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Marine to fresh water: the sedimentology of the interrupted environmental transition (Ludlow–Siegenian) in the Anglo-Welsh region

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The transition occurred in a period of approximately 15–20 Ma, in broad association with the closure of the Iapetus Ocean, and as thick marine sediments in the Welsh Basin (?small fore-arc basin) became welded on to the Midland Microcraton and East Anglian Foldbelt to the east, with a consequent inversion of relief.

The earlier Ludlow sediments are sharply differentiated between the basin, where deep-water turbidites accumulated, and the microcraton, on which calcareous shallow-marine deposits formed. Slope facies, in association with canyons, mark the basin margins. Environmental differences became increasingly less marked as the Ludlow advanced, but the distinction between basin and microcraton was never entirely eradicated. By the end of the Ludlow, shallow and restricted marine conditions prevailed, but only in central Wales was deposition apparently continuous into the Downton. The earliest Downton sediments, in the areas to the south and east, give evidence of either non-deposition or the temporary withdrawal of the sea. Later Downton sediments spread widely but environmentally are somewhat enigmatic. Except in southwest Wales, where a valley had been alluviated before being transgressed, they point to the replacement of a shallow-marine shoal and barrier complex by extensive and uniform coastal mudflats, influenced for a substantial period by both rivers and the sea.

The distant northerly complex of regionally metamorphosed rocks which furnished the Downton sediments became isolated from the Anglo-Welsh area in the early Gedinnian, as deformation in the region of the Irish Sea brought Lower Palaeozoic (including Downton) rocks of the Welsh Basin into the zone of weathering. A major consequence of this shift of sediment source and rearrangement of drainage, marked by an extensive spread of unusually well developed and closely spaced palaeosols, was the sudden appearance and subsequent rapid southward advance of wholly fluvial environments. Under the continued pressure of deformation nearby to the northwest, comparatively stable and frequently meandering streams (Gedinnian) were replaced by larger and more unstable sand-bed rivers of low sinuosity (Siegenian–Emsian).

1. Introduction

The purpose of this paper is to outline how sedimentary environments evolved in the Anglo-Welsh area during the transition from the deep-water marine conditions that had dominated Lower Palaeozoic times to the continental, fresh water conditions so important during the Upper Palaeozoic. Evolution occurred during, and was largely impelled by, the gradual closure of the Iapetus Ocean and the eventual suturing of the lithospheric plates that lay to either side, both of which find representation in the British Isles (Phillips *et al.* 1976). Similar transitional conditions were in fact widespread at about this time in many parts of the North Atlantic region, notably in Norway, Spitsbergen and eastern and northern Canada. The Anglo-Welsh area is singled out for detailed discussion, however, because it is particularly well known and

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contributes more than any other region to our understanding of the evolution of the early plants and the fresh water vertebrates.

Two points should be kept in mind in following the environmental evolution of the Anglo-Welsh area. Not only was there a rapid replacement of one kind of depositional environment by another, but the erosional environments – the source-lands that yielded sedimentary materials – also experienced major changes. The geological and physiographic evolution of long-standing source areas was continued, as erosion bit deeper and the pace of tectonism varied, and important new source-lands became established. Thus the evolving flora and fauna conceivably encountered an increasing development and diversification of continental environments, ranging from marginal-marine and riverine lowlands to upland habitats. The second point is that the replacement of marine by fresh water conditions took place neither continuously nor at a uniform pace. In the Anglo-Welsh area there were at least two periods of unusually rapid and pronounced change, in one case with a reversal of sense, making the transition an interrupted or punctuated one.

The period of time considered in this paper is from the Ludlow (late Silurian), when the youngest Lower Palaeozoic turbidites known in the Anglo-Welsh area were deposited, to the Siegenian–Emsian (early Devonian), by which time south-flowing rivers had become firmly and widely established. Taking the main time-points of Gale *et al.* (1980), the time scale suggested by the work of Ross *et al.* (1982) and Wyborn *et al.* (1982), and the *pro rata* proposals of McKerrow *et al.* (1980) (but see Gale & Beckinsale 1983), the Ludlow may represent about 8 Ma, the Downton 2 Ma, the Gedinnian 6.25 Ma, the Siegenian 4 Ma, and the Emsian 6.25 Ma. Thus the transition was complete in a period of the order of 15–20 Ma. The rocks involved are those of the traditional marine Silurian and the Lower Old Red Sandstone red bed facies (Downton–Emsian). It would be out of place here to consider even in outline their complex and still partly debatable stratigraphy, but recent comprehensive reviews are available for both the Silurian (Holland *et al.* 1963; Cocks *et al.* 1971; Holland *et al.* 1980; Bassett *et al.* 1982) and the Lower Old Red Sandstone at outcrop and in borings (House *et al.* 1977; Allen 1979).

