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Conference Report

Deformation of soft sediments

W. R. Fitches & A. J. Maltman

In recent years there has been increasing awareness of the variety of structures formed by the deformation of unlithified sediment and of the difficulty in distinguishing these from structures produced in rock. Confusion of the two types clearly has serious implications, for example, in orogenic strain sequences where it is necessary to separate pretectonic slump structures from the results of later tectonism. Unfortunately, the features of soft sediment tend to fall within the realm of sedimentology whilst those of deformed rocks are the concern of structural geology so that the region where depositional, compactional, diagenetic and tectonic processes merge tends to be viewed from divergent positions and perhaps receives relatively little attention.

The meeting held at Aberystwyth provided a *forum* for workers of differing background to assess recent progress in clarifying this difficult area. About a hundred participants heard a day of talks which made it clear that a wide diversity of phenomena on a variety of scales falls within the conference topic.

The first part of the meeting was concerned with the structures and processes which are largely confined to soft sediments, beginning with large scale features such as olistostromes and moving on to the smaller structures such as convolute laminations. The later part continued through the compaction-lithification sequence, and the effects of tectonic strains became increasingly emphasised.

Discussion following the talks returned repeatedly to two points. Firstly, there was the problem of *defining terms* and especially how to define the onset of lithification. The second recurrent topic was the apparent paucity of reliable *criteria* for distinguishing between penecontemporaneous and post-lithification structures. Observations which seemed potentially useful for this purpose were scrutinised and many were rejected. For example, a series of folded layers truncated above and below by planar beds was not considered by some to be diagnostic of pre-lithification movement in the absence of, say, undeformed worm burrows crossing the junctions.

It seems that generally the geologist has to assess the cumulative weight of several not infallible criteria in order to identify those structures which were formed by the deformation of soft sediment.

Report of Tectonic Studies Group Meeting held at the University College of Wales, Aberystwyth, on 5th and 6th March, 1977.

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Soft-sediment deformation: a review. N. H. Woodcock

Soft-sediment deformation was first demonstrated by mid-nineteenth century observations of glaciogenic structures and clastic dykes, but it was only with the descriptions by Milne and Heim of recent 'sub-aqueous landslides', at about the turn of the century, that the concept was applied to other structures. This proven instability of some sediments on low slopes resulted in many 'intrastratal contortions' and some conglomerates being explained by a 'slumping' mechanism. By the 1930's several experimental studies of soft-sediment deformation had been published, as had the first theoretical analyses. In the later 1930's the controversy between Boswell and Jones over Silurian structures in Wales had a large audience, and focused attention on criteria for distinguishing soft-sediment from 'tectonic' structures.

In the post-war period, three new approaches have contributed to knowledge of soft-sediment deformation. The turbidite concept has shown the distinction between slump structures and convolute lamination, and has suggested that slumping and turbidity flow may be linked by a spectrum of sediment gravity flow mechanisms. The soil mechanics approach has solved some of the theoretical problems of soft-sediment behaviour. The growth of submarine profiling has revealed many recent examples of sediment deformation by slumping, diapirism and subduction.

The many unsolved problems in this field are of two types. Firstly, there are the difficulties of describing, classifying, and interpreting accepted soft-sediment structures. For example, the terms 'glaciogenic structures' and 'convolute lamination' each encompass a range of geometrically and genetically differing structures, and need clarification. Also unclear is the relationship between slumping and the other mechanisms of sediment gravity flow.

The second type of problem is the recurring one of distinguishing 'tectonic' from 'non-tectonic' structures. There is some semantic confusion here, particularly over the meaning of the term 'tectonic'. However, the discrimination of soft-sediment structures remains a meaningful and often vital exercise, although complicated by the lack of widely applicable criteria. The geometric distinction of soft-sediment folds, fabrics and microstructures from their 'tectonic' counterparts is one aspect of this problem. The distinction of tectonic mélanges from olistostrome mélanges is another, and one where the importance of 'sedimentary' processes is now being realised. Assessment of the general importance of soft-sediment deformation must await further work. Those who doubt its potential significance should be cautioned by the fact that most 'slumps' on modern continental margins are of a scale which would probably make them 'tectonic' if preserved in the geological record.

