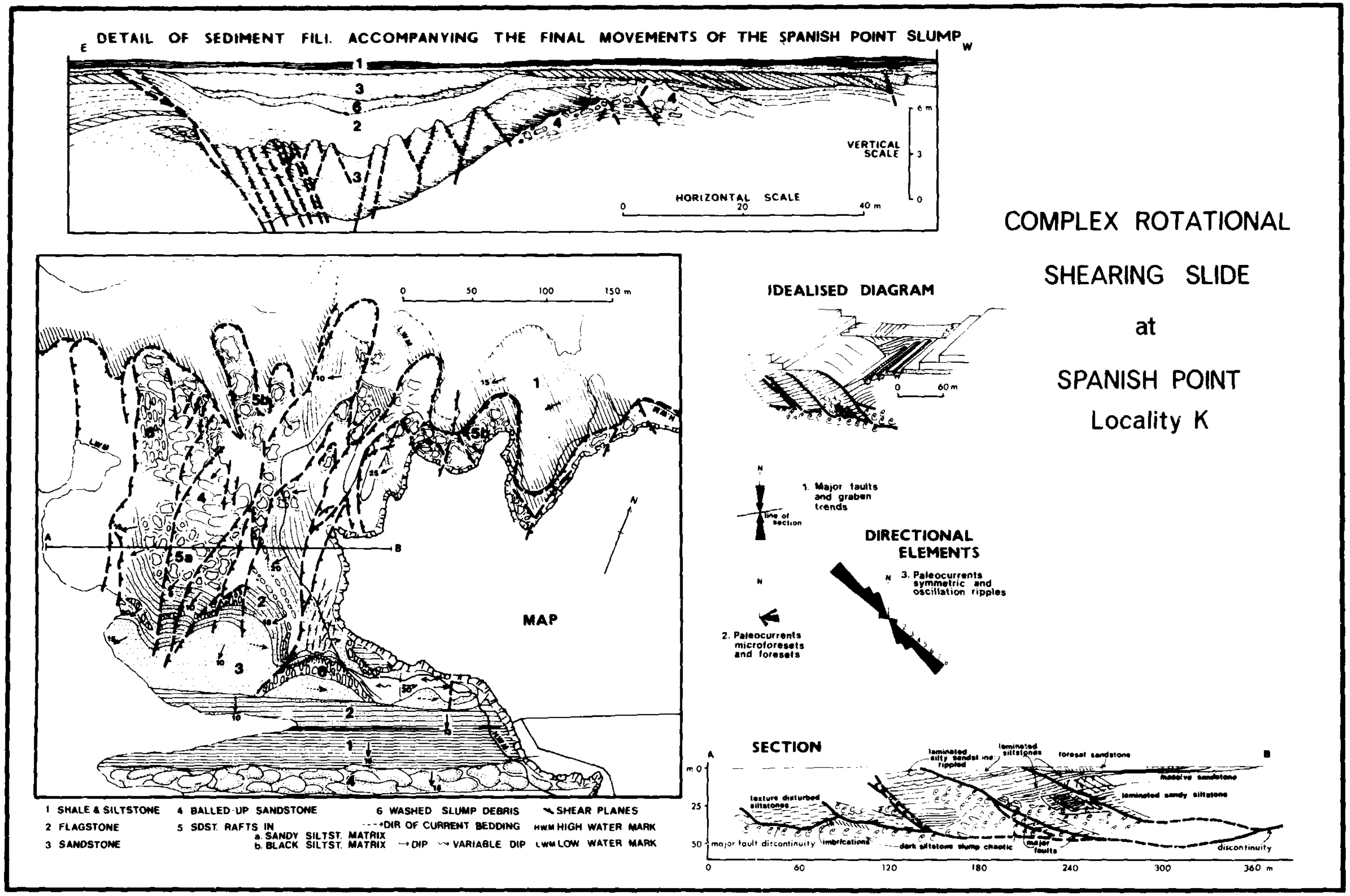

Figure 55. Complex rotational Shearing Slide at Spanish Point. (After Rider, 1969). Locality K, Figure 1.

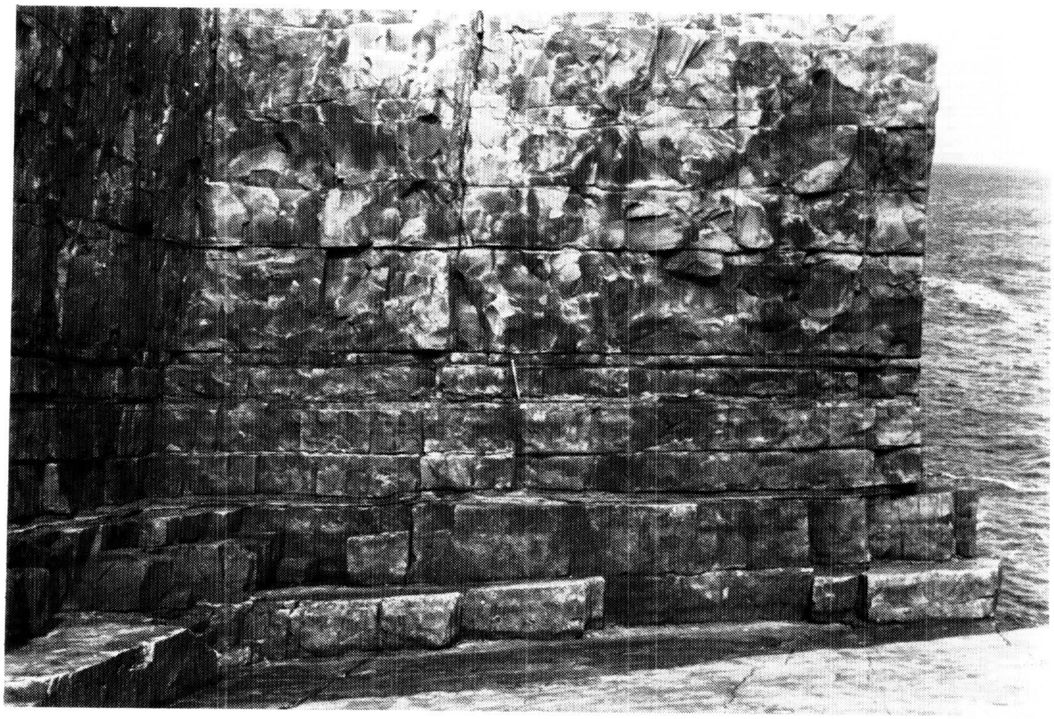

Plate 1. Ross Sandstone Formation, Locality A, Figure 1. Section approximately 5 m thick.

