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Early veins as evidence of detachment in the Lower Palaeozoic rocks of the Welsh Basin

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Abstract—A suite of small-scale structures, mostly represented by veins, is widely developed in the Lower Palaeozoic clastic sedimentary rocks of the Welsh Basin. The structures were produced after considerable dewatering and diagenesis had taken place, but before the host rocks were folded and cleaved by the regional deformation. Bedding-parallel veins, the most common of the structures, are attributed to hydraulic jacking and mineralization by overpressured pore-fluid, probably at burial depths of several hundred metres or more. The various burial processes, including the production of the veins, are regarded as being diachronous, in effect propagating up through the sediment pile as sedimentation and burial progressed. The bedding-parallel veins are considered to represent inclined detachment surfaces, sole planes down which rock masses slid due to gravity. These surfaces were initiated in rocks originally inclined from the horizontal for sedimentological reasons or in horizontal host rocks which were later tilted by tectonic processes. Other structures in the suite were produced as a consequence of the downslope movement, for example by compression in the toe regions of the glide-sheet and by extension at the trailing edges. Slip directions, recorded by striations on the bedding-parallel veins, were mainly WNW–ESE in W Wales and N–S in NE Wales. It is anticipated that these structures will be found to be widespread in other sedimentary basins. That they have escaped attention elsewhere is probably because individual structures are small and seemingly insignificant whilst in deformed rocks they may have been attributed to folding or faulting.

INTRODUCTION

THE WELSH Basin (Fig. 1) is viewed widely as having been a marginal basin situated to the SE of the Iapetus suture (e.g. Phillips *et al.* 1976, Anderton *et al.* 1979, Kokelaar *et al.* 1984). The basin had its inception in Late Precambrian and early Cambrian times, continued to subside during the Ordovician and Silurian before losing its identity at the close of the Silurian, probably in Downtonian times.

Traditionally, the basin history was divided into a chronological sequence of separate events: sedimentation and volcanism, followed by alteration during burial, then a climactic end-Silurian compressional deformation and uplift. It is now recognized that this concept is over-simplistic and that many processes were commonly diachronous, progressive and interdependent over much of Early Palaeozoic time. For example, there are numerous documented records from various parts of the basin of tilting, folding and faulting accompanying and partly controlling sedimentation and volcanism (e.g. Shackleton 1954, Rast 1969, Kokelaar 1979, Campbell 1983, 1984, Reedman *et al.* 1984). Moreover, the concept of early extension promoting subsidence and late contraction causing regional deformation is also under revision. Woodcock (1984a, b), in his reviews of early Palaeozoic sedimentation and tectonics in Wales, has postulated an important role for strike–slip deformation at various stages of basin evolution. The Welsh Basin seems to have been a site of complex interplay between transtensional and transpressional tectonics during much of its history.

It has also been recognized that the rocks have been widely altered by diagenesis (Evans & Adams 1975, 1976, Craig *et al.* 1982) and by low-grade metamorphism (e.g. Roberts 1981, Bevins & Rowbotham 1983) due to burial and, according to Merriman & Roberts (1985), to deformation.

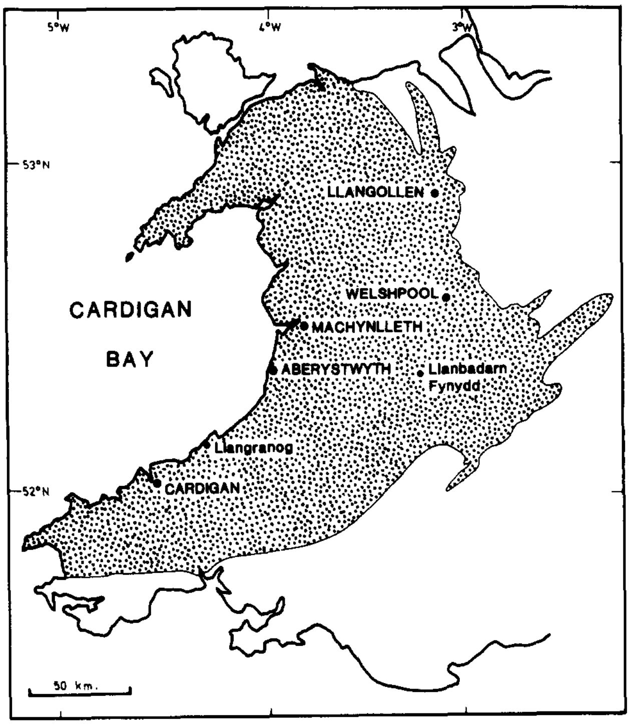

Fig. 1. The outcrop of the Welsh Lower Palaeozoic Basin and the positions of the main localities referred to in the text.

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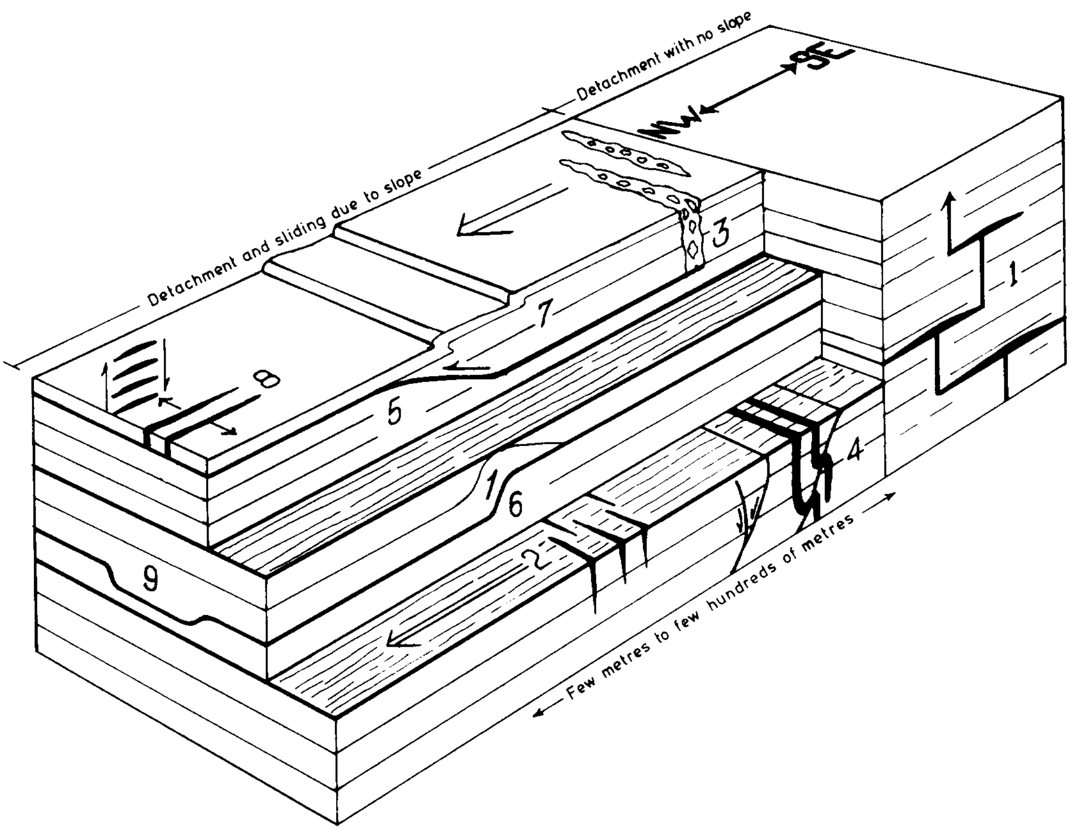

Fig. 2. Synoptic diagram to show geometrical interrelationships between various types of detachment structure in the Welsh Lower Palaeozoic Basin. Bedding, fine lines; veins, thick lines. Note that in parts of W Wales the glide direction is opposite to that indicated whilst in NE Wales the direction is N or S. 1, Upward propagating system of steep veins and bedding-parallel veins (after Fyfe *et al.* 1979, figs. 11 and 18); 2, striated bedding-parallel veins and NE–SW bedding-normal veins (e.g. Figs. 3b and 5a); 3, hydraulic fracture vein breccia (Fig. 5b); 4, extensional structures at trailing edge of glide sheet (Figs. 5b and 6d & e); 5 and 6, thrusts and backthrusts (Figs. 4c & d); 7, asymmetric folds (Figs. 6a–c); 8, NW–SE bedding-normal veins and tension gash arrays (Fig. 7f); 9, lateral ramp.