Deformation in sediments. M. D. Max

Describing the range of sedimentary deformational features and their relationship to tectonism using the conventional episodic structural nomenclature is inaccurate as the processes are inherently semi-continuous and locally controlled. Therefore, a general system of orders of sedimentary deformation is proposed (Table 1) with the transitional term tectono-sedimentary introduced. In the lower orders compaction predominates, whereas higher orders reflect increasingly dynamic conditions. Some individual features such as loading, slumping, folding and the imposition of fissility are found in all orders but differ in degree and profusion. Olistostromes are confined to the higher orders and nappes to the tectono-sedimentary and tectonic fields.

Sedimentary deformation comprises a structure of *adaptation* mainly by local control and tectonic deformation comprises secondary or newly formed structures of *imposition.* Solution effects related to compaction, such as stylolites, are held to be structures of local adaptation related to diagenesis and are regarded as sedimentary deformation. Localized tectonic structures such as cleavage may occur well down into the field of sedimentary deformation with which structures they may be contemporaneous. There is no precise set of criteria that everywhere can be used to delineate a time of abrupt passage from sedimentary to tectonic deformation.

Sedimentary deformation has gravity as the driving mechanism, seismic shocks and the degree of slope as a control and the sediments themselves exert a variable and significant influence on their final appearance. In sedimentary deformation, shape fabrics are irregular, folds are characteristically triclinic and changes in symmetry are usually not meaningful in respect of a regional post-sedimentary tectonic pattern. The internal microscopic deformational history is poorly represented and is anomalous in respect of the macroscopic appearance. Most important, a larger scale stratiform arrangement is preserved; undisturbed biological activity in deformed beds is the only unequivocal criterion for determining deformation entirely of a sedimentological aspect.

Slope morphology and sediment deformation, Santa Monica Basin, Southern California.
Barbara E. Haner

The continental borderland of Southern California illustrates the inter-relationship of tectonics and sedimentary environments in creating (a) zones of slope stability, (b) progressive slumping on gentle slopes where sediments fail at canyon axes and (c) large scale rotational slumping associated with fault scarps and rapid sediment influx. Syndepositional tectonics may also affect unlithified slump toes which terminate close to basal slope faults.

Transform NW–SE faulting has created a continental margin characterised by *en echelon* deep water basins separated by islands and banks. The eastern slope of the nearshore Santa Monica Basin is controlled by faulting parallel to this regional trend but the northern slope of the basin is controlled by a major E–W fault associated with frequent earthquake epicentres. A subparallel minor fault 6 km south passes through the upper axis of Santa Monica Canyon. This latter fault system enclosed a 20 km infilled E–W graben extension of Santa Monica Basin bounded on the north by a steep fault scarp and a canyon to the south. The trough extends from shelf break (70 m) to 800 m with average westward gradient of 1°15′.

High resolution seismic profiling revealed three contrasting areas of slope morphology. In the south, NW-SE slopes average 8° and may exceed 15°. Sedimentation is slow as the canyon head traps sediment, resulting in slow compaction and slope stability. Only adjacent to the canyon has minor slumping occurred.

The eastern back wall of the trench is characterised by 1°8′ slopes decreasing to less than 1° between 150–250 m. Here, high sediment input by the Los Angeles River occurred during periods of low sea level. Seismic reflectors are indistinct and discontinuous while surface topography is stepped. Downslope this zone terminates either with tensional notches or rotational slumps into the canyon. This indicates semi-continuous sediment deformation caused by slope creep and downslope slumping following rapid sediment influxes and sediment failure.

Westwards, along the northern boundary fault, gradient and slope length increase to 7° and 5 km. Sediment is thin or absent on the upper fault scarp, while downslope arcuate slump glide planes develop. Slumped material covers 110 km^2 and averages 9 m in thickness. Slump toes are