2. Structural foundations

Five main elements contribute to the structural foundations of the area that experienced the marine to fresh water transition (see figure 1). Already a long-standing positive feature, the Irish Sea Ridge carried thin, broken, onlapping sequences, and may have been bounded on each flank by a staircase of down-faulted blocks (Shackleton 1954; Crimes & Crossley 1968; James & James 1969; Brück *et al.* 1979). The Welsh Basin, thought to be a fore-arc basin (Okada & Smith 1980), contained several kilometres of marine clastics and volcanics, including thick turbidite formations ranging between Cambrian and Silurian in age (Jones 1956). Closing off the basin to the S and SW, and with a northern flank marked by thinning and non-sequence (George 1970), lay a mildly positive feature which Cummins (1969) called St David's Land (see also Stamp 1923). The Midland Microcraton E of the basin carried thin and broken mixed clastic–carbonate sequences (Wills 1951), the Ordovician being represented only by igneous intrusions.

The most shadowy of the five elements remains the East Anglian Foldbelt, first recognized by Wills (1951) and later claimed to be an aulacogen (Evans & Maroof 1976; Evans 1979). This broad concealed zone of NW–SE trend (see figure 1) is characterized by high heat flow

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(Richardson & Oxburgh 1978, 1979) and by two belts of bold magnetic anomalies. Those of the southwestern belt, ranging along a line trending roughly through Nottingham and Cambridge, are comparatively small and steep-sided. They appear to be caused by granite plutons, to judge from their association with such rocks at depth in the Warboys (Le Bas 1972; Cribb 1975) and Kirby Lane (Evans & Maroof 1976) borings and with exposed late Ordovician to early Silurian granodiorites at Mountsorrel (Cribb 1975). The northeastern anomalies, trending between Lincoln and Norwich, are generally broader and less steep. There is little direct evidence as to their cause, but an association with intrusive granites is not impossible (Chroston & Sola 1982).