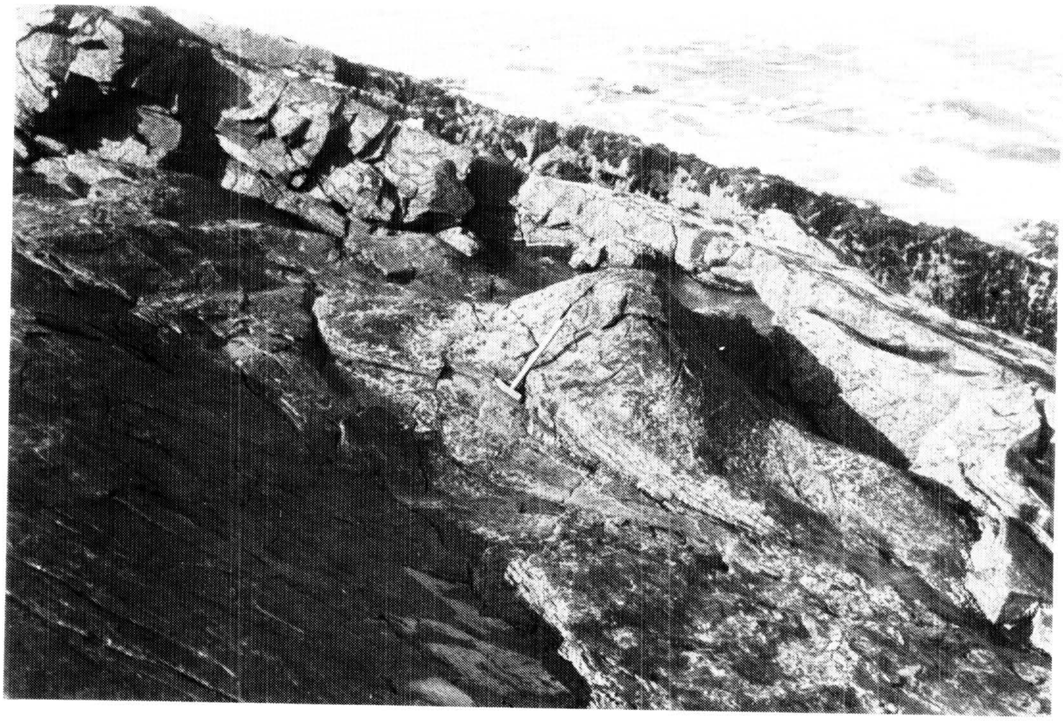

Plate 2. 'Point up' and 'cusp' structures in the Tullig Sandstone Member at Killard. Locality G, Figure 1. About one metre of section visible.

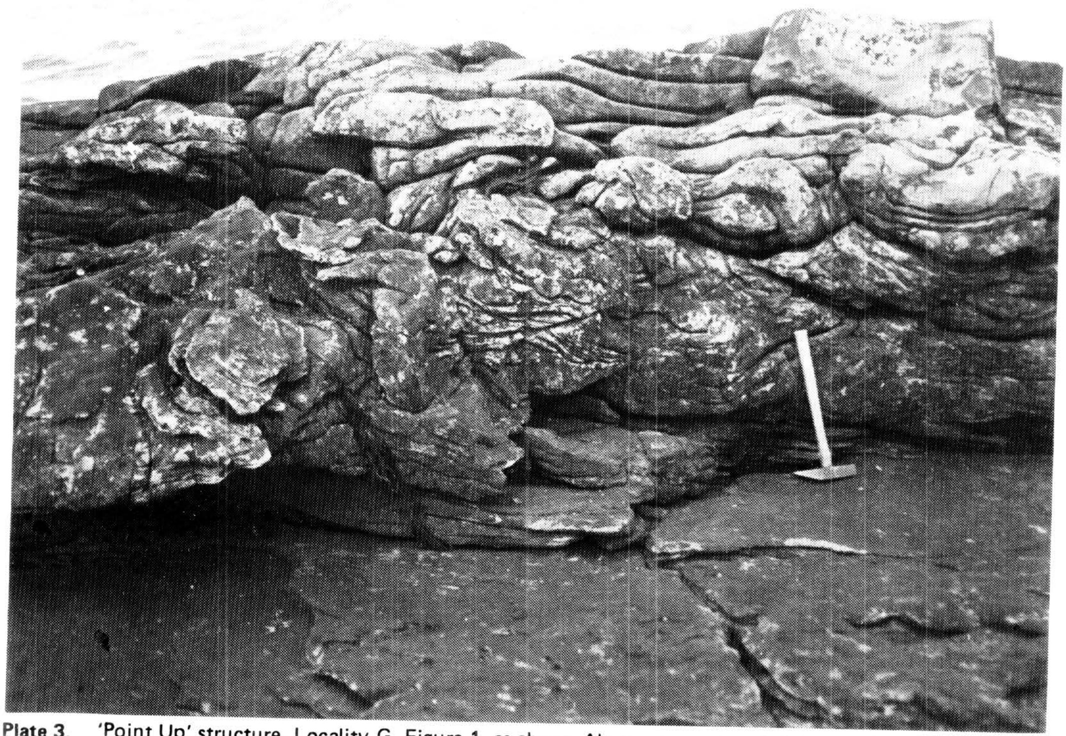

Plate 3. 'Point Up' structure, Locality G, Figure 1, as above. About one metre of visible section.

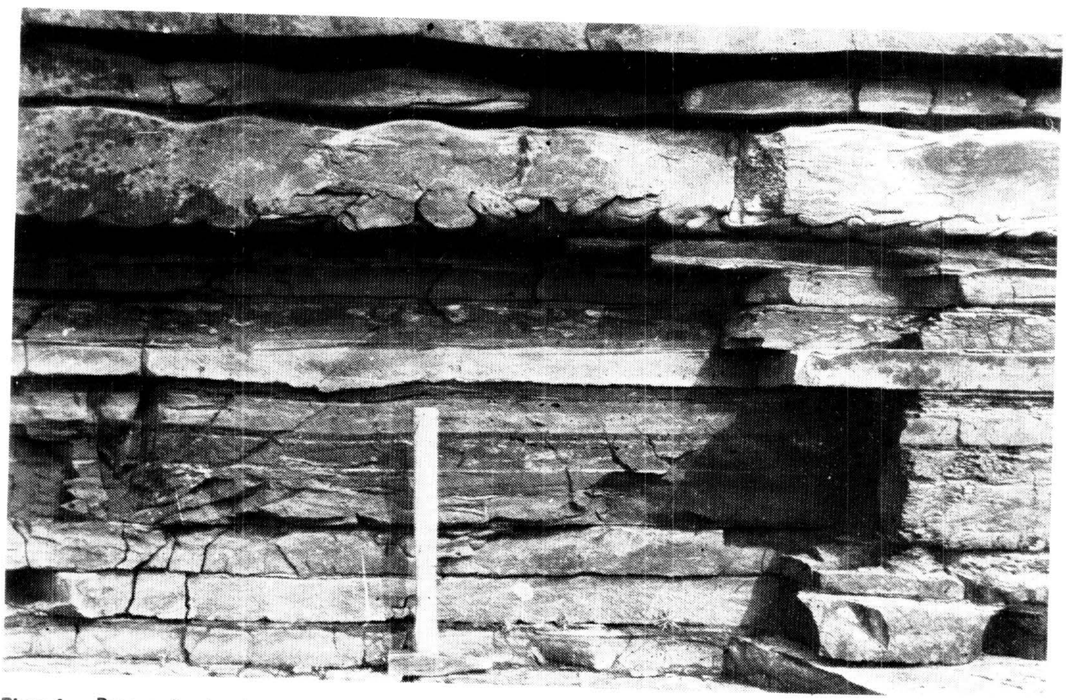

Plate 4. Progressive load casting in cyclothemic rocks on the face of the Black Rock, Mal Bay. Locality K, Figure 1.