Table 1

DEFORMATION TYPE	CHARACTER	DEFORMATIONAL COMPONENT	JUNCTIONS
TECTONIC	DISRUPTION → REDISTRIBUTION	SECONDARY OR IMPOSED STRUCTURES CONTROLLED BY IMPOSED STRESS	TECTONIC DISLOCATIONS
TECTONO-SEDIMENTARY		WHOLESALE REDISTRIBUTION	SEDIMENTARY PASSAGES (NORMAL ↔ ABNORMAL)
HIGH ORDER SEDIMENTARY		SEMI-CONTINUOUS RESEDIMENTATION	
		GROUPS OF SETS OF BEDS	
INTERMEDIATE ORDER SEDIMENTARY	LOCAL DISTURBANCES: UNRESTRICTED	SETS OF BEDS	
LOW ORDER SEDIMENTARY	LOCALIZED	INDIVIDUAL BEDS	
	RESTRICTED	BED SURFACE OR INTERFACE	

slightly contorted, terminating at a WNW trending ridge near a basin and graben junction. Gravity anomalies indicate that this ridge may be either a large slump toe or a rising anticline, causing disturbance of unlithified sediments. During active uplift and low sea level, short, steep streams dumped material close to the shelf edge. Downslope slumping could be triggered by earthquakes, or rapid changes in overburden and sediment instability.

Slumping and debris-flow in the Palombino Limestone, Northern Appenines. M. A. Naylor

In the Trebbia Valley (Provincia di Genova) of the Northern Apennines, the eugeosynclinal sequences include a Cretaceous succession of 300 m of olistostrome overlain by 250 m of Palombino Limestone. The latter consists of alternating beds of turbiditic micritic silty limestone and shale, with thin sandstone bands. Beds range from 2 to 200 cm thick.

Soft sediment disruption of the Palombino sequence occurred by folding (of sandstones and limestones) and boudinage (of limestones). These processes are attributed to repeated gravitational slumping down a SW-facing palaeoslope. Evidence for the palaeoslope orientation is provided by consistently oriented asymmetric slump folds. The folds, often recumbent, may form trains with curved axes in the continuous sandstone beds, may be sheared out, or occur as isolated hinges in shale. Folding is intrafolial and disharmonic, even between two adjacent beds. Boudins show a range of shapes, reflecting deformation at different stages in the lithification of the limestone. Parallelism of boudin and fold axes is consistent with gravity sliding.

A continuum of processes is reflected in the lateral and vertical transition from olistostrome ('distal', resedimentation) to disrupted and undeformed Palombino ('less distal'; slumping and turbidite deposition) which later prograded on to the olistostrome. The olistostrome represents many thin debris-flows (0.3 to 6 m); in these, the limestone clasts were supported by the strength of the mud matrix. Channels, slump folds and other features of debris-flows confirm the palaeoslope orientation. Earlier models for the origin of olistostromes, involving shedding of debris from NE-advancing nappes, are inapplicable here.

Criteria for positive identification of soft-sediment deformation are few, but may include the following: folds and boudins lack geometrically related mineral veining; folds are associated with convolution and liquefaction of sand layers; folds may be load-casted after formation. Low strength is implied by collapse of lower fold limbs under the load applied by recumbent upper fold limbs. Sediment may be ponded by, or drape over, slump folds, but slumps are not eroded, a reflection on the low energy of the depositional environment.

The olistostrome contains large (up to 1 km) slide blocks of ophiolitic lithologies. Is this a sedimentary accumulation, or a tectonic mélange? The former is true: using the above criteria the olistostrome is demonstrably sedimentary, and also the slide-blocks show a unique distribution of emplacement-related structures (at base) and scree-like breccias (at sides and top). A tectonic mixing model could not generate such a bipartite distribution. Many earlier criteria for discriminating tectonic and sedimentary mélanges are not valid; each case must be viewed individually.

Slump structures in the western end of the Lower Palaeozoic Longford–Down Inlier, Ireland. J. H. Morris

Pre-slaty cleavage structures are present in an Ordovician greywacke sequence of massive sandstone, medium (12–50 cm) and thin bedded turbidite and siltstone, in the Longford–Down inlier, Ireland. Overall, the sequence is inverted, but in detail the strata regularly alternate between inverted and right way up and are cross cut by the cleavage with a consistent orientation.

The structures are considered in two categories, *coherent* and *disaggregated*, although they are spatially inter-related throughout the sequence.

The *coherent* structures vary from isoclinal, generally concentric folds with axial planes parallel to bedding to sinusoidal folds of variable shape. Fold wavelengths vary from a few centimetres to many metres. A welded contact is considered diagnostic of soft sediment deformation and, as the truncated beds young in the reverse direction to the regional stratigraphy, the bedding inversion is also a product of the same deformation.