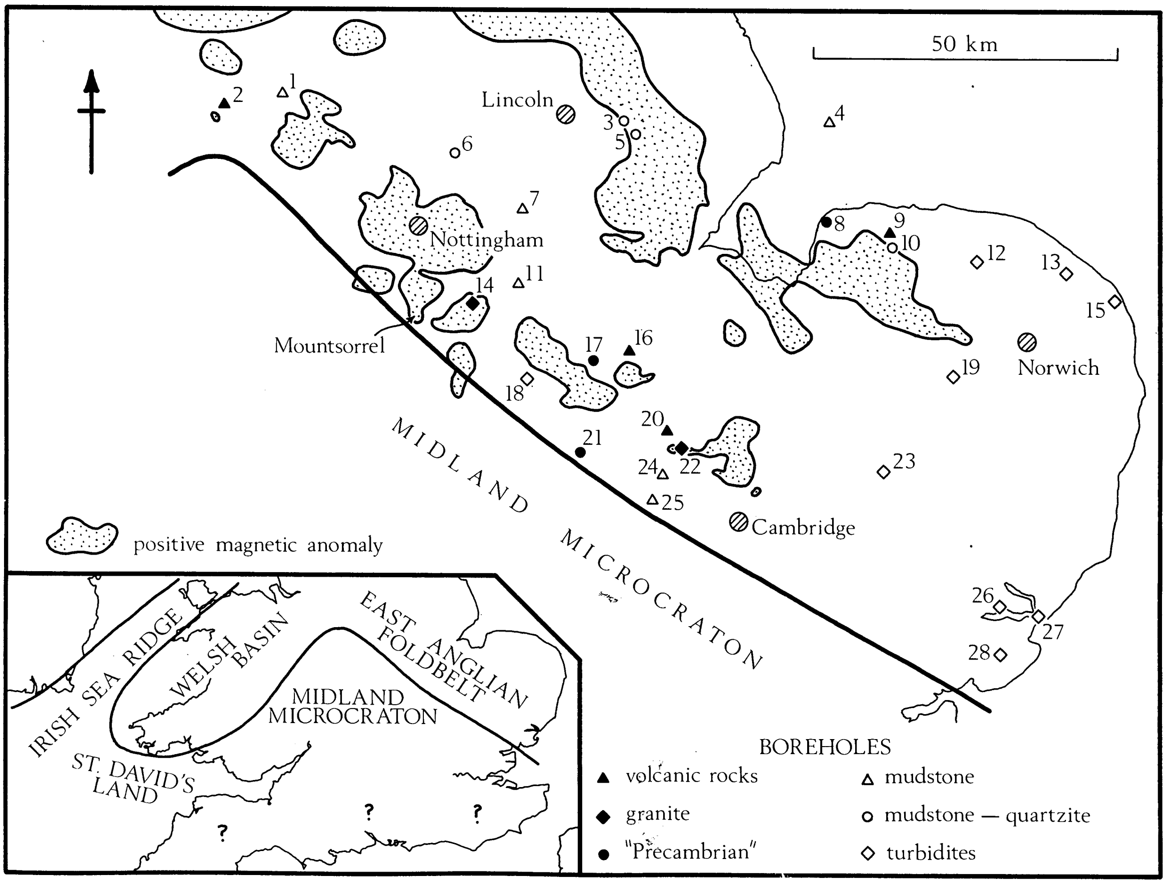

Figure 1. Structural elements (inset map) relevant to the Siluro-Devonian marine-freshwater transition in the southern British Isles, together with details of deep boreholes and magnetic anomalies in the East Anglian Foldbelt. Magnetic anomalies drawn mostly at the +50 gamma contour, after the Geological Survey's (1965) *Aeromagnetic Map of Great Britain*. Boreholes: 1, Eyam; 2, Woo Dale; 3, Nocton; 4, Burmah Oil; 5, Stixwould; 6, Eakring; 7, Foston; 8, Hunstanton; 9, North Creake; 10, South Creake; 11, Sproxton; 12, Saxthorpe; 13, East Ruston; 14, Kirby Lane; 15, Somerton; 16, Glinton; 17, West Wittering; 18, Thorpe-by-water; 19, Ellingham; 20, Upwood; 21, Gas Council; 22, Warboys; 23, Culford; 24, Huntingdon; 25, Great Paxton; 26, Stutton; 27, Harwich.

The granites appear to have been intruded into a complex of sediments and volcanic rocks, steeply dipping and in many places phyllitic or cleaved, which range definitely up to Llanvirn in age. Pyroclastic rocks occur in borings at Upwood (Kellaway 1966), Glinton (Kent 1962) and North Creake (Kent 1947), and at Woo Dale (Cope 1973; Evans 1979) accompany lavas

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and Llanvirn marine fossils. A facies of mudstones to phyllitic shales is known from borings at Eyam (Dunham 1973), Foston (Lees & Taitt 1946), Sproxton (Lees & Taitt 1946), Huntingdon (Kellaway 1967) and Great Paxton (Jenkins 1983). An Arenig–Llanvirn shelly fauna is present at Eyam. The graptolites in the rich early Llanvirn shelly–graptolite fauna at Great Paxton have Baltic and East European affinities. Rocks attributed to the Llanvirn also occur in the Burmah Oil well (Wills 1978). At Nocton (Lees & Taitt 1946), Stixwould (Kent 1968), Eakring (Lees & Taitt 1946) and South Creake (Chroston & Sola 1982) occurs a facies of thick mudstones and quartzites. Wills (1978) identified 'Precambrian' rocks at Hunstanton, West Wittering and in a Gas Council well, all within the general area of the boreholes already mentioned. What may be a turbidite (or shelf-storm) facies is reported from Thorpe-by-Water, yielding a Llanvirn whole-rock radiometric age (Richardson & Oxburgh 1978).

A strongly deformed turbidite facies is widespread over the southeastern part of the East Anglian Foldbelt (see figure 1), occurring at Saxthorpe (Chroston & Sola 1982), East Ruston (Chroston & Sola 1982), Somerton (Chroston & Sola 1982), Ellingham (Chroston & Sola 1982), Culford (Strahan 1913), Harwich (Bullard *et al.* 1940), and at Weeley and Stutton (Bullard *et al.* 1940) where early Silurian fossils are recorded (Cocks *et al.* 1971).

The sediments and volcanic rocks of the Eastern Anglian Foldbelt may therefore have been welded on to the Midland Microcraton and intruded by granite plutons in late Ordovician to early Silurian times. Their present distribution suggests that the foldbelt may comprise complex structures plunging to the SE. In view of the suggested age of the plutonism (?folding a little earlier), it is significant that the Midland Microcraton should lack Ordovician and earliest Silurian sediments.

3. Ludlow marine basin and shelf

The Ludlow saw in the Welsh Basin and Midland Microcraton the gradual replacement of sharply differentiated marine basin, slope and shelf sediments by widespread deposits of a uniformly shallow-marine aspect (Holland & Lawson 1963). In thickness terms, there is almost an order of magnitude difference between the basin sequence and that on the microcraton (see figure 2*a*). Moreover, the succession on the microcraton is probably less complete than in the basin, in view of its numerous either condensed or non-sequential horizons rich in vertebrate remains and locally phosphatic and pebbly (King & Lewis 1912; Lawson 1954, 1956; Squirrell 1958; Squirrell & Tucker 1960; Cave & White 1971; Turner 1973). These horizons probably record episodes of at least shoaling on the microcraton, if not actually regressive–transgressive events.