Several stages in this complex continuum of processes, from sedimentation through diagenesis and burial metamorphism to the development of cleavage and folds, have already been defined or at least a framework established for future detailed work. This paper is concerned with a suite of small-scale but widespread structures, mainly brittle and vein-filled, many of which mark detachment and lateral displacement of beds. The structures were produced progressively during burial after substantial dewatering and diagenesis but before the development of the regional cleavage and folds. They are interpreted here as largely a response to overpressured pore fluids.

Most examples of the detachment structures and the interpretations of them are derived from the Caradocian to mid-Llandovery sedimentary rocks of the Cardigan Bay coast (Fig. 1). Here the widest variety of structures is developed and their relations to the diverse fabrics and structures produced during sedimentation, burial and deformation are most clearly displayed. It will be shown, nevertheless, that similar structures are widespread in the Welsh Basin. Moreover, it is anticipated on theoretical grounds that these types of structures are likely to be common in most sedimentary basins, but because individual structures are small and seemingly insignificant they have been largely overlooked.

The various members of the suite are described in the next section. The preceding burial events and the later deformation are then briefly outlined so that the chronology of the detachment structures in the continuum of events can be demonstrated. Lastly, consideration is given to the origin of the structures.

DETACHMENT STRUCTURES

The term detachment as used in this paper refers specifically to the separation of sedimentary rocks along a bedding-parallel fabric and also to concomitant bedding-normal tensile separation of these rocks. The term detachment structures encompasses all small-scale structures, such as veins and certain folds, which we consider to be consanguinous with detachment and sliding. Figure 2 illustrates the various structures in diagrammatic form.

Cardigan Bay coast

Bedding-parallel veins. These structures, first recognized by Davies & Cave (1976), are the commonest members of the suite. The veins are composed mainly of quartz and ferroan dolomite (Cave & Hains in press) in various proportions and some contain small amounts of chlorite, white mica and pyrite; galena was found in one vein in the core from the Glanfred borehole near Aberystwyth (Cave & Hains in press). The veins are mostly less than 5 mm in thickness but some attain 5 cm (Fig. 3a). In places they can be traced for at least 25 m to

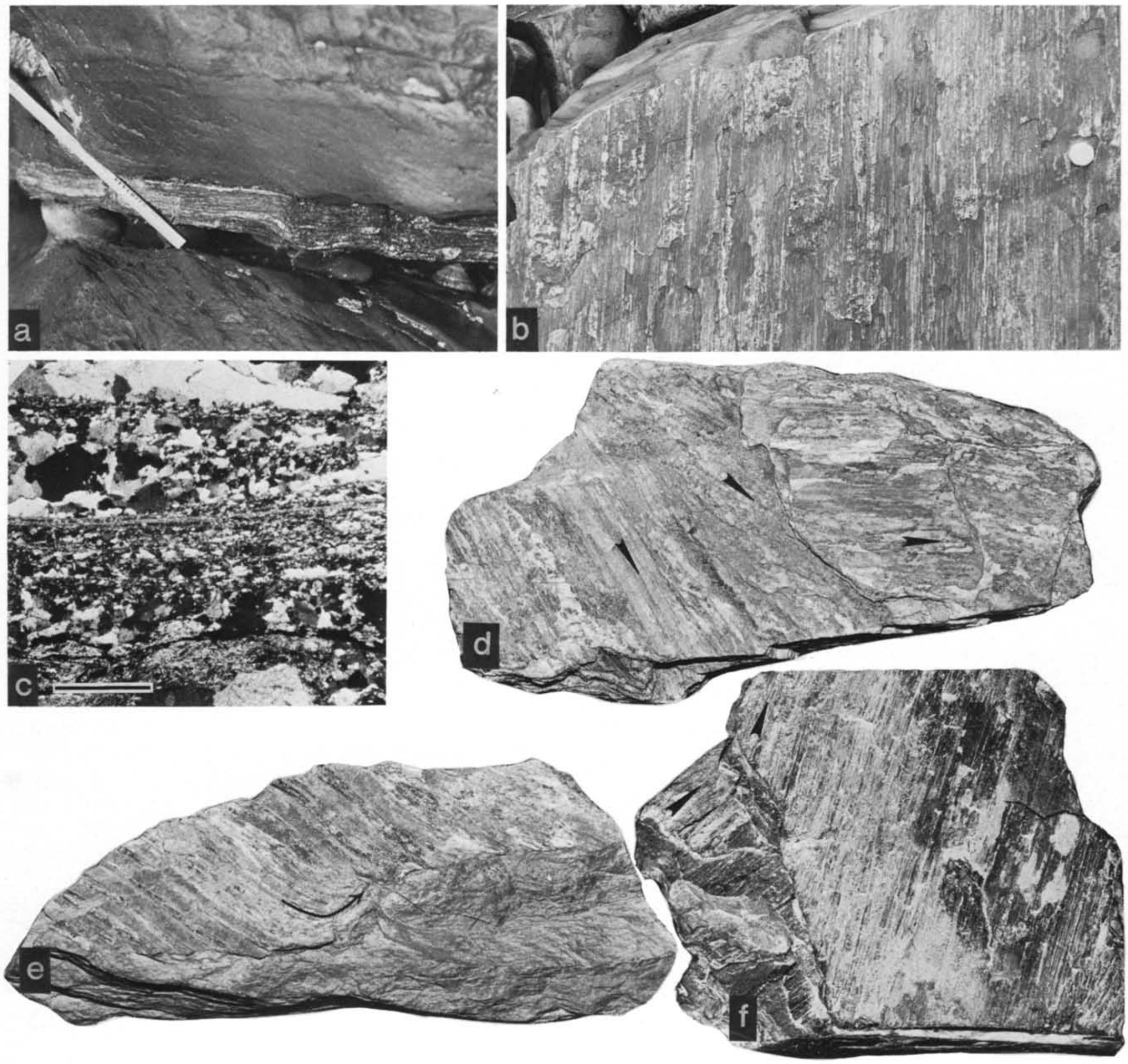

Fig. 3. Examples of detachment structures. (a) Composite bedding-parallel vein, Borth beach [SN 603886], 8 km N of Aberystwyth. Bedding visible in top part of figure. Pencil for scale. (b) Striated bedding-parallel vein near Cormorant Rock [SN 582829], 2 km N of Aberystwyth. Coin for scale. (c) Photomicrograph of part of quartz–carbonate vein, showing zones of sub-grained quartz interlayered with large quartz grains. Crossed polars. Scale bar: 1 mm. (d) Striations with different orientations (arrows) on successive surfaces of a composite bedding-parallel vein. Hand-specimen, maximum length 20 cm. Pant Glas Quarry [SJ 215476], near World's End, Llangollen area. (e) Curved striations (arrow) on single, plane surface of bedding-parallel vein. Hand-specimen, maximum length 15 cm. Bryn-y-afr [SN 746878], 9 km SE of Machynlleth. (f) Striations with different orientations on successive surfaces of a composite bedding-parallel vein. Hand-specimen, maximum length 15 cm, from the tilted, but otherwise undeformed, late Precambrian Stoer Group, Stoer Point, NW Scotland. Specimen provided by M. R. Evans.