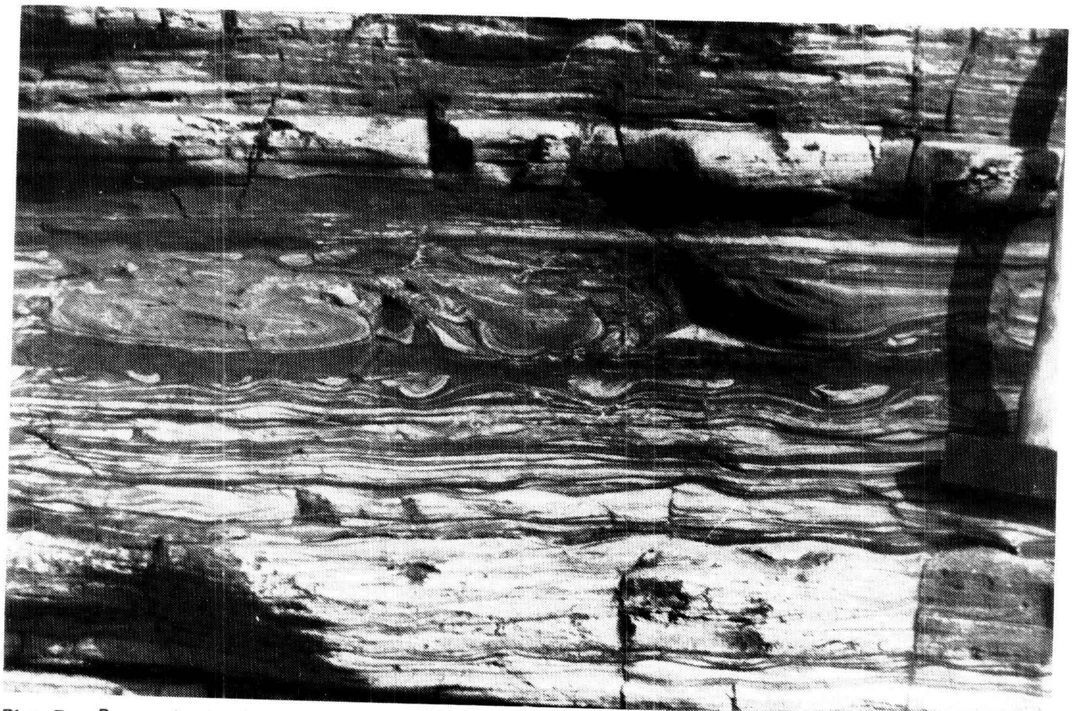

Plate 5. Progressive load casting in cyclothemic rocks on the face of the Black Rock, Mal Bay. Locality K, Figure 1.

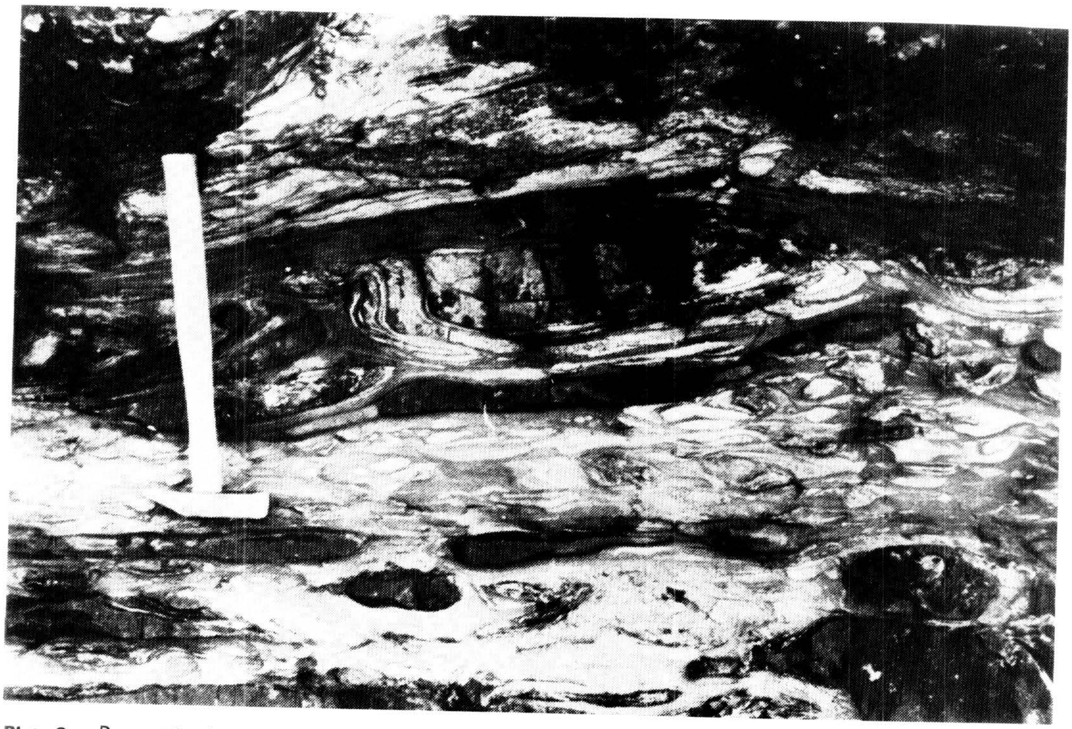

Plate 6. Progressive load casting in cyclothemic rocks on the face of the Black Rock, Mal Bay. Locality K. Figure 1.

Plate 7. Ball and pillow structure.

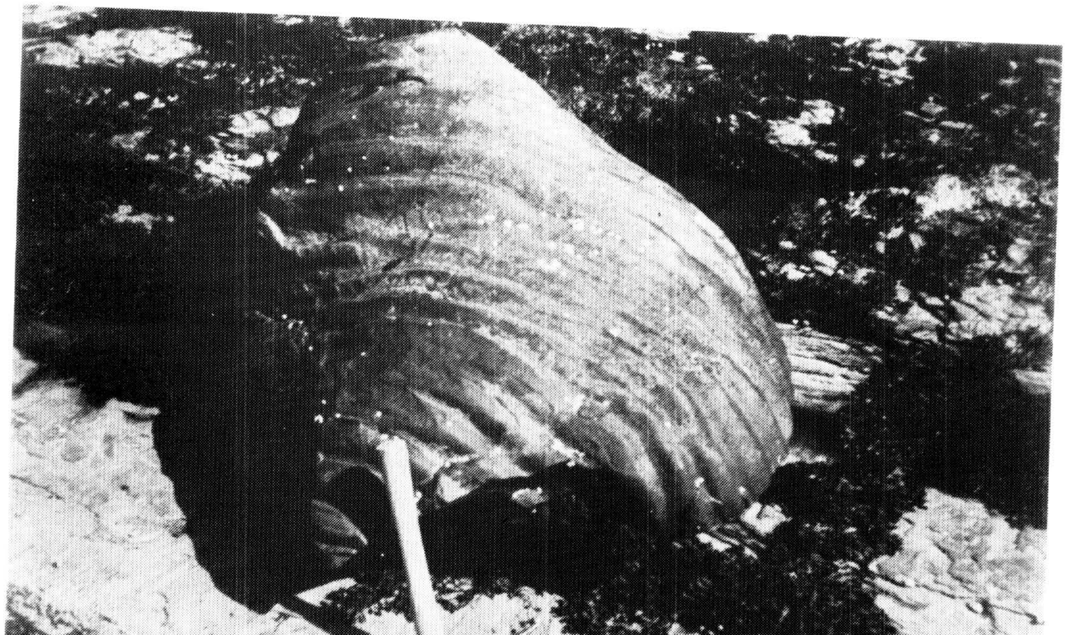

Plate 8. Pillow from foreshore at Quilty, opposite church. Locality I, Figure 1. See also Figure 14.