Figure 2*a* emphasizes the pattern of sedimentation in the early Ludlow, when the sharpest facies differences are apparent. Both margins of the Welsh Basin can be glimpsed. A shelf may have existed on the flanks of the Irish Sea Ridge to the N and NW, to judge from the shelly mudstones transported southward by submarine slides into the Denbigh branch of the basin's axial trough (Jones 1937, 1939). This stem was separated by a submerged ridge – the Derwen Ridge of Cummins (1969) – from the Montgomery branch to the south. Slides moving toward respectively the SE and NW occurred on the eastern flank of the Derwen Ridge and on the opposite slope between the Midland Microcraton and the Welsh Basin (Bailey 1964, 1969; Woodcock 1976). Near the shelf-edge at Ludlow, features interpreted as submarine-canyon

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heads lead westward downslope (Whitaker 1960). A thick turbidite facies of siltstones and very fine-grained sandstones occupies the bottoms of the troughs (Cummins 1959*a*, *b*; Bailey 1964). The turbidites interfinger with the slides in some parts of the Welsh Basin but in other places underlie them. Graptolitic shales, mudstones and siltstones at times spread over the basin slope and floor.

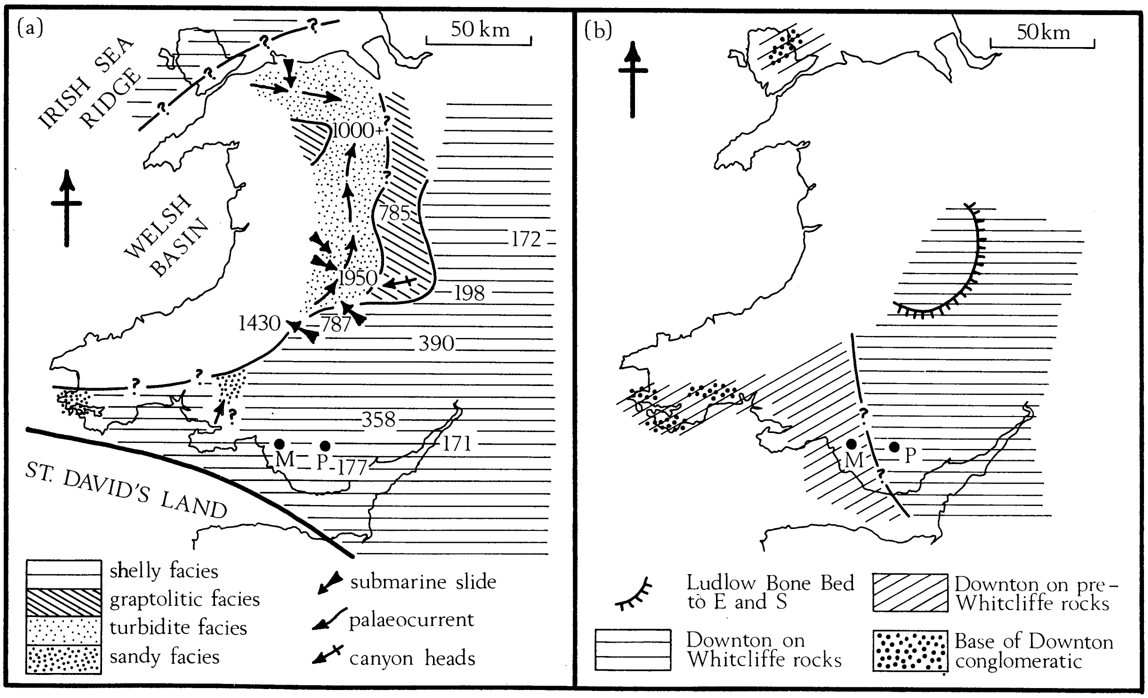

FIGURE 2. Ludlow sedimentation (early Ludlow emphasized) and the relationships of the basal Old Red Sandstone in Wales and the Welsh Borders. (*a*) Ludlow facies distribution and palaeocurrents (partly after Holland & Lawson (1963) and Cummins (1969)). Figures show the thickness of preserved Ludlow strata in metres. (*b*) Distribution of the Ludlow Bone Bed and basal relationships of the Lower Old Red Sandstone. Boreholes: M, Maesteg; P, Pontypridd.

The shelf facies, thinning eastward and southward over the microcraton (see figure 2*a*), mainly comprise richly fossiliferous, calcareous shales, mudstones and siltstones, with some argillaceous biogenic limestones (King & Lewis 1912; Ball 1951; Lawson 1954, 1956; Walmsley 1959; Squirrell & Tucker 1960; Holland *et al.* 1963; Cave & White 1971). The possible presence of a low-lying St David's Land from which terrigenous clastic sediments might have been supplied northward is suggested by the (?)deltaic sandstones and conglomerates of the Trichrûg Beds near Llandeilo (Squirrell & White 1978) and the perhaps Ludlow (Allen & Williams 1978) deltaic or shallow-marine quartzites of SW Dyfed. The breadth of the shelf on this northern margin of St David's Land is emphasized by the report of Ludlow rocks in the Maesteg and Pontypridd borings (Wills 1978).