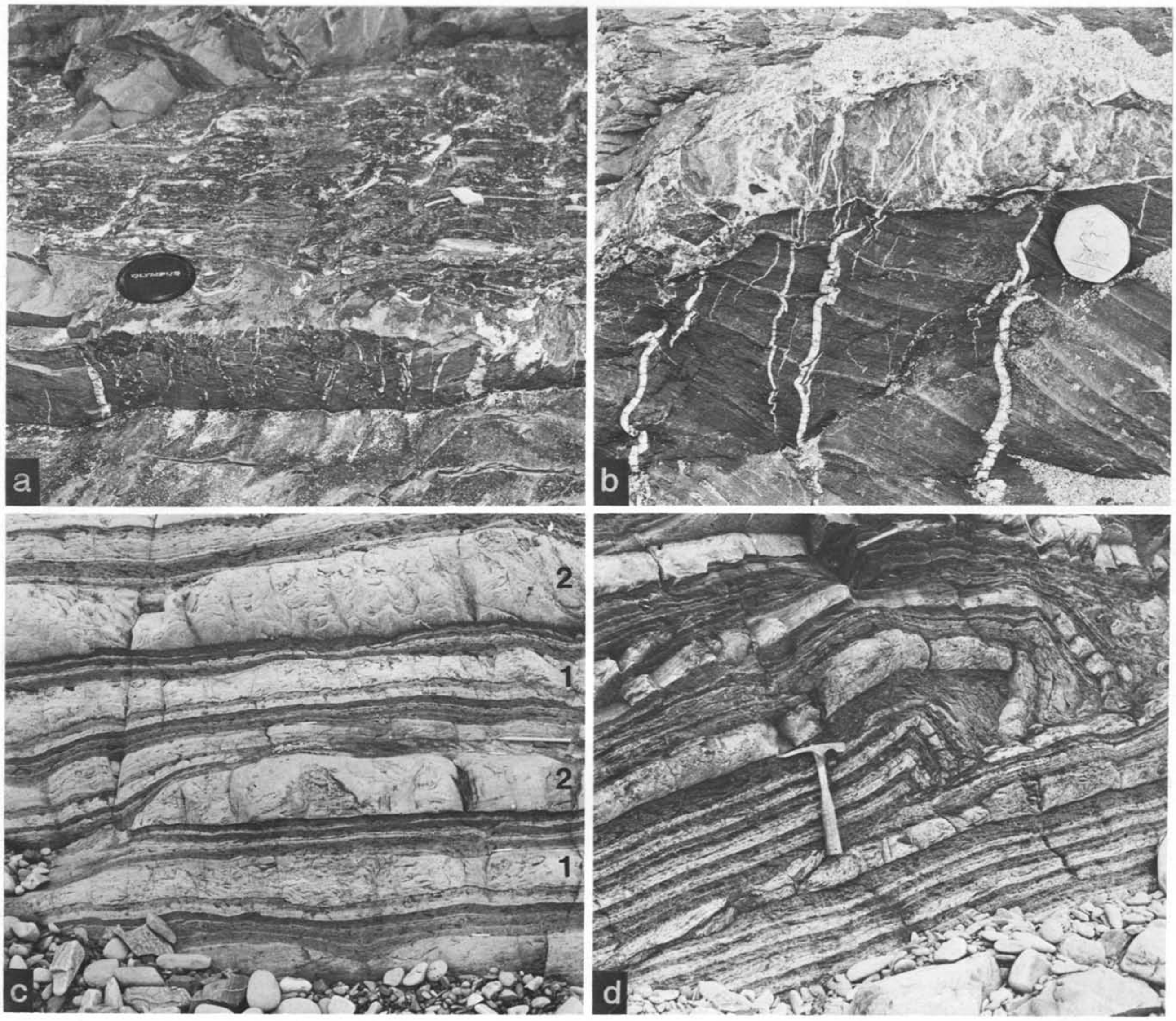

Fig. 4. Examples of detachment structures. (a) NE–SW bedding-normal veinlets exposed in steep dark face below lens-cover. Lens-cover rests on gently folded striated bedding-parallel vein. Cove 200 m N of Llangranog [SN 310542]. (b) Vein breccia (above coin), at high angle to bedding, cut by folded NW–SE veins. Traeth-Penbryn [SN 288521], 3 km SW of Llangranog. Coin for scale. (c) Thrust with staircase geometry duplicating strata of the Aberystwyth Grits Formation. Allt Wen [SN 575794], 2 km S of Aberystwyth. Viewed from SW. Pencil for scale. (d) Thrust with hangingwall anticline. Locality as for Fig. 2(g). Viewed from SW. Hammer for scale.

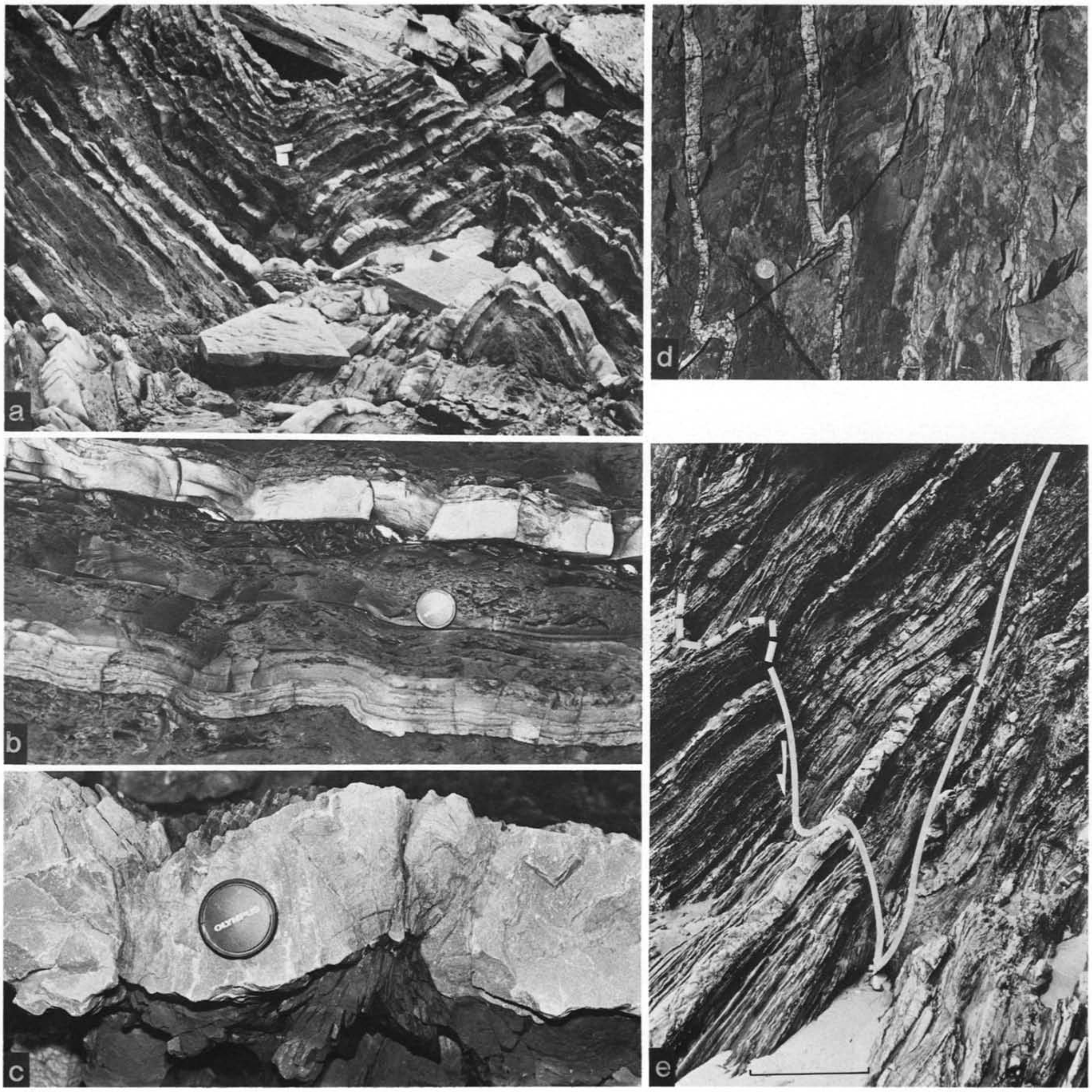

Fig. 6. Folds, listric faults and other structures associated with detachment. (a) Syncline–anticline pair produced by the regional deformation. Near Cormorant Rock [SN 582829], 2 km N of Aberystwyth. Scale: notebooks in syncline core. See text for discussion of structure. (b) Asymmetric folds, with carbonate-filled inner arc saddle-reefs indicated by arrow, in left (SE) limb of syncline shown in Fig. 6(a). Lens cap for scale. (c) Asymmetric fold with cleavage fan, convergent in dark argillite and divergent in light arenite. Left (SE) limb of syncline shown in Fig. 6(a). Lens cap for scale. (d) Listric normal fault, downthrowing to left (NW), and associated veins. Crug Cove [SN 178515], 8 km NE of Cardigan. Coin for scale. (e) Bifurcating faults, each with concordant striated veins, viewed from SW. NW fault has listric normal geometry and up to 20 cm displacement; irregular trace is due to rough topography of exposure. Amount of displacement on SE fault is unknown. N side of Llangranog bay [SN 310542]. Scale line: 1 m.