Disaggregated structures are best developed in medium bedded turbidite. They vary from pull-apart and pinch-and-swell beds to discrete lenticular boudins which only crudely preserve the attitude of bedding. These structures occur sandwiched between and pass laterally into undeformed turbidite. The interpretation that these structures arose in soft sediments is based upon such features as the association with pre-cleavage folds; development of irregular necking towards a channel; inverted boudins overlain by a channel infill sequence and their presence in the medium bedded part of a fining-upwards channel infill sequence.

By comparison with the slump structures described by Gregory (1969), Helwig (1970) and Kleist (1974), it is concluded that both the coherent and disaggregated structures are the products of slumping. The spatial relationships suggest that the initiation of slumping may have been favoured by channeling. Liquefied flow deposits (Lowe 1976) are also thought to be present. These are fine- to medium-grained massive sandstones containing irregular patches of coarse sandstone—fine conglomerate and shale rafts up to 10 m in length.

A fore-arc basin model (Coulborn & Moberly 1977) of greywacke accumulation is suggested as a working hypothesis for the interpretation of the sedimentary environment of the area.

Sedimentary deformation structures in the fluviatile Cosheston Group, Lower Old Red Sandstone, SW Dyfed, Wales. R. J. Thomas

The three oldest ('lower Cosheston') formations of the five that comprise the 1500–1800 m thick Cosheston Group consist of intercalated intraformational conglomerates, coarse- to very fine-grained sandstones, and siltstones arranged in fining-upward cycles. These formations contain a greater abundance and diversity of sedimentary deformation structures than any other Old Red Sandstone sequence in South Wales. Over 95 per cent of these structures are products of vertical (as opposed to lateral) water and sediment movement.

Two groups of structures are recognised. Those of Group A affect 'heterogeneous lithologies', i.e., originally unlaminated or flat laminated siltstones interbedded on scales of

0.001–1.00 m with medium to very fine grained sandstones. Small-scale examples include load casts, ball-and-pillow structures, piled loaded ripples and convolute lamination. Laminae are preserved but disrupted. The formation of the structures involved differential loading which induced hydroplastic behaviour and partial liquefaction. Spectacular 'pillow beds', reaching 4.42 m in thickness also occur. Individual pillows are typically composite, contain 'wrap-around' structures, and apparently the development of subsequent (tectonic) 'concentric' fractures.

Group B structures are small- and large-scale rounded folds (maximum amplitudes and wavelengths: 1.90 m and 4.00 m, respectively). These affect 'homogeneous' medium to very fine grained, originally flat-, or low-angle parallel-laminated sandstones. Fold axes vary in orientation, whilst axial planes dip at 70°. Interlayer shear of some fold limbs caused a 'venetian blind'-like opening of micas within laminae, which facilitated the escape of water 'displaced' by adjacent foundering syncline cores. Laminae are intact because water escape rates were sufficient to promote folding but too slow to initiate fluidisation.

The incidence of sedimentary deformation within the 'lower Cosheston' increases upwards, reaching a peak in its uppermost (Mill Bay) formation and thereafter decreases rapidly, eventually dying out in the overlying 'upper Cosheston' succession. Cessation of faulting (hence of associated earthquake activity) that had led to the gradual emergence of the source area of the 'upper Cosheston' during 'lower Cosheston' times, would explain the observed distribution of deformation structures if they were seismically initiated. More significantly perhaps, during deposition of the Mill Bay sequence conditions for sedimentary deformation were ideal, but were generally unsuitable in the 'upper Cosheston'.

Seismically-induced load structures. J. D. Weaver

Within the basal Coal Measures (zone of *Anthraconaia lenisulcata*) near Abercraf (25 km northeast of Swansea), south Wales, and within the upper Visean (P_2 zone), near Wirksworth (19 km NNW of Derby), south Derbyshire, a number of liquefaction layers are seen displaying ball- and-pillow structures.