By late Ludlow times a facies of calcareous shelly siltstones had spread widely over both microcraton and basin (Holland & Lawson 1963). Although basin and shelf are no longer recognizable in depth terms (there is a continuing structurally controlled deposition-rate difference), sediment transport directions tend to parallel the former shelf-edge (Bailey & Rees

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1973). The well-sorted siltstones contain frequent shell lags, and locally display patterns of storm-related planar to hummocky lamination, cross-lamination and wave-current ripples (for example, see Goldring 1966).

4. End-Ludlow movements

Although numerous either condensed or non-sequential horizons can be recognized in the Ludlow of the Midland Microcraton, none introduced a radical change in sedimentary style away from the predominance of calcareous muds. At the end of the Ludlow, however, there occurred a widespread shrinking of the sea that had occupied the Welsh Basin and its surroundings, and a change to new styles of sedimentation. This significant event is marked by the Ludlow Bone Bed, a widespread and conspicuous condensed bone bed horizon (or complex of bone beds), more likely than any similar earlier horizon to record regression followed by transgression (Allen & Tarlo 1963; Antia & Whitaker 1978; Antia 1979*a*, *b*, 1980).

The Ludlow Bone Bed (see figure 2*b*) is well known on the microcraton in the Ludlow–Much Wenlock area (Robertson 1927; Whitaker 1960), the central and west Midlands (Stamp 1923; Ball 1951), the Malvern–Abberley Hills (Phipps & Reeve 1967), Woolhope (Squirrell & Tucker 1969), May Hill and Gorsley (Lawson 1954, 1956), Tite's Point (Cave & White 1971), Usk (Walmsley 1959) and Cardiff (Waters & White 1980). It resembles the older condensed horizons lithologically and rests on a planar to slightly uneven scoured surface of older deposits that in places had already become lithified and bored. Channels are associated with the bone bed at Tite's Point and at a locality in the Usk area.

Westward from the Welsh Borders and SE Wales (see figure 2*b*), where the Ludlow Bone Bed rests everywhere on Whitcliffian rocks (highest substage of the Ludlow), the corresponding basal measures (locally conglomerates or breccias) of the Lower Old Red Sandstone facies step across successively older and progressively more deformed rocks (Straw 1930; Potter & Price 1965; Squirrell & White 1978), until in SW Wales the Precambrian is locally reached (Strahan *et al.* 1907, 1909, 1914; Cantrill *et al.* 1916). Much of this southwestern area therefore may briefly have been land toward the end of the Ludlow and in the early Downton. There is no compelling factual basis (Allen & Williams 1978; Squirrell & White 1978) for the recently stated view that the earlier Lower Old Red Sandstone rocks in SW Wales are Wenlock to Ludlow in age (Sanzen-Baker 1972; Hurst *et al.* 1978).

Deposition seems to have been continuous from Ludlow into Downton times only in the heart of the Welsh Basin (see figure 2*b*), there being no proven Ludlow Bone Bed in the Clun (Stamp 1919; Earp 1940), Kerry (Earp 1938) and Knighton (Holland 1959) outcrops. The Ludlow Bone Bed is doubtfully present at Long Mountain to the north (Stamp 1923; Austin 1925), also lying off the edge of the Midland Microcraton. The Irish Sea Ridge to the NW was probably land for much of the Ludlow and Downton. Here in Anglesey a coarse fluvial facies of late Downton age rests with marked angular unconformity on folded rocks no younger than early Silurian (Greenly 1919; Allen 1965).

By the end of Ludlow times the sea had shrunk to a very small central area. Although a short-lived transgression followed, mainly toward the S and SE, the previous conditions of shallow-marine deposition were never to return, as the succeeding Lower Old Red Sandstone facies demonstrate.

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5. Early Downton alluviation in SW Wales

To judge from recent developments in stratigraphical knowledge, early Downton times saw the alluviation of a broad shallow valley on the northern margin of St David's Land (see figure 3). The stratigraphical key to this feature and event is afforded by the variously named *Lingula*-bearing intertidal facies (for example, Temeside Shales) which occur widely in the Welsh Borders and S Wales, in many areas directly overlying sandstones of Downton Castle type and equivalence.