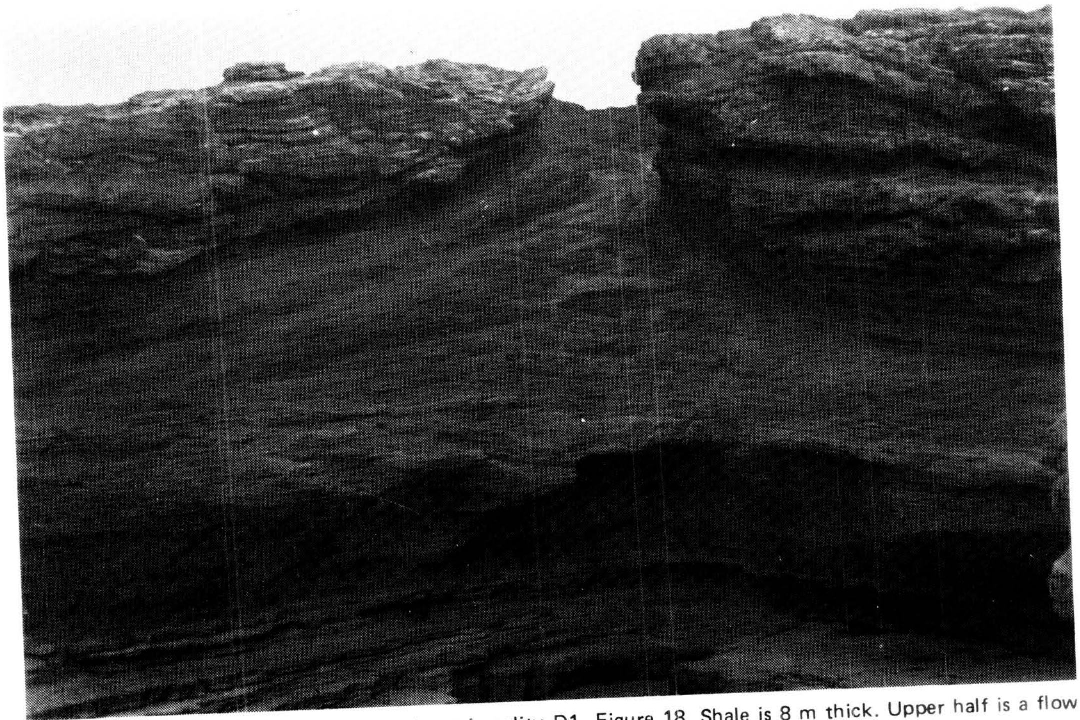

Plate 9. Shale diapir, with adjacent rafts at Locality D1, Figure 18. Shale is 8 m thick. Upper half is a flow breccia. Moore Bay, Kilkee.

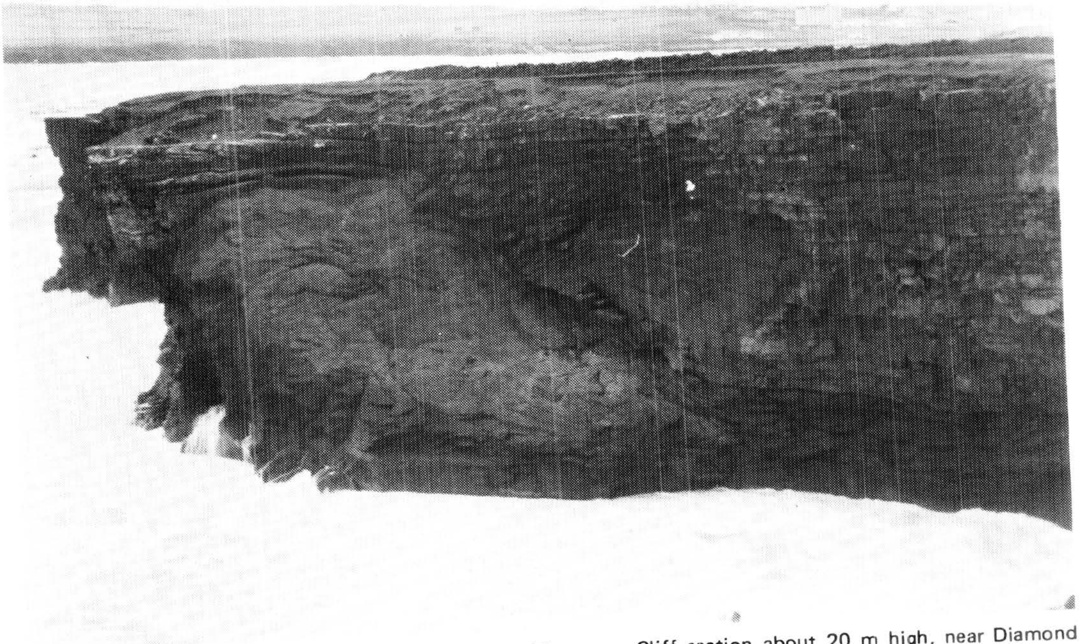

Plate 10. Shale diapir with undisturbed unconformable cover. Cliff section about 20 m high, near Diamond Rocks. Locality C3, Figure 18. Moore Bay, Kilkee.

Plate 11. Flow breccia in silty shale. Killard, Locality G2, Figure 52. Scale from match box.

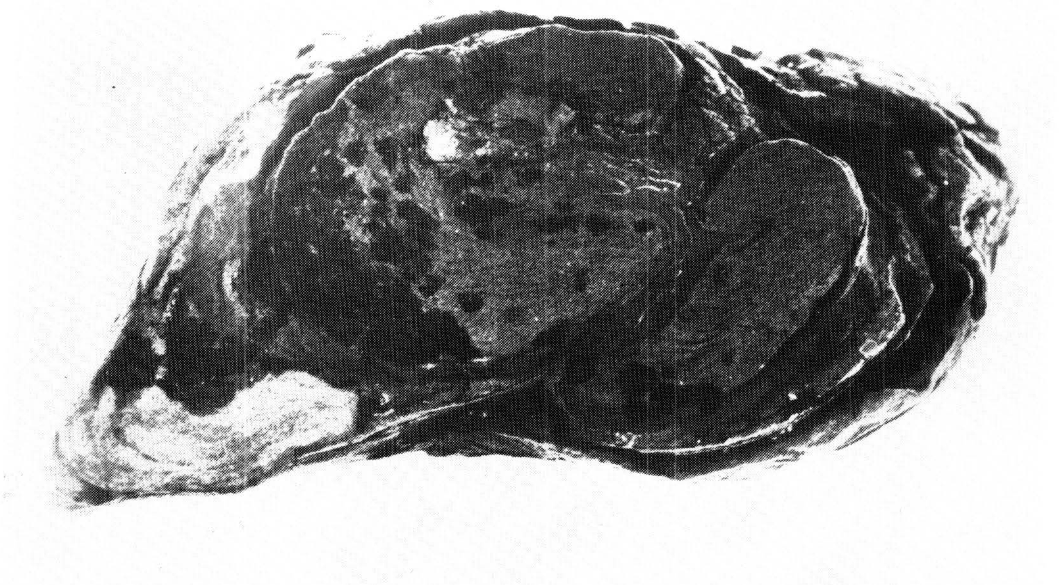

Plate 12. Section of sandstone ball derived from refolded cylindrical folds. Locality D2, Figure 18. Moore Bay, Kilkee, Specimen about 30 cm in longest direction.

Plate 13. Six-inch (15 cm) diameter core from Namurian beds showing fold-style of low viscosity sediment and nappe-like sheets.

Plate 14. Bed of sandstone balls below sand volcano sheet, Freagh Point. (Gill and Kuenen, 1958. Figure 2). Bed is about 1 m thick (see Figure 21).

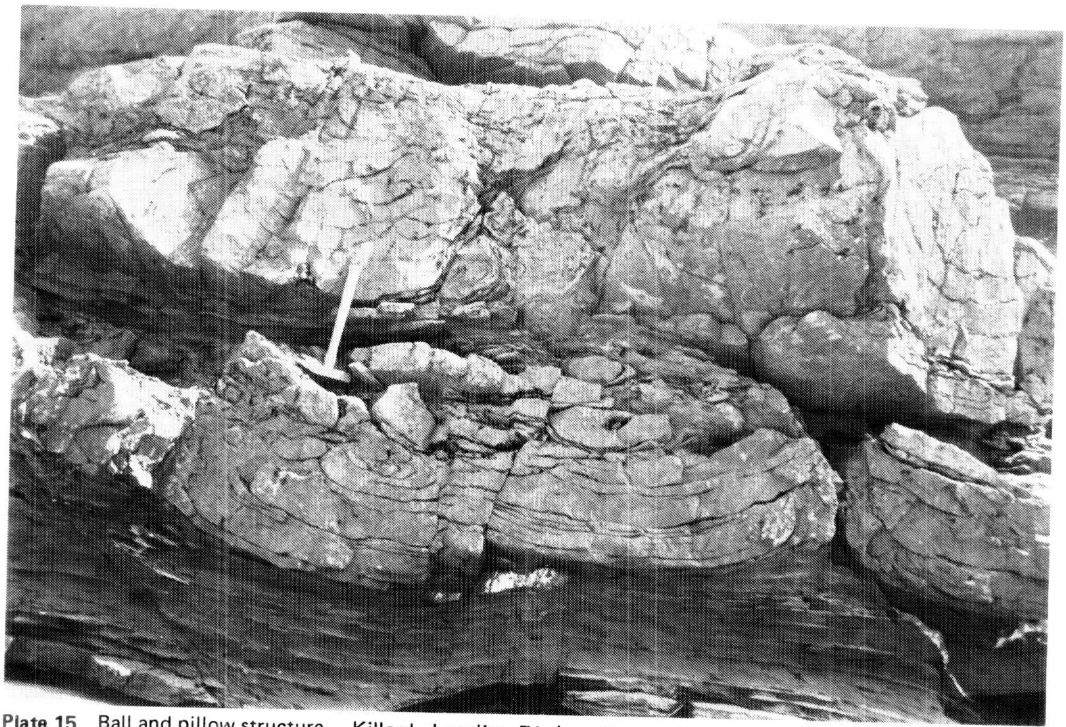

Plate 15. Ball and pillow structure. Killard, Locality F1 (see Figures 15, beds 7 & 8 and Figure 49). Section about 2 m thick.