Those at Abercraf consist of balled-up, ovoid masses of sandstone 0.25–0.3 m thick and 0.5–0.7 m across within a matrix of siltstones (Weaver 1976). These sandstone masses have convex lower surfaces and planar tops, the laminae within the pillows are turned upwards at the margins and the underlying matrix shows the development of flame-like structures between contiguous pillows. These pillowed horizons are located just over 500 m northwest and southeast of the Swansea Valley Fault and die out into undisturbed sandstone beds away from the fault. Such horizons are not found in exposures of the same rocks further away from the fault. The ball-and-pillow structures near Wirksworth are composed of a coarse biosparite in a micritic matrix. These pillows, which are 0.6–0.8 m thick and 1.5–4.0 m across, show similar features to those at Abercraf and are located about 500 m east of the Gulf Fault.

Load structures of this sort are thought to be mainly the result of reversed bulk density gradients produced by the superposition of sand upon mud or by the deposition of more densely packed sand on less densely packed sand. The porosity of freshly deposited sediments is often high and loose packing will tend to make them unstable. A sudden change in this packing may result from the application of a sudden shock. This would generate a high excess of pore pressure which would support the grains. Experimental work carried out by Kuenen (1958) and Anketell *et al.* (1970) demonstrated that a shock wave could produce liquefaction structures of the ball-and-pillow type. Because the ball-and-pillow structures are restricted to exposures close to major fractures it is suggested that they were produced by the liquefaction of the underlying silt/micrite caused by shock waves from penecontemporaneous earthquakes.

If these structures are produced by shock waves emanating from fault lines, then their presence will indicate the development of such faults during sedimentation. It is therefore suggested that the Swansea Valley Fault was developing during basal Coal Measure times and the Gulf Fault during upper Visean times.

Convolute laminations in the Turbidites of the Aberystwyth Grits. R. K. Leppard

Convolute laminations observed in the distal turbidite sequence of Silurian age at Aberystwyth were produced by deformation which occurred contemporaneously with deposition. The mechanism responsible for the vertical movement of the sediment during deformation was probably compaction concomitant with dewatering. It is postulated that the lateral movement could have been produced by the following features.

1. Rippled surface. Rapidly deposited climbing ripple sequences, a common feature in the distal turbidites, are composed of muddy, very fine-grained, poorly sorted sand of high porosity which readily becomes thixotropic when shocked. In this quasi-liquid state the change of surface relief from rippled to flat necessitates lateral movement within the sediment.

2. Mud capping. Convolute laminations could result if the main dewatering phase takes place after a mud capping has been deposited on to the unit. When dewatering occurs the grains in the lower portion of the unit will suffer minimal displacement whereas the upper portion of the unit will become thixotropic. The expelled water will rise through the sediment until it comes into contact with the mud layer. It will then flow laterally until it can escape through a rupture in the mud. A complex pattern of flow will result, creating varying degrees of deformation within the unit.

3. Differential Compaction. It is feasible to suppose that when the sediment is subjected to some form of shock, compaction could evolve about discrete foci creating a series of depressions at the surface which would be filled instantaneously by the surrounding sediment, thus producing the requisite lateral movement in the upper portion of the unit. Different rates of collapse could result from variations in the packing density of the sediment.

It is thought that deformation is initiated by microseisms produced by an impending turbidity current. Seismic activity associated with a developing sedimentary basin would probably be too far from these distal units to initiate dewatering.

The theory relating to the deformation of climbing ripple sequences agrees in part with findings published by Kuenen (1953 & 1968).

Magnetic fabric measurements on flysch from the Southern Uplands H. ab Iorwerth & W. H. Owens

Measurements have been made of the magnetic fabric of Ordovician and Silurian rocks from the Rhinns of Galloway, Wigtownshire (ab Iorwerth 1968). This technique has potential uses in detecting very weak, otherwise indiscernible, fabric elements, and for distinguishing various modes of origin.