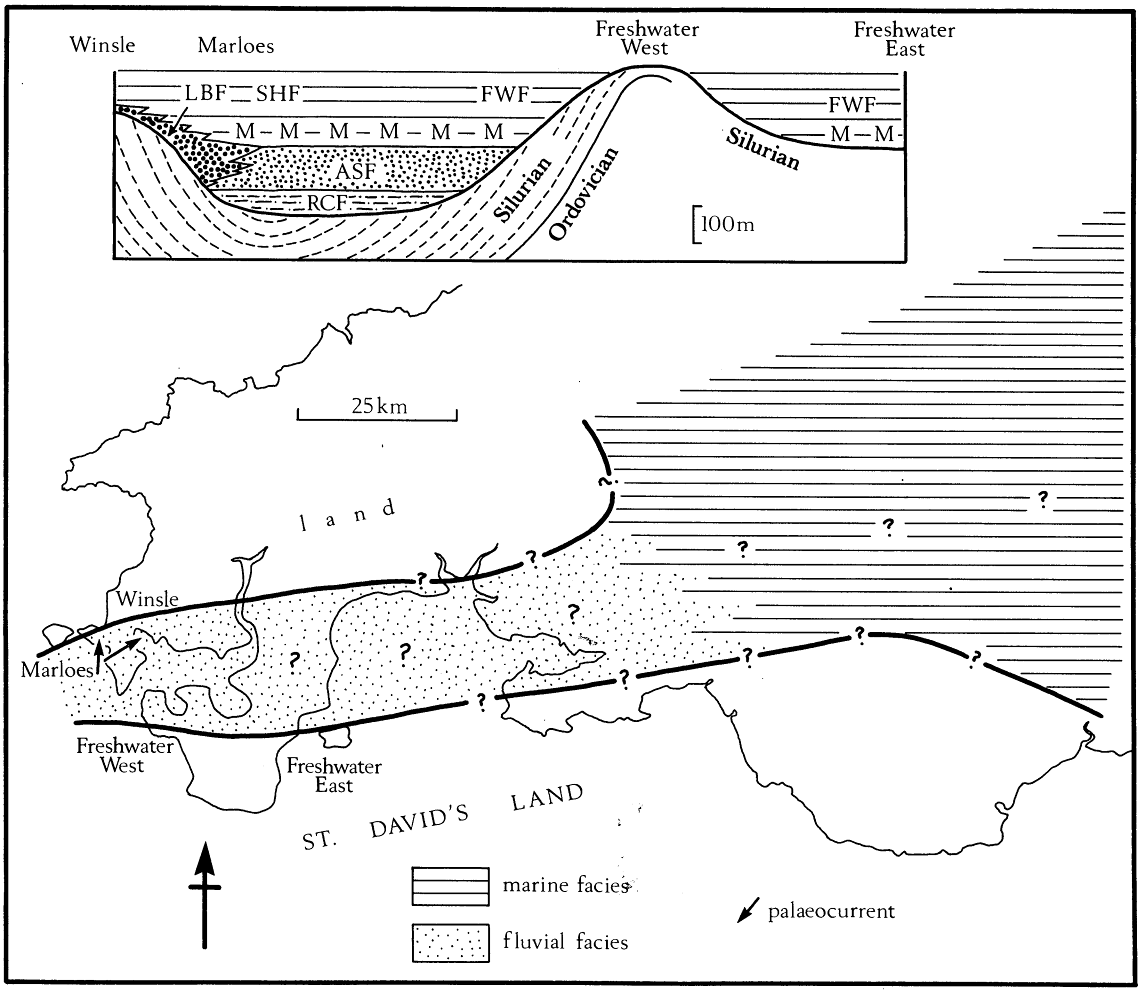

Figure 3. Tentative palinspastic reconstruction of the early Downton palaeovalley in SW Wales. The inset vertical section connects the localities named on the map. M, *Lingula* facies; RCF, Red Cliff Formation; ASF, Albion Sands Formation; LBF, Lindsway Bay Formation; SHF, Sandy Haven Formation; FWF, Freshwater West Formation.

The evidence for the valley is clearest in SW Dyfed, which has been resketched palinspastically in accordance with independent structural (Hancock 1973; Hancock *et al.* 1981, 1983) and palaeomagnetic (McClelland Brown 1983) evidence suggesting a substantial clockwise rotation of the region, in association with both fold- and fault-related shortening. The sequence below the intertidal facies is thickest (*ca.* 150 m) and most complete in the Marloes area (Allen &