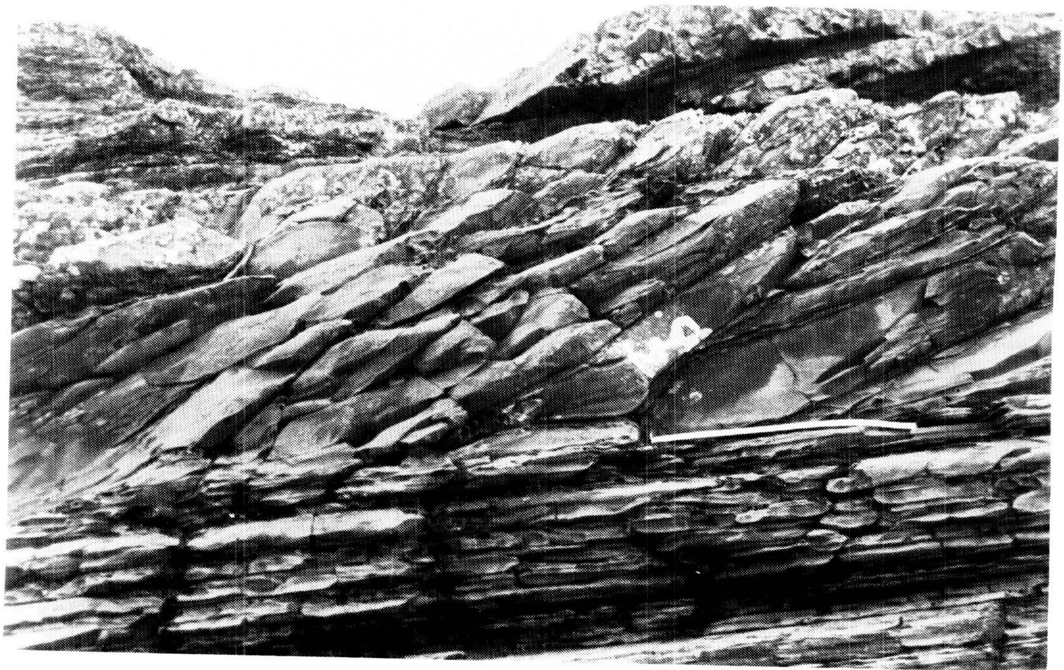

Plate 16. Chaotic slab-sheet, about one metre thick. Locality B, Figure 1, Goleen (Coast Road, south of Kilkee). See Figures 25 and 26.

Plate 17. Rotational shear cutting a channel-like body of sandstone balls about 5 m thick. Foreshore at Quilty (opposite Church) Locality I, Figure 1. See also Figure 14.

Plate 18. Chaotic slab sheet at the Atlantic Slump (Figures 16 and 17). Bed of silty flags broken up into polygonal or oval slabs separated by partially re-slurried material. Atlantic Lodge, Mal Bay, Locality K south, Figure 1.

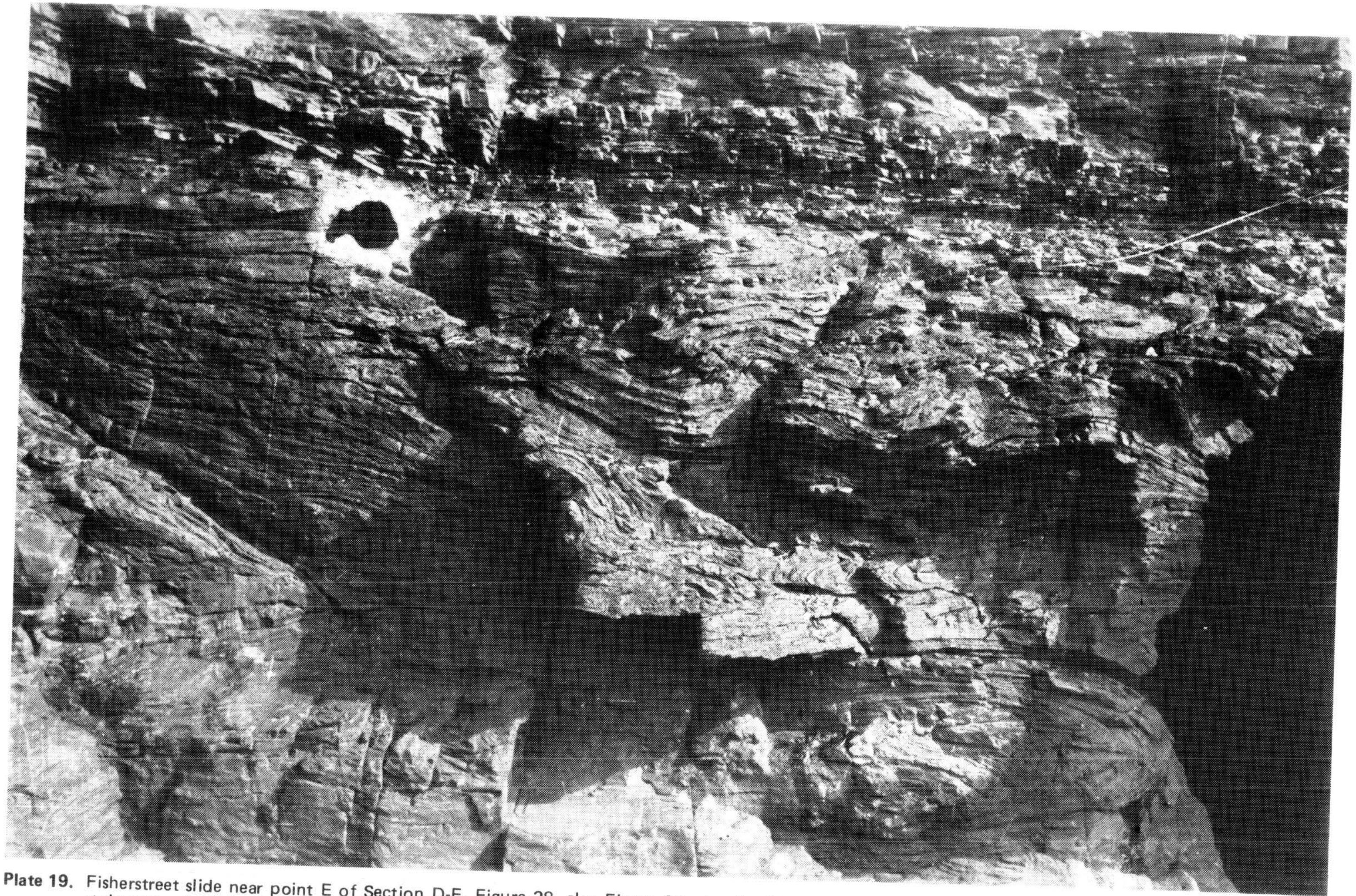

Plate 19. Fisherstreet slide near point E of Section D-E, Figure 28, also Figure 27, showing highly fluidal fold style and refolded folds (right lower part of picture). Also note erosional top and unconformable cover. Locality M, Figure 1, Fisherstreet Bay.