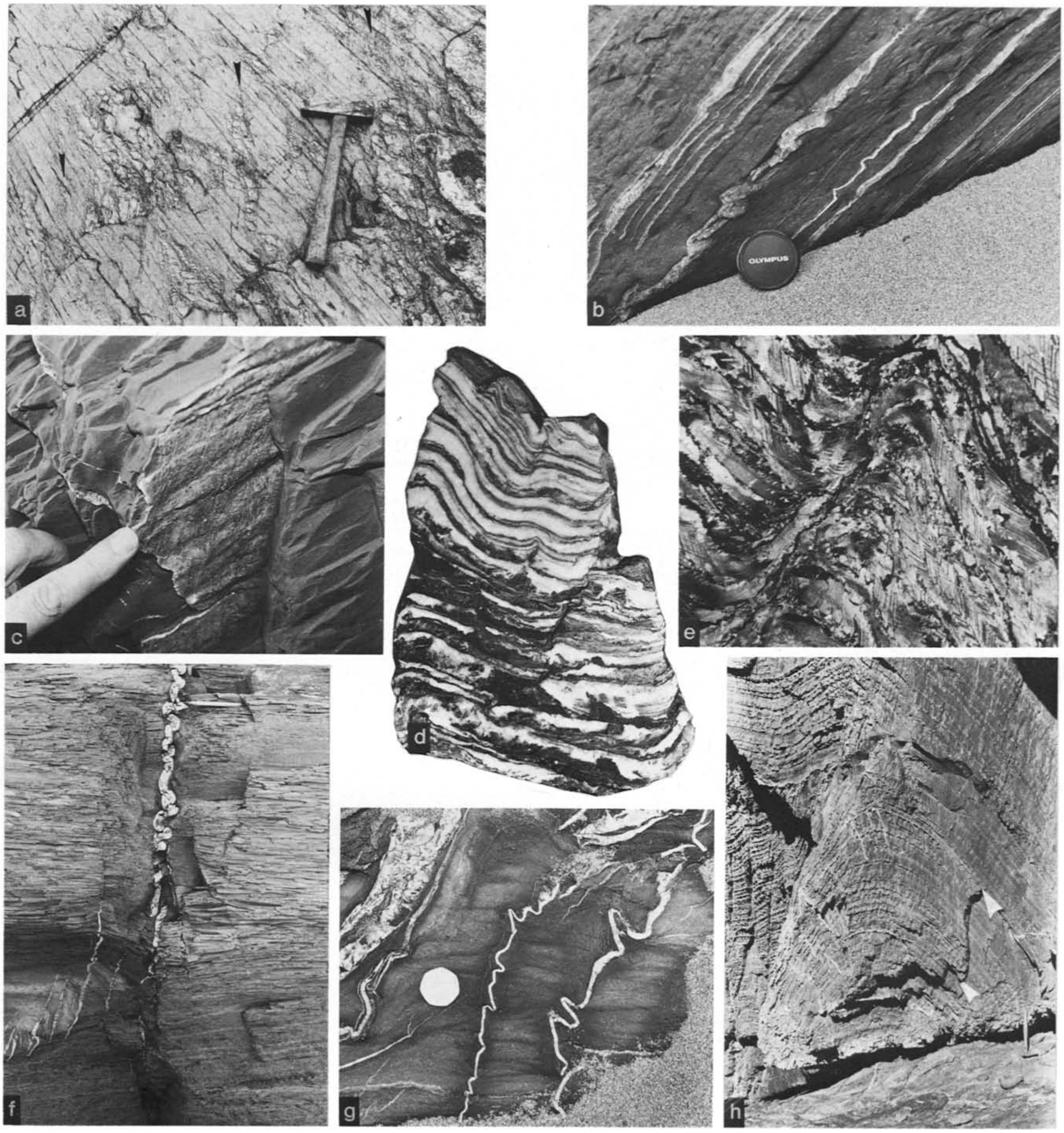

Fig. 7. Detachment structures and their relationships to burial features and regional deformation structures. (a) Bedding plane grooved by movement of concretions. Grooves indicated by arrows. Dark areas on right edge are impressions of concretions. Cleavage traces from top left to bottom right. Crags on N side of Machynlleth golf course [SN 755004]. (b) Bedding-parallel veinlet (partly covered by lens-cover) buckle-folded with thin, light coloured sandstone layers. Cove 200 m N of Llangranog [SN 310542]. Lens cap for scale. (c) Bedding-parallel veinlet gently buckled on fold hinges plunging parallel to finger. Note the strong striations making a 45° angle with the fold hinges. Locality as Fig. 6(b). (d) Gently folded composite vein. Hand-specimen, maximum length 20 cm. Pant Glas Quarry [SJ 215476], near World's End, Llangollen area. (e) Photomicrograph of folded calcite–quartz layers in a composite vein. Calcite shows twinning. Axial-plane pressure-solution surface runs top centre to bottom left. Scale bar: 1 mm. Locality as Fig. 6(d). (f) NW–SE vein and tension gash array deformed by folds of the regional deformation. Cleavage, marked by pencil, is congruent to the folds. Traeth-Penbryn [SN 288521], 3 km SW of Llangranog. (g) NE–SW veins buckled in cleaved argillites. Coin for scale. Locality as Fig. 6(f). (h) Bedding-parallel veins (indicated by arrows) and bedding-normal veins folded by anticline produced by the regional deformation. Hammer for scale, bottom right. Penyrhwbyn [SN 181517], 8 km NE of Cardigan.

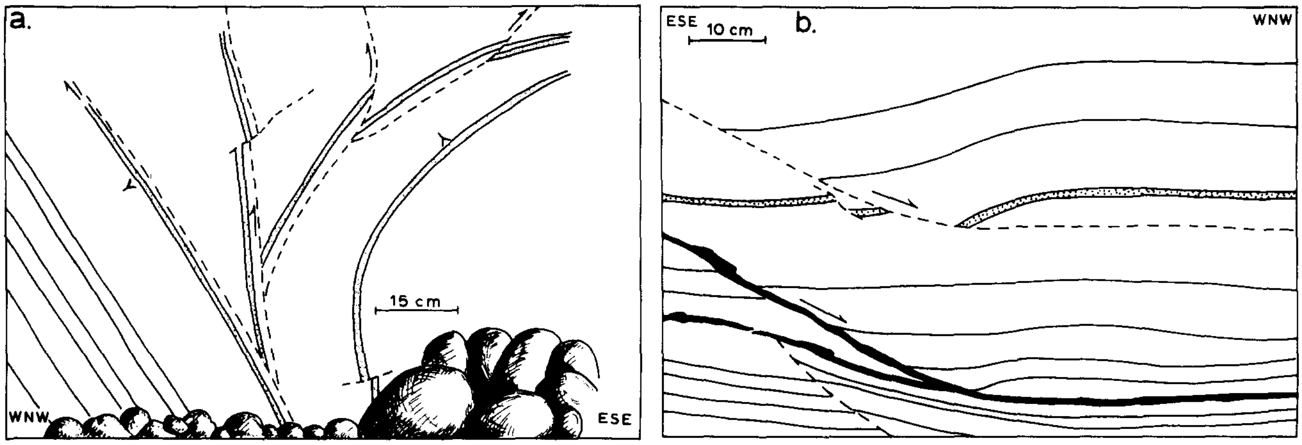

Fig. 5. Field sketches of detachment structures. (a) Imbricate fan, probably in an antiformal stack. Bedding, solid lines; faults, pecked lines; sandstones, stippled; younging, arrowed. Borth beach [SN 603886], 8 km N of Aberystwyth. (b) Listric normal faults with rollover anticlines. Veins along detachment planes shown in black. Traeth y Gaerglwyd [SN 321549], 1.5 km NE of Llangranog.

the limits of exposure. Most thin veins are megascopically homogeneous. Those thicker than 5 mm, however, are composed of several parallel laminae, each of microscopic to few millimetres thickness, with different mineral proportions and grain sizes. The laminae are separated either by discrete fracture surfaces or more rarely by laminae of country rock.