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Williams 1978), where a mud-dominated fluvial facies (Red Cliff Formation), derived from the S, is overlain by pebbly braided-stream alluvium (Albion Sands Formation), transported from the SW. Interfingering (perhaps from the N) with the Albion Sands and overlying Sandy Haven formations is an alluvial-fan conglomerate (Lindsway Bay Formation) rich in the debris of felsic volcanic rocks. At Winsle, a short distance northward from Marloes, only a thin development of the Lindsway Bay Formation is recognizable, resting with a sharp and clear break on Wenlock marine shales (Allen *et al.* 1976; Allen & Williams 1978). At Freshwater West, to the south of Milford Haven, the oldest Lower Old Red Sandstone rocks also rest sharply on the Wenlock, with Llanvirn beds lying close below, but substantially postdate the intertidal facies. The latter is developed again at Freshwater East (Dixon 1921) but without an underlying fluvial sequence. The opposite side of the ancient valley is ill-defined E of Winsle. The Lower Old Red Sandstone rests on Precambrian rocks not far to the N and is red and conglomeratic (Strahan *et al.* 1914). The basal Green Beds come on further E and eventually are found seemingly above the Tilestones (Strahan *et al.* 1907, 1914; Squirrell & White 1978), a correlative of the Downton Castle Sandstone. The eastward extension of the valley to meet in central S Wales the southern edge of the marine area is strongly suggested by the presence in the Tilestones of felsic igneous debris similar to that plentiful in both the Albion Sands and Lindsway Bay formations.

The margin of St David's Land drawn in SE Wales is rather speculative. Whereas the Downton Castle Sandstone has a correlative in the Usk inlier (Walmsley 1959), a facies of this type is lacking to the S at Cardiff (Waters & White 1980). The lowermost beds of the Lower Old Red Sandstone may therefore young southward.

6. Early Downton sand shoals and later mudflats

As the valley in SW Wales became infilled, sand shoals built up extensively in the shallow seas that ranged across the borders of the now virtually unrecognizable Midland Microcraton and Welsh Basin (see figure 4). These shoals are recorded by the Downton Castle Sandstone of the Welsh Borders (Allen 1974*a*; Antia 1979*a*, 1980), the Tilestones of central S Wales (Potter & Price 1965; Squirrell & White 1978), the *Platyschisma helicites* Beds and Yellow Downton of central Wales (Stamp 1919; Austin 1925; Earp 1938, 1940; Holland 1959), and variously named correlative formations elsewhere in the region (Ball 1951; Lawson 1954, 1956; Walmsley 1959; Squirrell & Tucker 1960; Phipps & Reeve 1967; Cave & White 1968, 1971).

Two main facies are recognizable in these early Downton deposits. The lowermost, thickest in central Wales (Montgomery trough) and confined to a central area, comprises an upward-coarsening alternation of mudstones, clean coarse siltstones and very fine to fine-grained sandstones. The siltstones and sandstones occur as thin, erosively based, normally graded units with sharp current- or wave-rippled tops. Cross-lamination and a planar to wavy horizontal lamination occur internally. The mudstones, and to a less extent the sandstones, are bioturbated and yield a locally abundant but somewhat restricted shallow-marine invertebrate fauna with drifted plants (Antia 1980). Vertebrate remains are few and chiefly disarticulated (Allen & Tarlo 1963). The facies is probably subtidal (Allen 1974*a*), as Antia's (1980) record of fine-scale desiccation cracking in the mudstones could not be confirmed. The uppermost and much thicker facies comprises well-sorted very fine to fine grained sandstones with a variety of sedimentary structures indicative of vigorous wave-action in the intertidal and inshore zones

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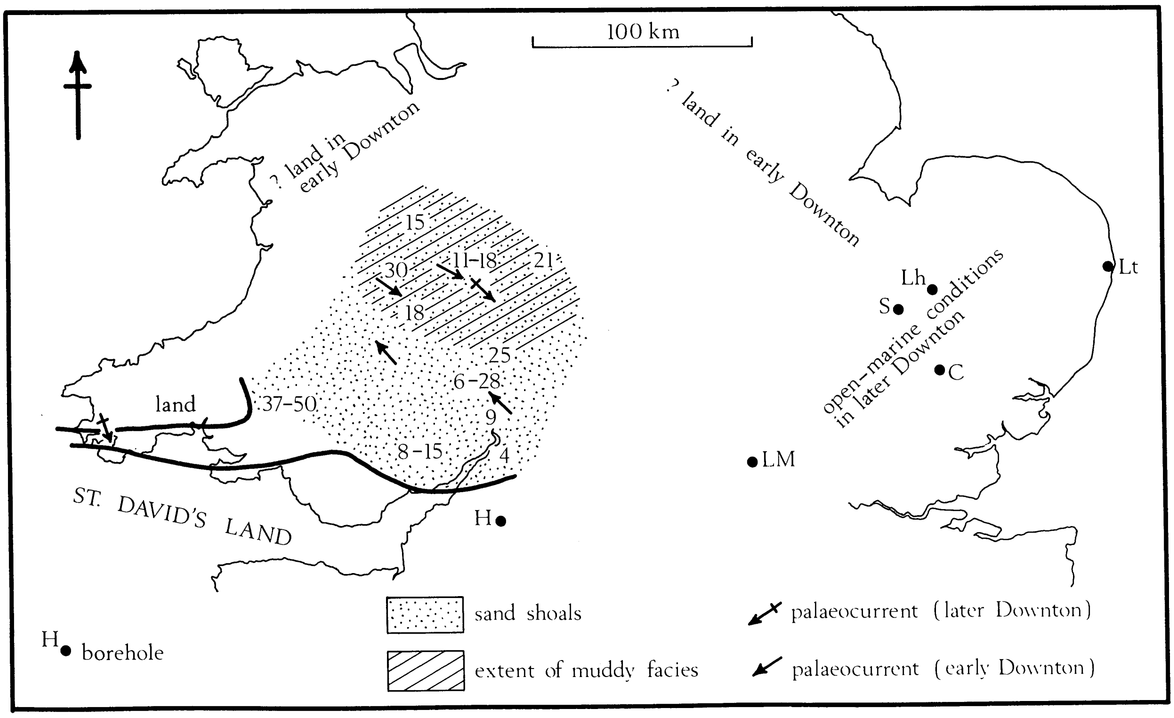