Magnetic fabric analysis (Owens & Bamford 1976) relies on the measurement of induced magnetization. It is thus not confined to ferrimagnetic minerals. The response of an anisotropic specimen will vary with the direction of the applied field, so that from measurements in different orientations the anisotropy can be calculated. Although all classes of magnetic minerals show a linear response to a weak (Earth's Field = 50 μT) magnetic field, they can be differentiated by measurements at high (0.1–1 T) fields, in which ferrimagnetic minerals saturate. All the specimens considered here showed a linear response at high field, suggesting that anisotropic susceptibility in paramagnetic minerals was responsible for the fabric. The magnetic fabric of a rock arises from the preferred orientation of grains which are themselves anisotropic. If the grain anisotropy is related to grain shape, either directly, or indirectly through crystallographic control, the technique may be used to provide information on grain preferred orientation. Its advantages are its speed and its sensitivity to weakly developed fabrics. Its disadvantage is that susceptibility anisotropy (the property usually measured) is inevitably of orthorhombic symmetry, irrespective of the grain fabric symmetry, so that the information conveyed by a measurement is limited to the description of an ellipsoid. Nevertheless, characteristic magnetic fabric styles, corresponding to conditions of deposition or deformation, can be recognised (Rees *et al.* 1968, table 1).

In the Rhinns of Galloway four basic fabric types are found:

(a) Depositional-oblate ellipsoids, often slightly imbricate with respect to depositional surface, with poorly defined maximum axes scattered about the current direction.

(b) Convolution-oblate triaxial ellipsoids with maximum axes well grouped parallel to the axes of convolution, and minimum axes girdled about the bedding pole. The fabric style is characteristic both of the laminated and of the underlying graded intervals of flysch units.

(c) Fold-ellipsoid shape and orientation are systematically related to position around the fold (cf. Owens & Bamford 1976. fig. 3) with maximum axes parallel to fold axes.

(d) Cleavage-oblate ellipsoids defining a magnetic foliation plane parallel to cleavage. Maximum axes within this plane may define a lineation not otherwise apparent.

In the Rhinns of Galloway depositional fabrics were noticeably rare. Convolution fabrics were common and often related to local folding, reflecting a possible relationship between soft sediment and larger scale deformation.

Compactional strains. D. Sanderson

Estimates of finite strain during compaction have been obtained by a variety of methods, most of which fall into 3 main groups:

(a) use of shape markers, particularly fossils, of which those with either a spherical or quasi-cylindrical form are particularly common and simple to use;

(b) measurement of shortening normal to bedding using features such as clastic dykes;

(c) comparison of differential compaction of sandstones and mudstones, usually assuming that the former undergo negligible compaction and hence calculating the compaction of the latter.

Development, testing and comparison of these methods has been undertaken using a wide variety of rock types. Decompaction numbers (N_D) are calculated, where $N_D = t_0/t_D$ and t_0 = original bed thickness, t_d = present bed thickness. Generalised results are shown in Table 1. We may interpret N_D for many argillaceous sediments in terms of initial porosity (ϕ_0) when compaction is achieved simply through reduction in pore space.

$N_D = 1\phi_D$ where ϕ_D = present porosity, which tends to zero with increasing depth and hence:

$$\phi_0 \approx 1 - N_D^{-1}$$

Estimates of ϕ_0 obtained from differential compaction studies accord well with observed initial porosities of marine sediments, whereas ϕ_0 from clastic dykes, for example, correspond more closely with porosities at about the Liquid Limit; open cracks could not form at higher porosities. Fossils give lower ϕ_0 estimates, between the Liquid and Plastic Limits.

Thus different methods record different aspects and amounts of the compaction history and calibration of the data may allow crude estimates of the Atterberg limits.

The large strains (N_D = strain ratio) during compaction must be considered in models of sedimentary sequences, in reconstructing sedimentary structures and fossils, for example, and as initial input in the interpretation of tectonic strains from deformed markers.

Experimentally deformed sediments and the early generation of cleavage. A. J. Maltman

Argillaceous sediments with up to 40 per cent saline water contents have been experimentally deformed by (i) triaxial compression at strain rates between 10^{-3}/sec and 10^{-7}/sec and confining pressures up to 700 bars, and (ii) direct and distributed simple shear at similar strain rates and with normal stresses up to 14 bars.

The materials are weak and usually macroscopically ductile, but in thin-section six types of microstructure have been recognized, *viz.* i) ductile shear zones, ii) kink-bands, iii)

Rock type (and method)	Decompaction number
Shale (differential compaction)	4.0–7.0
" (fossil markers, clastic dykes, etc.)	1.5–3.5
" (channel fill)	1.5–2.5
Siltstone	1.5–2.5
Sandstone	1.1–1.5
Carbonate mud	2.0–4.0
Limestone	1.0–2.0
Coal	3.0–15.0

TABLE 1
Typical ranges of decompaction numbers in various sediments

'flaser' structure, iv) layer-boundary crenulations, v) crenulations, vi) 'creases'.