The quartz and carbonate grains are mostly in the size range 0.1–0.2 mm whilst the other mineral components are much smaller. In thin sections the grains usually show no preferred dimensional or optical alignment. However, in small areas of some sections the quartz and carbonate have a weak preferred dimensional elongation in the plane of the vein, and in rare instances quartz is elongated at about 45° to the vein margin. The minerals are commonly strained, the carbonate containing twin-lamellae and the quartz showing deformation bands, sutured margins and sub-grains.

The fracture surfaces separating laminae in the thicker, composite veins are seen under the microscope as discrete planes. Some planes pass cleanly through large quartz and carbonate grains without offsetting or apparently disturbing them in any way. Others are adjoined by very fine quartz which is probably derived by sub-graining of adjacent large grains by cataclastic deformation (Fig. 3c).

Characteristically the surfaces of the veins and the laminae comprising the thicker veins are striated (Fig. 3b). The ridges and grooves have amplitudes and spacing of less than one millimetre in most instances although coarser striations are found locally. The lineations on any one surface are usually aligned uniformly and trend mostly in the azimuth range 050–120°. In composite veins, striations on surfaces of adjacent laminae are commonly aligned in different directions (Fig. 3d). In the Cardigan Bay area the azimuth difference is rarely more than 20° but elsewhere, notably in the Llangollen area discussed below, the differences can be much greater. Very rarely the striations on one surface curve abruptly through almost 90° in the plane of the vein (Fig. 3e).

The spacing of the veins is variable. In parts of the succession they are sparse and one vein is separated from the next by many metres of strata. Commonly, however, they are abundant and are spaced at intervals of 10–15 cm through several metres of strata (e.g. Allt Wen [SN 575794], near Aberystwyth). Some veins separate beds of contrasting lithology, notably sandstone from superjacent argillite, but the majority are located in apparently homogeneous argillite.

Thrusts and associated structures. Minor thrusts with displacements ranging from a few centimetres to several metres are found locally, notably in the Aberystwyth Grits. Some thrust planes are marked by striated quartz–carbonate veins indistinguishable from those described above. Bedding-plane thrusts are probably more common than has been recognized but in these rhythmically bedded rocks they are detectable only where ramps are exposed or where duplication of strata can be identified. Figures 4(c) and (d) show examples of duplication, ramps and folds associated with thrusts, and Fig. 5(a) shows an imbricate fan.

Ramps are rare and mostly exposed only on single surfaces so meaningful statistical orientation data have not been obtained. Some ramps dip southeast and others northwest, parallel to the NW–SE movement direction inferred below, indicating the presence of thrusts and backthrusts. Examples of ramps striking nearly NW–SE are regarded as oblique or side-wall ramps.

Directions and amounts of displacement. The striations give the azimuth of displacement on bedding-parallel veins, that is, between southwest and northwest but predominantly WNW–ESE. Only in exceptional circumstances has the sense of displacement been determined; for example, where the vein has a staircase thrust geometry, where objects such as pyrite which made the striations remain *in situ* or where opening up along minor steps in the surface can be recognized. From this sparse information displacements mainly towards the west-northwest and east-southeast have been determined. The movement direction on thrusts with striated veins along them is probably also shown by the striation directions. In rare instances where they can be observed the orientations of footwall or hangingwall ramps are consistent with mainly west-northwest or east-southeast movement.

The amount of displacement along bedding is difficult or impossible to measure in most instances. The lengths of striations produced by movement of recognizable objects such as pyrite show displacements of up to 30 cm on some surfaces. Similar displacements have been measured from the lengths of single striations.

Displacements of up to 50 cm can be recorded on some thrusts but the narrowness of outcrops, typically in wavecut platforms and cliffs, prevents measurements on larger thrusts. No examples have been found of a pair of footwall and hangingwall ramps belonging to the same thrust. A minimum displacement of nearly 5 m towards the northwest was measured on one thrust, from the point of truncation of a marker bed by a footwall ramp to the limit of exposure of that bed along the next flat.

The amount of movement on a single surface seems trivial. However, the striated bedding-parallel veins are abundant and the summation of their displacement is large. The veins are 10–20 cm apart over several metres of strata in places. With displacements of 30 cm or more on each surface, and very much more in total on the thicker, composite veins, shear strain values are greater than 1, possibly up to 2 or 3 locally. Accordingly, it is conceivable that several hundreds of metres simple shear displacement, without significant duplication, has been accomplished in places.

Bedding-normal veins. Bedding-normal veins are not widespread but in places are abundant, especially near Llangranog [SN 310 542] and on Traeth-Penbryn [SN 288 521] (Fig. 1). Several types and orientations have been distinguished.

NE–SW veins. These are mostly small structures, a few millimetres to 5 cm in width and rarely more than 10 cm in length (Fig. 4a). They occur variously as single veins, in groups comprising several veins spaced at intervals of 10 cm or more, and in dense arrays. Some of the latter are more than 5 m in width measured down the dip of the beds containing them and are composed of individual veins less than 5 mm wide spaced at intervals of 5 cm or less.

These veins are usually closely associated with the bedding-parallel veins. In places bedding-normal and bedding-parallel veins appear to merge or veins of either set can displace those of the other set. At Llangranog [SN 310 542] and locally elsewhere the bedding-normal veins form dense arrays beneath thick composite bedding-parallel veins (Fig. 4a). Here, and at several other sites, the latter are traversed by very fine bedding-normal veinlets which are completely contained within the bedding-parallel veins and intersect them on lines normal to the striations. Electron-microscopy revealed that the quartz of the veinlets is in fibres aligned normal to the veinlet walls and parallel to the striations on the bedding-parallel veins (Strong 1974).

A variant of the bedding-normal and bedding-parallel vein combination is shown in Fig. 6(d). In this case, quartz–carbonate veins in part follow a dip–slip shear surface which is slightly steeper than bedding, but elsewhere they are nearly normal to bedding and the shear surface; each vein takes on a step or fold-like form. Quartz fibres throughout the combination are aligned NW–SE, normal to the walls of the steep parts and parallel to the walls of the flatter parts. Evidently NW–SE extension took place across the steep veins whilst the flatter surfaces were used as shear planes.

NE–SW vein breccias. Vein breccias form irregular masses up to 50 cm across and lenses and planar sheets up to 10 cm wide, 2 m high and more than 10 m long (Fig. 4b). They are composed of country rock clasts in coarsely crystalline quartz which locally contains pyrite. The clasts are angular and the margins of some adjacent clasts show a perfect fit indicating only slight separation; some fragments remain partially attached to the wall rocks. All clasts appear to have been locally derived. Description of microtextural features of the clasts, pertinent to discussion of the timing of brecciation with respect to other processes, is deferred to a later section.

NW–SE veins and tension gashes. NW–SE veins are locally abundant, notably on Traeth-Penbryn [SN288521] (Fig. 1) where they accompany and in places cut the breccias described above (Fig. 4b). The veins are up to 10 cm in width and up to 3 m in length. They occur singly or in groups up to 5 m in width and more than 10 m in length. Quartz is the dominant mineral, usually massive, but locally as fibres elongated normal to vein walls; ferroan dolomite is found in small amounts.

Some veins are disposed en échelon in tension-gash arrays aligned WNW–ESE (Fig. 2f). This configuration and the NE–SW quartz fibre elongation indicate dextral slip along the arrays under NW–SE compression.

Small folds and cleavage. A set of 20–30 cm wavelength asymmetric folds, described by Wood (1958) as tectonic ripples, is exposed 2 km N of Aberystwyth at [SN 583 830]. These structures are tentatively assigned to the suite of structures described above. The folds (Figs. 6b & c) are in the east limb of a syncline (Fig. 6a) which has a steep NNE-striking axial plane and so is probably one of the regional tectonic folds. However, the small folds are not parasitic to the syncline because they have the opposite sense of asymmetry for that limb and are incongruent to the large structure. Some of the folds have cleavage fans, mostly weakly defined by closely spaced anastomosing surfaces but in some sandstones also by slight grain flattening (Fig. 6c).