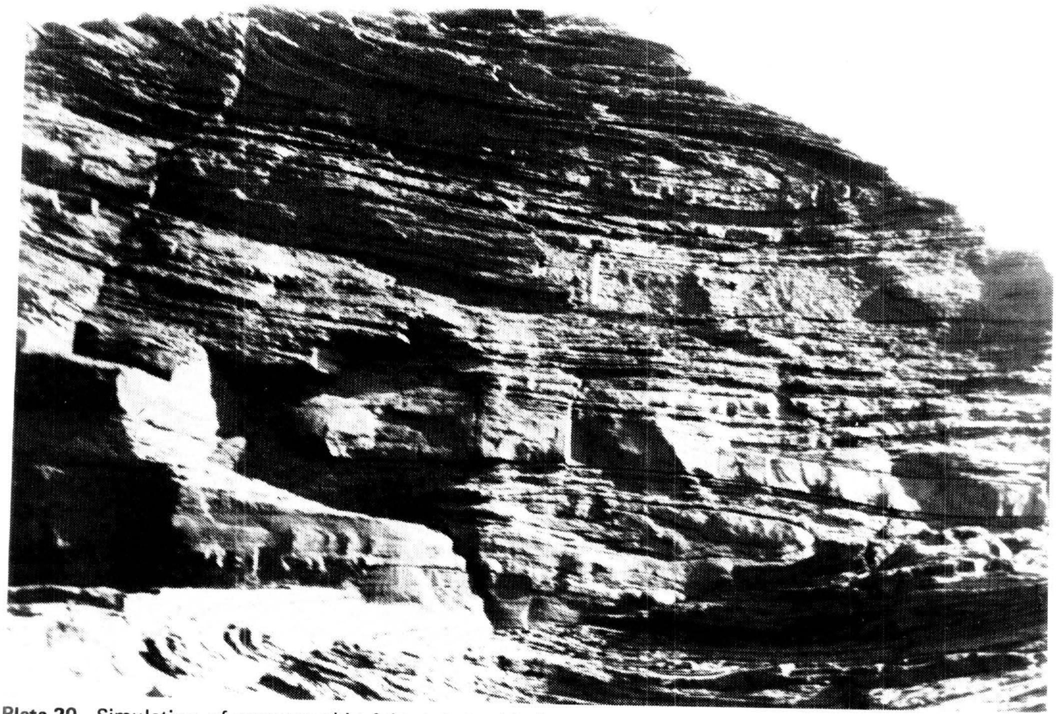

Plate 20. Simulation of metamorphic fold-style by black sandy siltstones and shales of the Fisherstreet Slide. Folds display a high degree of flank thinning and axial thickening, and occasionally axial plane cleavage. About the centre of Section D-E, Figure 28. Section visible about 5 m thick. Locality M, Figure 1, Fisherstreet Bay.

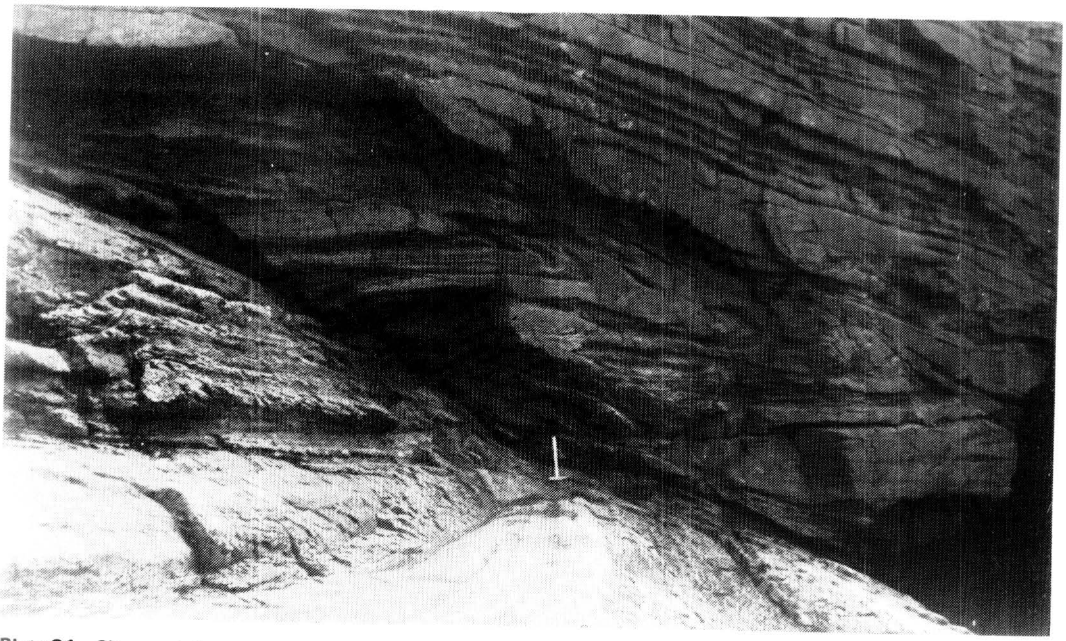

Plate 21. Clean-cut thrust at northern boundary of fold and thrust complex figured in Section A-B, Figure 29, Fisherstreet Slide. Near point A of the section. See also Figure 30. Hammer is 30 cm long. Locality M, Figure 1, Fisherstreet Bay.

Plate 22. Fold closure against northerly-dipping thrust near the southern margin of complex, section A-B. The fold axis is at an angle of about 30° to the trace of the thrust plane. Near point B, Figures 29 and 30. Fisherstreet Bay, Locality M, Figure 1.

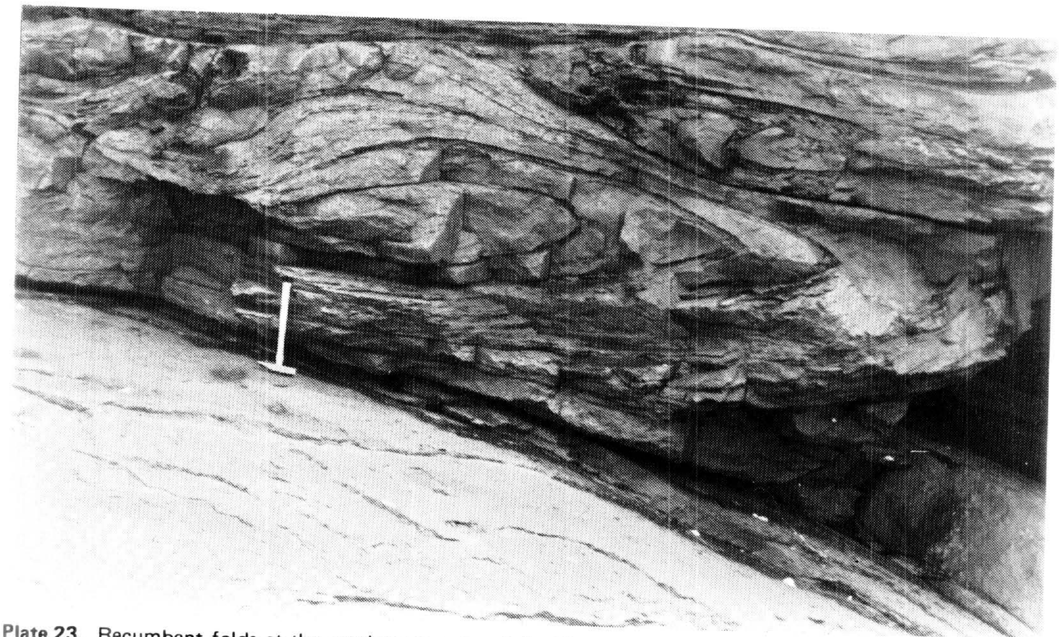

Plate 23. Recumbent folds at the southern margin of the Fisherstreet Slide. These form a separate upper sheet of structures deduced to have been formed by southwards gliding from upthrust masses further north. Near southern margin of section A-B, Figures 29 and 30. Visible section about 10 m thick. Fisherstreet Bay, Locality M, Figure 1.