Most of these microstructures are observed in sedimentary rocks, so that the apparent absence in nature of the 'creases' is puzzling, especially as they are commonly formed in the experiments. The explanation may lie in their being obscured by slaty cleavage, of which they may be a precursor. A sequence can be envisaged for the development of some slates, which begins with the primary settling fabric, important in many argillaceous sediments, being 'creased' by tectonism, as in the experiments, and a secondary foliation initiated. The spacing and orientation of this will have influence on the ensuing modifications. Continuing tectonism and lithification, and the onset of such processes as pressure solution and recrystallization enhance the 'creases' and establish a spaced slaty cleavage. Incipient metamorphic differentiation augments its domainal nature.

In contrast with the hypotheses of Maxwell (1962) and followers, this sequence does not depend on pore-water expulsion (and the concomitant convolution and disruption of laminae), and only an incipient cleavage is generated before lithification.

Petrographic examination suggests that this succession of events may have operated on the Lower Palaeozoic argillaceous rocks of central Wales.

Strain and planar fabric development in slate. W. Davies

Stereo-pairs of SEM photographs have been used to resolve the fine-scale structure of Lower Palaeozoic slates from Wales and of mudstones and marine band shales from the Yorkshire Coal Measures. The mudstones exhibit 'edge-to-edge' arrangement of clay flakes. Shale and slate reveal planar fabrics, by no means wholly penetrative, in which layers of platy minerals are separated by flakes which are randomly arranged. In some instances, the layers merge and part to produce a lenticular character. The spacing of these layers determines the fissility of the rock. Within the plane of fissility, the fine structure in shale exhibits no preferred elongation, nor do the flattened impressions of goniatites and pectinoids. In slate, phyllosilicates in the plane of fissility are corrugated in the direction of principal extension derived from distorted trilobite impressions.

It is considered that both planar fabrics developed by the progressive collapse of an edge-to-edge structure while pore-water was present. In slate, this was accompanied by slippage along the layers of flakes as when the higher cards in a pack move over the lower ones. Reduction spots are interpreted as post-slippage reduction controlled by the fabric.

The slippage mechanism is associated with the conditions of accumulation of the Welsh Lower Palaeozoic sediments, here compared with the present-day accumulations off the eastern coast of the USA in which creep, sliding, and décollement folding are important phenomena. Excepting local volcanic piles, possibly no more than 1500 m thickness of Lower Palaeozoic sediment ever accumulated at any one point on an irregular slope bounded by the Borderland shelf. The sediment, mainly clay of a consistency between the Atterberg plastic and liquid limits, responded to local changes in loading as the sedimentary pile accumulated. This, rather than a single regional, post-Wenlock compression, determined the cleavage pattern across Wales with all its sharp changes of course.

Well-cleaved slate is rare in Wales; possibly its origin lies in the localised development of quick-clay conditions on the Lower Palaeozoic slope, conditions which developed in limited thicknesses of strata.

Cleavage-fold relationships in the Aberystwyth Grits—a preliminary report. W. R. Fitches & Ruth Johnson

The Lower Palaeozoic rocks of mid-Wales have been involved in a deformation sequence that first produced the regional nearly N–S folds with steep axial planes and cleavage, then localised flat-lying folds, kink bands and crenulation cleavages, followed by rare, steep NW–SE folds and crenulation cleavages. The first structures are being studied in the Aberystwyth Grits in coastal exposures near Aberystwyth.

Fold geometries and fold-cleavage relations are commonly complex in detail giving features which can be considered in terms of soft-sediment deformation as an alternative to deformation of already lithified rocks. However, no satisfactory criterion has been found whereby distinction can be made. For example, some folds have quartz-carbonate fibre slickensides parallel to the slip direction expected during flexural slip on bedding planes, others have veins with quartz and carbonate fibres consistent with development during outer arc extension. But until it is known if extension fibre growth is restricted to lithified sediments it is not possible to use these structures as a criterion for deciding the state of lithification at the time of folding.