The small folds and localized cleavage are overprinted in places by the regional cleavage, itself weakly defined and incongruent to the syncline. Accordingly, the small folds and their cleavage appear to have preceded the regional deformation. Before rotation in the limb of the syncline, the axial planes of the small folds were inclined gently southeast and their hinges were aligned NE–SW.

In the inner arcs of some of the anticlines the sandstone layers have been detached from the subjacent argillites. Saddle-reefs have thus been produced which were filled by carbonate during or after detachment (Fig. 6b). Evidently the rocks at this stage had sufficient cohesion to deform under the prevailing strain-rates partly in brittle manner. Folds of the sandstone beds are Class 1B, those of the argillites are Class 1C, indicating a marked ductility contrast between the two rock types.

Listric normal faults. In very rare instances striated veins curve away from bedding planes to take on a listric normal fault geometry. Of the veined faults illustrated in Fig. 6(e) that on the left has a demonstrable normal throw of 5–20 cm in different places. The other fault has a throw of more than 5 m but its sense of displacement is unknown because beds cannot be matched across the fault; striations on the vein indicate dip–slip movement. In the other illustrated example (Fig. 5b) one listric fault plane is veined, the other not. Rollover anticlines developed above the faults. The faults shown in the figures strike NE and, where determined, downthrow to the NW. However, insufficient data are available to demonstrate whether or not these are typical attitudes and displacements.

Detachment structures in other parts of the Welsh Basin

Striated bedding-parallel veins are widespread in the Welsh Basin. In the hinterland to Cardigan Bay, for example, there are few major outcrops where they cannot be found. Further afield, we have observed them in various parts of E Wales, at least as far southeast as Llanbadarn Fynydd [SO 099778] (Fig. 1) and what is probably the same type of structure was described from gently dipping but otherwise undeformed strata in the Long Mountain district by Palmer (1972).

Particularly well-displayed examples of striated bedding-parallel veins, accompanied by bedding-normal veins, are exposed in the Wenlock strata of the Llangollen and Denbigh Moors areas of NE Wales (Fig. 1). Described first by Wedd *et al.* (1927, p. 91), the textural features, origin and relations to regional deformation of these structures have more recently been discussed by Nettle (1964), Nicholson (1964, 1966, 1978) and Warren *et al.* (1970).

The veins in Wenlock strata of NE and E Wales differ from those in W Wales in several details. They comprise mainly calcite with minor quartz in contrast to quartz and ferroan dolomite with minor chlorite in W Wales. Several bedding-parallel veins near Llangollen attain greater thicknesses; an example in Pant Glas Quarry [SJ 215476] near World's End is 40 cm thick and is composed of many tens of laminae.

A characteristic feature of the calcite in the veins of NE and E Wales, but not of the ferroan dolomite of W Wales, is its platy habit. Nicholson (1966) has described single crystals less than 1 cm in thickness extending over 70–80 cm^2 in the plane of a vein. Nettle (1964) and Nicholson (1966) have recorded a strong preferred alignment of the *c*-axes of both calcite and quartz normal to vein walls and therefore also normal to the calcite plates and to bedding. No preferred crystallographic orientation has been detected in minerals of the W Wales bedding-parallel veins.

Another difference between the bedding-parallel veins of NE and W Wales is the orientation of the striations. In W Wales the striations are aligned mainly WNW–ESE. Most of those in NE Wales are almost N–S but on different veins or on various laminae in a single composite vein can deviate by as much as 70°.

CHRONOLOGY OF EVENTS

The timing of the structures described in the previous section is established by referring to the burial and regional deformation history of the host rocks. Figure 8 summarizes this history and this section briefly comments on the main stages. Only W Wales is considered here.

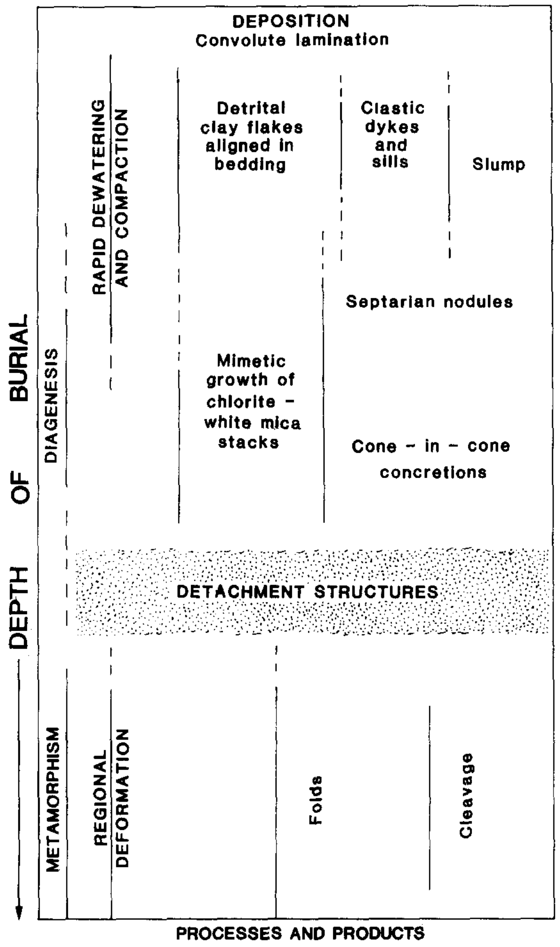

Fig. 8. Detachment structures in relation to depth, processes and products of burial and regional deformation. Sediment was buried to progressively deeper levels through time so that processes and products of burial were diachronous. See text for discussion.

Burial history

Partial dewatering, compaction and local slumping. The earliest structures in this part of the basin developed immediately after sedimentation and others followed after slight burial. They include:

(i) convolute lamination, which developed widely in sandstones almost synchronously with sedimentation;

(ii) the collapse of randomly oriented detrital clay flakes into alignment with bedding; this began soon after deposition and light burial (Maltman 1981, Maltman & Fitches 1982);

(iii) emplacement of clastic dykes and sills, some dykes being buckled by compaction processes;

(iv) development of small intraformational steep, normal and reverse faults (these faults are probably due to down-slope extension of lightly buried sediment as some are enclosed by concretions which themselves grew during compaction);

(v) slump-sheets, containing a variety of small folds, faults and other structures, were generated locally by larger-scale downslope movement of sediment. These slumps have been described from the Llangranog area [SN 310542] (Fig. 1) by Anketell (1963) and Carn Owen [SN 731881] about 20 km ENE of Aberystwyth by Cave (in IGS Report 1966), Tremlett (1982) and Cave & Hains (in press).

Diagenesis. Various concretions are megascopic signatures of diagenesis in the rocks. They comprise: phosphatic layers and lenses near the top of turbidite E intervals (Cave 1979), some of which are deformed and disturbed by bioturbation; septarian nodules; and cone-in-cone layers and nodules. Craig (1985) has inferred that the septarian nodules preceded strong compaction and grew at shallower levels of burial than the cone-in-cone concretions which in places enclose them.

Under the microscope the most conspicuous signs of diagenesis in argillites are the chlorite–white mica stacks which mimetically replaced the bedding-aligned clay fabric before the development of the regional cleavage (Craig *et al.* 1982). Craig (1985) has inferred on evidence from thin-sections that this recrystallization also preceded the growth of cone-in-cone concretions and therefore began under conditions of lighter burial than those concretions.

Chronology and depth of burial. The various features outlined above are related to different stages in a complex continuum of processes that began to operate immediately after sedimentation. The features were controlled by diverse factors during progressive burial and, at deeper levels, by processes perhaps better ascribed to anchimetamorphism. It is important to note that although each stratal unit in the sedimentary pile may now exhibit similar alteration features that developed in the same sequence, the processes causing them were diachronous and were controlled particularly by depth of burial (Fig. 8). Accordingly, diagenetic changes such as concretionary growth and recrystallization of phyllosilicates effectively migrated up through the sedimentary pile with time; or, more accurately, took place as the sediments were progressively more deeply buried and passed down into and through the appropriate depth zone.