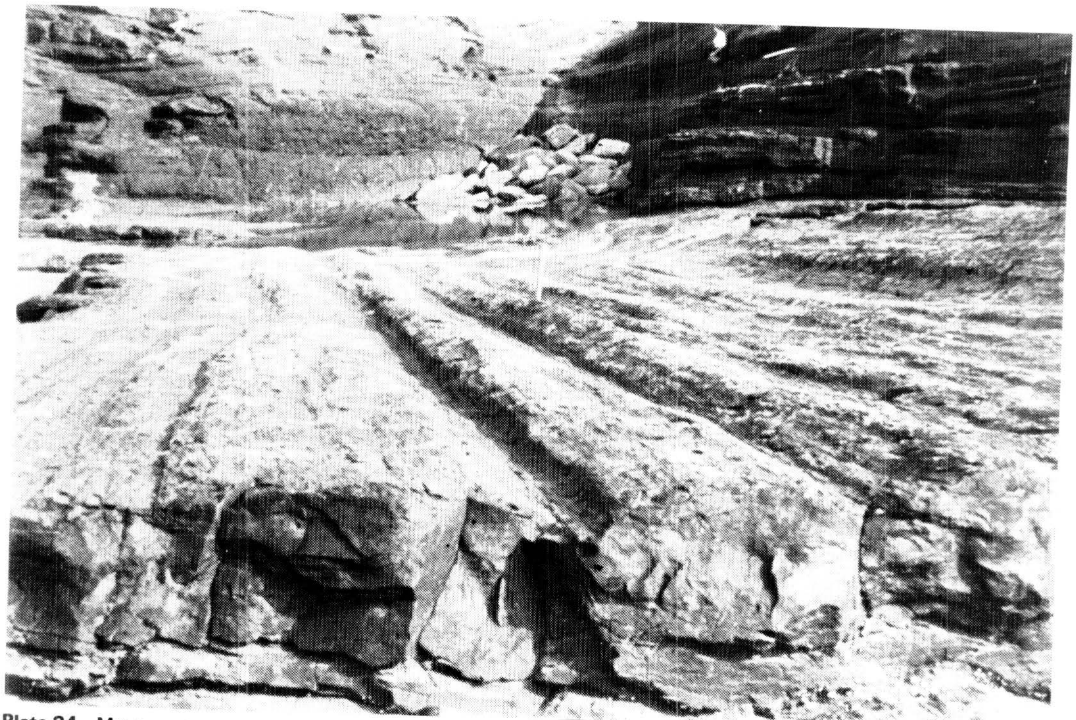

Plate 24. Movement grooves in flat-lying beds of the Fisherstreet Slide. The grooved bed, about 40 cm thick, is not folded and the base unaffected. See Figure 30 for groove distribution. Fisherstreet Bay, locality M, Figure 1.

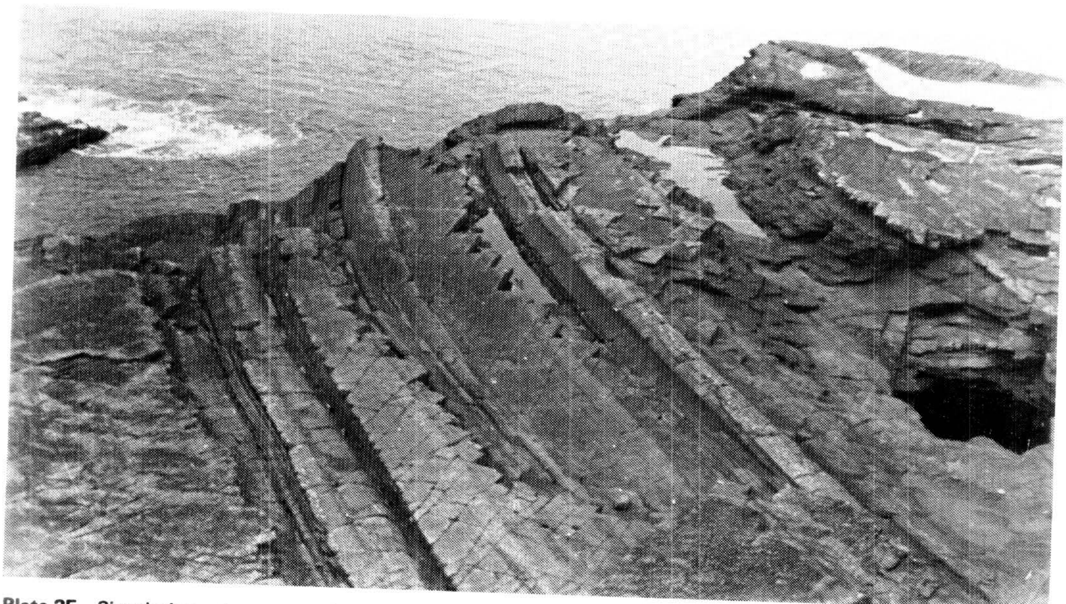

Plate 25. Simulation of metamorphic-style fold mullions in the Fisherstreet Slide. See Figures 31, 32 and 33 for scale and internal form. Near the northern end of the Section A-B, Figure 29.

Plate 26. The Ross Slide. Terminal 'bulldozer'. The forward-moving slumped mass on the left has rucked up the bedded sandstones on the right (see Figure 44) giving clear kinematic parameters. Height of section above flat platform about 3 m. Locality A5 (Figure 39) Bridges of Ross.

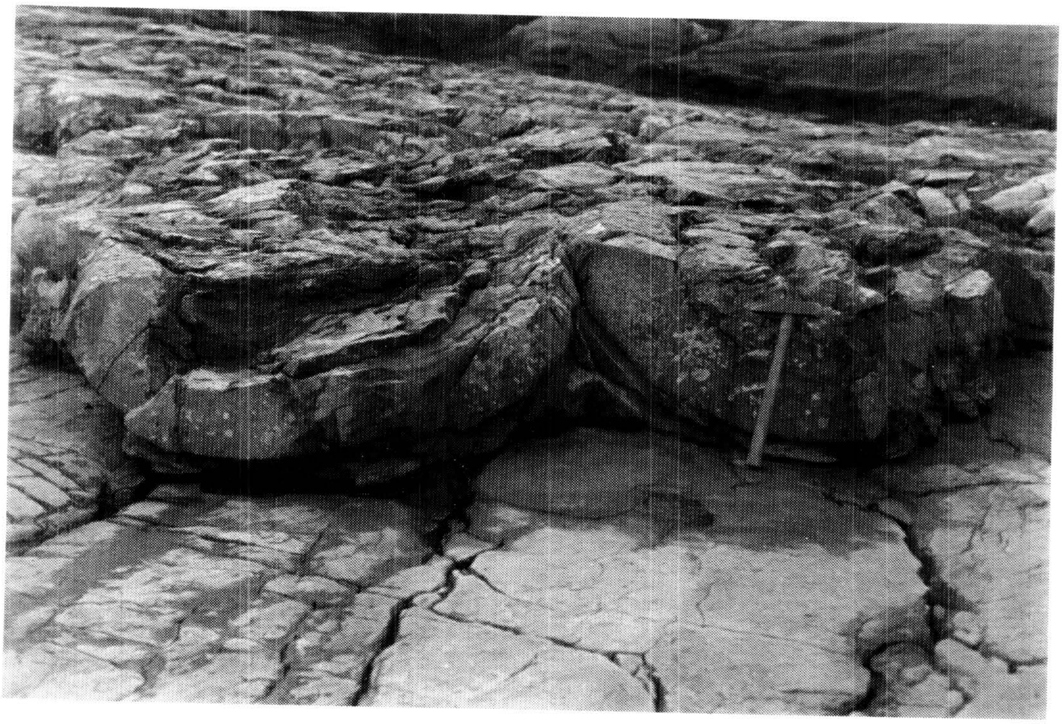

Plate 27. Basal rucked zone of Ross Slide showing fold core below imbrications of the same trend. Locality A4, Figure 39. Bridges of Ross, looking westwards.