A beach platform at Clarach (SN 5855 8425), about one mile north of Aberystwyth, is being mapped on a scale of 1:50 to study the detailed relationships between cleavage and folding. In thin section the cleavage appears as thin seams of opaque dust, poorly developed and widely spaced in the sandstones but well developed in the argillaceous rocks in which they are spaced about 0.01 mm apart. There is usually no preferred alignment of phyllosilicates in the argillites except locally in narrow zones parallel and adjacent to the seams. Large (0.1 mm) penninite chlorites are abundant in the pelites, characteristically aligned with their 001 cleavages parallel to bedding. Many are bent or kinked on axial planes parallel to cleavage whilst some appear to have had their ends or middles removed by pressure solution on planes which are parallel to and continuous with the cleavage seams. It appears, therefore, that these chlorites were formed before the cleavage and that this structure involved flattening and pressure solution.

The origin of the chlorites is uncertain. Bjørlykke (1971) favoured a detrital origin whereas Evans & Adams (1975) regarded them as products of diagenesis or metamorphism. Current work is designed to elucidate their origin as clearly they are potentially useful in establishing the state of the rocks at the inception of cleavage, whether soft or lithified.

The folds can be divided into three types according to their relationships to cleavage. Some folds are cut obliquely by the cleavage and are referred to for convenience as 'pre-cleavage folds', in others the cleavage is itself folded in places ('post cleavage folds'). The situation appears to be analogous to that described from Tasmania by Powell (1974) and there are the same problems of interpretation; were the folds, cleavage or both diachronous with respect to each other? On current evidence it is our view that in the Aberystwyth Grits folding took place over a relatively longer period than the cleavage.

Report of field excursion, 6th March, 1977. R. Cave

Morning. Dr R. Cave led the party of about 75 persons over the north–south pericline at Carn Owen, $7\frac{1}{2}$ km east of Talybont (SN 7337 8809). Here the Drosgol Grits Member of the Ordovician sequence is well exposed both naturally and in two quarries. In the quarry in the east limb the 'Grits' consist of thick, graded and evenly bedded greywackes.

This regularity of bedding persists up the limb but in the core of the fold, as seen in the main quarry, severe disruption has occurred, contorting part of the greywackes beneath the envelope of mudstones (Cave 1967, pl. 1). Penecontemporaneous slumping is considered to have been the disruptive process. In the west limb, especially near the foot of the old tramway (SN 7285 8795) the style of deformation is again different. Dispersion of sedimentary grains and reaggregation has produced a new textural homogeneity in the form of juxtaposed balls and pillows.

Afternoon. The remainder of the meeting was conducted on the foreshore around Aberystwyth, on exposures of the Aberystwyth Grits Formation (Silurian).

1. At Constitution Hill (SN 5835 8267), Mr R. Leppard illustrated his paper of the previous day on the production of convolute lamination by penecontemporaneous collapse of a primary ripple sequence. He used the thicker turbidite sandstones to show that whilst a few retained ripple-lamination the majority had become convoluted.

2. At Clarach (SN 5855 8425), Dr W. R. Fitches and Mrs R. Johnson showed examples of the three main cleavage-fold relationships (see abstract) seen on the wave-cut platform. Discussion centred particularly on whether folding, cleavage, or both structures evolved over a protracted period or whether cleavage could be regarded for reference purposes as marking a relatively short time interval.

3. At Allt Wen (SN 5764 7954), Dr Cave pointed out several surfaces which he interpreted as inter-turbidite bedding slides. The 'slide' surfaces are grooved and commonly coated by thin films of fibrous quartz and carbonates. At other places such surfaces are thrown into small ripples. Warren *et al.* (1970) attributed similar phenomena in Denbighshire to flexural slip during folding. Dr Cave considered such an origin unlikely in the Aberystwyth area as both the grooves and ripples pass unmodified from the limbs over the hinges of folds. Nearby a décollement structure separates uniformly dipping beds from a zone of highly folded strata above. He considered that most of the sliding occurred before folding and before lithification, but acknowledged that inherent in this explanation is the need for fibrous growth in unlithified or only partially lithified materials.

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