There are no accurate constraints on the depths at which these processes began although the concretions possibly offer some information. The rimming of septarian concretions by cone-in-cone layers resembles that described from British Mesozoic shales by Marshall (1982). He inferred from oxygen-isotope data that those rims grew after some tens or probably hundreds of metres. Possibly the Welsh concretions grew at a similar depth range. The growth of chlorite–white mica stacks preceded cone-in-cone development and so seems to have required shallower depths of burial. This conclusion is consistent with observations that diagenetic chlorite can crystallize in very lightly buried sediment (Powers 1957).

Regional deformation

After alteration by the burial processes outlined above the sediments were folded and cleaved during events conveniently ascribed to regional deformation. According to traditional views this deformation occupied a narrow time interval, in the late Silurian, and was more or less contemporaneous across the basin. It is now recognized, however, that like the burial processes the regional deformation was perhaps diachronous and only culminated at that time.

Only the regional deformation structures of W Wales are outlined here as they are better known and their relationships to the detachment structures have been investigated more closely.

Folds are ubiquitous, with wavelengths on all scales from a few millimetres to several kilometres. They have steep axial planes which strike NNE in the north and swing progressively to ENE in the south. The typically complex non-cylindrical geometry of the folds has been described from the Cardigan Bay coast by Price (1962, 1967) and from the hinterland by Jones & Pugh (1915), Tremlett (1982) and Cave & Hains (in press).

Cleavage is found in almost all argillites but normally is poorly developed in arenites. Typically it is a spaced fabric, close in argillites, wide in arenites, comprising anastomosing surfaces on which some pressure solution has taken place and along which very fine phyllosilicates are weakly aligned (Craig *et al.* 1982).

In many localities cleavage displays normal simple fanning relationships to folds. However, there are many examples of complex fold–cleavage relations. Fitches & Johnson (1978), for instance, recorded from the Aberystwyth Grits that cleavage is commonly oblique to axial planes, is locally folded and that both cleavage and folds commonly deviate from regional attitudes. Some of these complications are perhaps due to accommodation in major fold cores (Price 1962), but other explanations involving large-scale detachment and progressive strike–slip deformation are other possibilities under investigation by the authors.

DETACHMENT STRUCTURES IN RELATION TO BURIAL AND REGIONAL DEFORMATION

Almost all the detachment structures of W Wales are demonstrably later than many burial processes and earlier than the regional deformation. Their relationships to burial in NE Wales have not been studied, but Nicholson (1966) concluded that there too the veins preceded the regional deformation.

Relationships to burial processes

Bedding-parallel detachment surfaces. That the bedding-parallel veins developed after several burial changes had taken place in the sediments is established in three main ways. (1) Some cone-in-cone concretions, formed relatively late in the burial history, have short bedding-parallel veins of quartz and carbonate attached to one end. This arrangement is considered to be due to displacement of the concretions by simple shear in the plane of bedding and the concomitant growth of the vein minerals as low-pressure fringes behind the trailing edges of the rigid concretions. (2) Diagenetic pyrite embedded in the hangingwall or footwall of some detachment surfaces has produced parallel grooves with rectangular profiles on adjacent surfaces during shear. (3) In very rare instances concretions have also acted as tools to groove adjacent surfaces (Fig. 7a). The veined faults illustrated in Fig. 6(e), one of which is demonstrably a listric normal fault, are clearly later than the slumping that took place in the host rocks. Slumped layers a few centimetres in thickness which are intercalated with the strata are offset by the faults.

Bedding-normal veins. The bedding-normal vein breccias, and therefore the NW–SE veins cutting them, followed several stages of burial according to evidence from the breccia clasts. Under the microscope the fragments are seen to contain a strong bedding-plane fabric of aligned chlorite–white mica stacks which, by analogy with the fabric-forming sequence in the country rocks, enhanced the earlier compaction fabric. This diagenetic fabric is aligned differently in each clast showing that compaction preceded brecciation and rotation of clasts. At the margins of some clasts the bedding fabric is overprinted by a second fabric which is morphologically similar to, but is itself overprinted by, the regional cleavage. The origin of this fabric at clast margins is unknown. However, the significant point is that it was imposed on diagenetic chlorite–white mica stacks and was also rotated by and preceded the brecciation. On this evidence the vein breccias are later than the growth of diagenetic phyllosilicates.

Relations to regional deformation

Bedding-parallel detachment structures. The bedding-parallel veins and their striations are considered to have preceded the regional deformation for the reasons listed below. The timing of thrusts without veins along them has not been determined, however, because none have been observed at a site where there are small folds belonging to the regional deformation and their relations to cleavage remain equivocal. Some thrusts are probably cogenetic with those with veins but the possibility remains that others are later and related to the regional deformation.

(i) Many bedding-parallel veins have been buckle-folded on a small-scale with their host rocks. They have behaved as relatively rigid layers much like the sandstone beds of the host sequence (Figs. 7b, d & h). Pressure solution seams in carbonate (Fig. 7e) are parallel with the axial planes of, or have a fanning relationship to, the folds and are approximately parallel with cleavage in the host rocks. These small folds and the cleavage are congruent to the regional deformation structures.

(ii) The striations on the veins commonly make high angles with fold hinges. In this way they resemble products of flexural-slip folding. That the striations are earlier than and unrelated to folding, however, can be shown in several ways. There are numerous examples of bed sequences with several bedding-parallel veins which are deformed together by cylindrical coaxial folds. Despite the uniform hinge alignment, the directions of striations on veins in the sequences vary widely, and can make angles of up to 30° with the fold hinges. Indeed, it is common in folds of single composite veins for the striations on successive laminae to be aligned in various directions, almost 90° apart in several instances (Fig. 3d). There are numerous examples of the striations passing over hinges (Fig. 7c). These striations show no change of character in passing from fold limbs over hinges nor with variations in dimensions or tightness of folds; some deeply striated surfaces are only gently warped by small folds in which flexural slip was minimal. These unsystematic and incongruent relationships are considered to show that the striations, and hence the veins containing them, are genetically independent of the small folds.

It might be argued that the veins and striations are due to flexural slip in the limbs of early large-scale folds. Against this interpretation, however, are the observations that the striated veins preceded the small folds which are parasitic to the major folds of the region; in buckling processes parasitic folds develop before the major folds (e.g. Ramberg 1964, Ramsay 1967).

In the absence of argillaceous–arenaceous sequences totally free of tectonic strain in the Welsh Basin, there are no examples of detachment structures which are obviously unrelated to regional deformation. It is pertinent to note, however, that striated bedding-parallel veins, indistinguishable from those described above, do occur elsewhere in essentially undeformed strata. Figure 3(f) illustrates an example from the tilted but otherwise undeformed late Precambrian Stoer Group of NW Scotland. Similar veins have also been recorded from the Lower Palaeozoic rocks of N England (Nicholson 1966), SW Australia (Boulter 1979) and the Alps (Beach 1981). However, the host rocks in each of these areas have been tectonized and the origin of the veins remains equivocal.

Bedding-normal veins. The NE–SW bedding-normal veins have been shown to be closely associated spatially with the bedding-parallel veins and like them are deformed by small folds in places. Few are folded, however, because most are nearly parallel to regional fold axial planes except where rotated in fold limbs (Fig. 7h). In W Wales the strains associated with regional deformation were too low to boudinage these veins although in NE Wales they were deformed in this way (Nicholson 1966).

The NE–SW vein breccias are considered to have preceded the regional deformation for two reasons. First, they are cut by NW–SE bedding-normal veins which are themselves deformed by small folds of the regional deformation (Fig. 4b). Second, the late fabric developed in the margins of some clasts, where suitably oriented, is crenulated by the regional cleavage. The NW–SE bedding-normal veins and associated tension gashes were deformed by small folds in places and have acted as rigid struts which caused the cleavage of the regional deformation to deflect around them (Fig. 4f).

ORIGIN OF THE DETACHMENT STRUCTURES

Figure 2 is a model showing the geometrical interrelationships of the various detachment structures. The model is applicable on various scales and the structures are considered to have developed almost synchronously at several sites at similar levels in the sediment pile. Through time the sediment was progressively buried to depths appropriate for each type of deformation so that the structures propagated up through successively younger strata.