Plate 28. Moulded fold core below imbric shear. Profile F-G (Figure 41) midway between F and G. Visible section about 6 m thick. Locality A6 (Figure 39) Bridges of Ross.

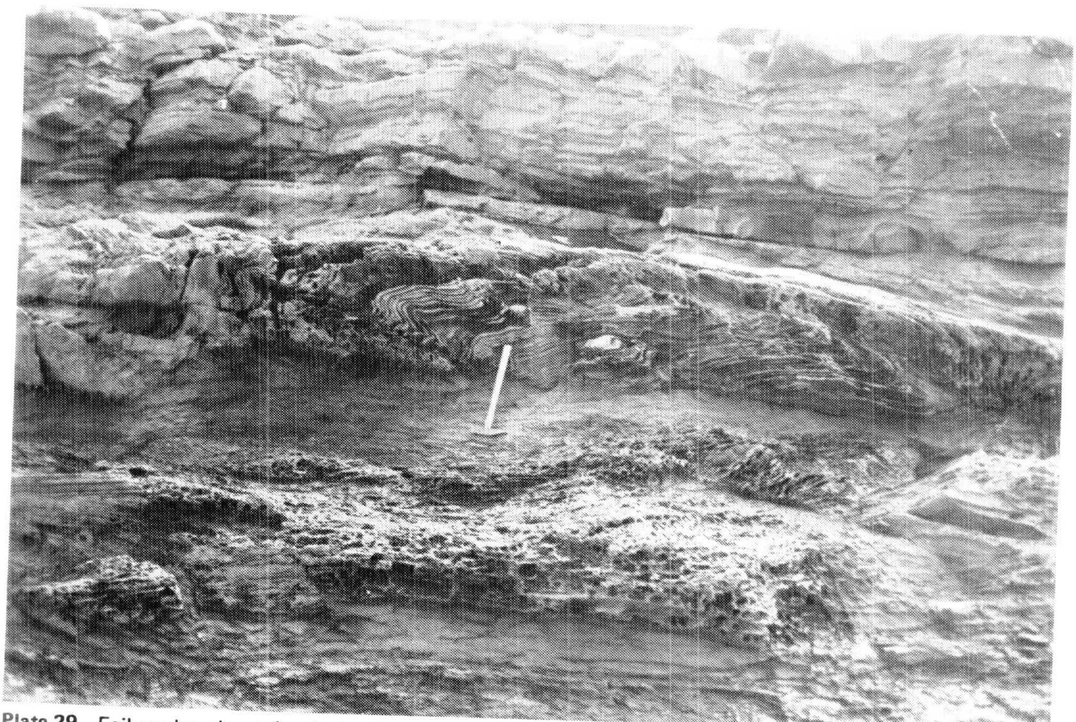

Plate 29. Failure by shearing of the footwall of a rotational shearing slide at Blind Horse's Cave, Killard. Locality F1 (Figure 49).

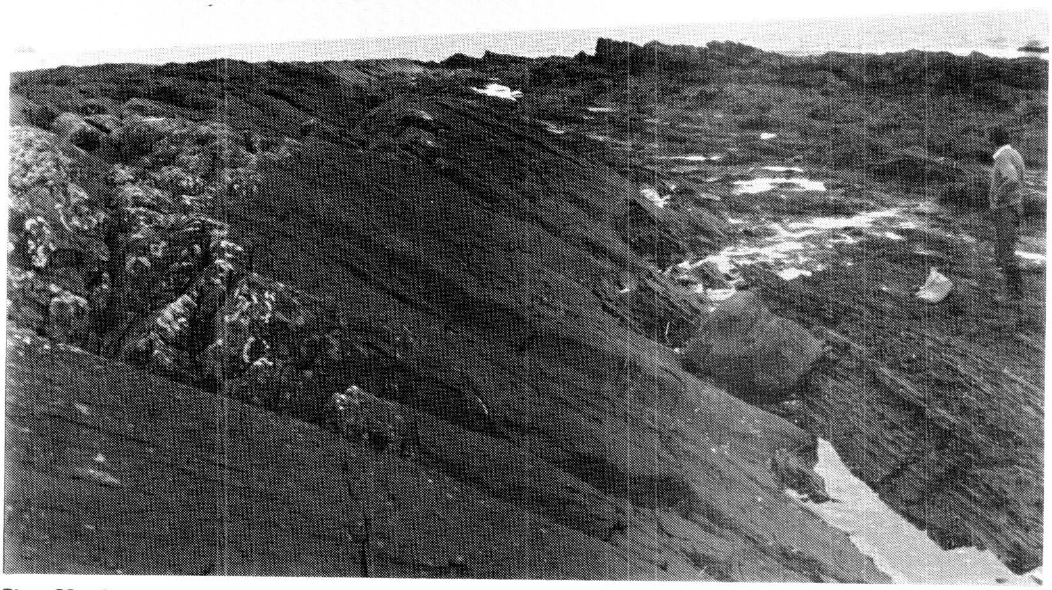

Plate 30. Graben and horst distension structures at the top of the Carrowmore Point Slump. Section, Figure 54 near locality 7. Overlying infilling silty shale clearly not affected. Locality H, Figure 1.

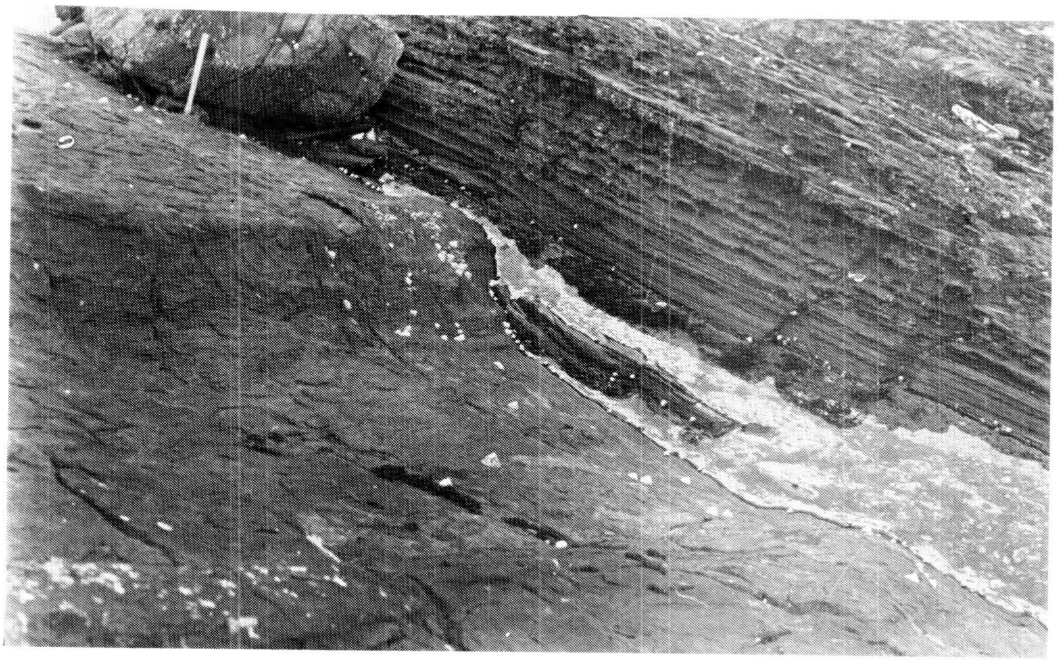

Plate 31. Silty shale infilling graben at the top of the Carrowmore Point Slump. Hammer is 35 cms long. Locality H, Figure 1.