It is considered that overpressured pore fluids were the prime cause of the detachment structures; overpressuring caused brittle-failure and the pore fluids deposited the vein minerals. The role of the pore fluids in producing the various structures is conveniently discussed under five headings: depth of development; fracturing processes; slip processes; directions of slip and mineral deposition.

Depth of development

There is little information on the depth of formation. The structures developed at levels deeper than those required for the several diagenetic changes which had preceded them. In particular, the detachment structures developed after the cone-in-cone concretions and hence at deeper levels. It was concluded above that those concretions, produced at levels deeper than other diagenetic changes recorded in the rocks, probably grew at depths of a few tens or few hundreds of metres. The depths at which the later regional deformation took place are not known and hence no maximum depth can be inferred.

Fracturing processes

It is considered that the flow of pore fluids migrating up through the sediment pile was impeded by beds of low permeability and became ponded beneath them in the manner predicted by Price (in Fyfe *et al.* 1978). In the argillites compaction had previously induced a preferred bedding alignment of detrital clays which had been enhanced by mimetic diagenetic recrystallization thus creating a barrier to upward flow. Fluid pressure increased with burial and became sufficient to cause fractures parallel to bedding, exploiting the low tensile strength of the planar anisotropy. Having jacked open bedding-parallel fissures the fluids deposited minerals on their free surfaces, beginning the process of vein formation.

Flexing of the roof of a fissure as it opens up under fluid pressure can produce stress conditions suitable for vertical hydraulic fracturing (Fyfe *et al.* 1978, figs. 11.17 and 11.18). The NE–SW bedding-normal veins and breccias are attributed to this process. If fluids were released up the steep fissures and fracture zones in this way, they could pond beneath higher-level impermeable layers whilst the bedding-parallel fissures would collapse abruptly. In consequence, a stack of collapsed but commonly mineralized bedding-parallel fissures could build up, interconnected by steep veins and vein-breccias.

The bedding-parallel veins composed of several laminae are regarded as products of fluid pumping. Each of several fluid pressure pulses opened the bedding-parallel fissure and deposited a mineral veneer on its roof and floor before being expelled upward through steep fractures, leaving the bedding fissure to collapse. Each fluid pressure pulse could have produced two mineral laminae. This repeated use of one bedding-parallel fissure is probably because once a thin vein had been deposited it too provided both a low permeability layer capable of ponding fluid beneath it and surfaces of low tensile strength vulnerable to further jacking.

The above model considers a sediment pile in which bedding is horizontal. It is likely, however, that bedding was inclined, due to differential compaction, original depositional slopes or rotation by contemporaneous faulting, for example. The model outlined by Price (1977) and Mandl & Crans (1981) can be invoked in which inclined bedding-parallel fractures, again caused by high fluid pressures, became potential detachment surfaces and soles to gravity-glide sheets. Shear displacement would have been inhibited until the build-up of fluid pressure and tensile stresses in the superjacent sheet combined to produce steep extensional hydraulic fractures (NE–SW veins and breccias) above the upslope tip of the bedding-parallel fracture. Development of the SE-dipping steep fractures enabled the detached sheet to glide down slope, provided the movement could be accommodated by various compressional structures towards and at the toe of the sheet and by lateral detachment at its side (cf. Farrell 1984).

Slip processes

It is envisaged that the bedding-parallel and NE–SW bedding-normal veins originated by one, or possibly both, of the mechanisms described. A detached sheet was inclined at the time of its inception according to the second model or at a later stage if generated according to the first model. In the latter case, tilting by faulting or perhaps the earlier stages of regional deformation are invoked. Once the sole and trailing edge of the sheet became detached, its slide downslope would produce various other structures. Compressional brittle structures in toe areas are represented by the thrusts, backthrusts and imbricate fans whilst more ductile behaviour, perhaps due to lower strain rates, is shown locally by the small asymmetric folds and their cleavage. The NW–SE bedding-normal veins are considered to be products of extension normal to the NW–SE compression in toes, whilst the tension-gash arrays marked incipient lateral ramps to the sheets. Extension detachment structures, marking the trailing edges of glide sheets, are apparently much less common than other detachment structures. Possibly they have been overlooked or are hidden in the very poorly exposed ground behind the coast.

The striations are the visible record of slip on the soles and are considered to be the result of the following processes. The development of bedding-normal hydraulic fractures led to abrupt collapse of the bedding-parallel fissures. In consequence the mineral veneers on the roof and floor of a fissure impinged on each other across an irregular surface defined by crystal faces that grew into the once-open cavity. Slip on this surface due to gravity gliding and possibly facilitated by residual fluid pressure, reduced from its former high value (but nevertheless still elevated), caused the harder minerals of the vein, notably quartz, to striate the softer minerals, especially the carbonates.

The rare growth of mineral fibres aligned with the striations of some bedding-parallel veins was probably due to unusually low slip rates giving time for tensile strain crystallization. The local development of cataclastic textures in quartz probably indicates rapid slip rates. Some of the twinning in calcite, so characteristic of the veins in NE and E Wales, is also tentatively attributed to deformation during gliding.

Slip directions

In W Wales slip directions inferred from striations on bedding-parallel veins (see above) and the very few orientations available from thrusts and associated structures are mainly WNW and ESE, but there are insufficient data to decide which was the dominant direction. The slip directions can differ widely, however, even in single composite veins (Fig. 3d). This variation is considered to mark changes in slope direction; in the composite veins, the changes took place over the period in which the vein developed. The inclination of slip surfaces was not necessarily more than a few degrees, provided that residual fluid pressure remained high, and only a slight rotation of the strata was required to change the slip direction by a large angle.

Mineral deposition

No attempt is made to analyse all the factors which controlled mineral deposition but two aspects are considered. One is the conspicuous contrast between the bedding-parallel veins of NE and W Wales; calcite is found in the former and ferroan dolomite in the latter. This contrast relates to the difference in composition of the host rocks in the two areas. The Wenlock host rocks in NE Wales commonly contain abundant calcite in the matrix and cement, whilst the arenites of W Wales commonly have a dolomitic matrix (Cave 1979). It appears, therefore, that the pore fluids causing the veins had acquired their major chemical components locally.

The other aspect concerns the platy habit of calcite in the bedding-parallel veins of NE and E Wales and the alignment of the calcite *c*-axes normal to the plates and to vein walls. It is usual for calcite to grow with *c*-axes normal to the walls of extensional fissures so the lattice arrangement here is perhaps a reflection of a primary growth orientation. It is suggested that calcite first grew as elongate crystals into a fluid-filled cavity or in fibrous form as in 'beef' layers produced in overpressured shales (Stoneley 1983). Subsequently, during a later stage of diagenesis perhaps, they recrystallized into single-crystal plates which retained the original lattice orientation.

SUMMARY AND DISCUSSION

The commonest detachment structures, bedding-parallel and bedding-normal veins, were direct responses to overpressuring by pore-fluids. They developed in horizontal or inclined strata once the host sediment had been buried beyond a depth of perhaps a few hundred metres. Allied structures were produced during slip on inclined bedding-parallel veins which acted as soles to sheets gliding downslope. Movement on the soles is marked by striations. Compressional structures, mainly thrusts, developed in the toe regions of the gliding sheets and rare extensional structures at the trailing edges are represented by listric normal faults. The slip (and slope) directions were variable but mostly WNW and ESE in W Wales and N or S in NE Wales. Displacement on each sole was on a centimetric to metric scale but where soles are abundant the aggregate displacement was probably one or two orders of magnitude greater.

Why the host strata were inclined during or after the development of the bedding-parallel veins has not been established. Only a slight tilt was necessary to allow slip, provided fluid-pressure on the soles was high. In a subsiding basin there are numerous ways of gently tilting strata, especially if sedimentation was accompanied by faulting.

The structures mark an important but previously undocumented stage in the burial history of the sediments and of pore-fluid activity in the Welsh Basin. From theoretical considerations, similar structures are to be expected in most sedimentary basins. That they have not been extensively reported is perhaps because individual structures are small, seemingly insignificant, and are easily overlooked whilst in strongly deformed strata it is likely that they will be assumed to be products of folding or faulting.

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