FIGURE 4. Downton sedimentation in southern Britain. Figures show the total thickness (muddy plus sandy facies) of the Downton Castle Sandstone and correlative sand shoals (early Downton). Boreholes: H, Hamswell; LM, Little Missenden; S, Soham; Lh, Lakenheath; C, Clare; Lt, Lowestoft. Palaeocurrents after Bailey & Rees (1973) and Allen (1974*a*).

(Allen 1974*a*). Locally, desiccation cracks in mud partings are preserved beneath sandstone beds (Allen 1971). Invertebrate fossils are scarce and mainly drifted, except in the Tilestones, which yields a rich and varied partly transported molluscan fauna (Squirrell & White 1978). Articulated vertebrates have occasionally been found but these animals are chiefly represented by fragmentary remains (Allen & Tarlo 1963). Diverse sediment-transport directions are revealed by the Downton Castle Sandstone and its correlatives (Bailey & Rees 1973; Allen 1974*a*), which may point to more than one provenance. The Tilestones of central S Wales, for example, appear to have been partly derived from the W, to judge from their inclusion of felsic igneous debris like that found in the basal (fluvial) Lower Old Red Sandstone of SW Wales. These sand-shoal deposits occupy a minimum area of about 12000 km^2, and therefore somewhat exceed in scale the tidal shoals and islands of the Danish Friesian system, as well as those of the Martha's Vineyard–Nantucket complex of the northeastern U.S.A.

The remaining several hundred metres of mudstones with minor sandstones attributable to the Downton are well known only in the Welsh Borders (Robertson 1927; Allen 1974*a*), SE Wales (Welch & Trotter 1961; Squirrell & Downing 1969; Allen & Dineley 1976), and SW Wales (Dixon 1921; Allen & Williams 1978, 1981, 1982). The lowermost of these deposits (for example, Temeside Shales) are clearly intertidal in origin and yield vascular plants (Edwards 1970*a*, 1979) and a highly restricted marine invertebrate fauna. Vertebrate remains, locally plentiful in the Downton rocks, in many instances are articulated (Allen & Tarlo 1963). The higher Downton sediments in most places lack invertebrate body fossils, but despite this and their now predominantly red colour, appear to have formed on extensive featureless mudflats

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partly intertidal and partly river-influenced. The air-fall tuffs which became catastrophically interspersed among the muds smother not only what may be intertidal gullies and ponds, but also a variety of trace fossils and a profusion of faecal debris (Allen & Williams 1981, 1982; Williams *et al.* 1982). What is undeniable is that the later Downton sediments, derived from the N or NW, record a substantial enlargement of the depositional area established in Downton Castle times. The coastal mudflats spread southward onto St David's Land, as shown by their occurrence in SW Wales (see figure 3) and at Cardiff (Waters & White 1980), and perhaps northward onto the Irish Sea Ridge at Anglesey (Allen 1965). Similar to more fully marine rocks of possible to probable Downton age occur to the E in borings (see figure 4) at Hamswell (Cave 1977), Little Missenden (Straw & Woodward 1933; Shaw 1969), Soham (Butler 1981), Clare (Bassett *et al.* 1982), Lakenheath (Butler 1981) and Lowestoft (Bullard *et al.* 1940; Stubblefield & Bullerwell 1967). Thus the East Anglian Foldbelt and St David's Land may between them have defined a broad and rapidly widening gulf opening to the SE.

7. Early Gedinnian movements and drainage modification

A striking feature of the Lower Old Red Sandstone rocks above the Downton Castle Sandstone and its correlatives is the number and frequency of pedogenic calcretes (Allen 1974*b*), Quaternary examples of which are reviewed by Goudie (1973, 1983), Reeves (1976) and Milnes & Hutton (1983). These fossil soils, with their distinctive profile, pseudoanticlinal structures, and suite of carbonate fabrics (Allen 1973*a*, 1974*a*, *b*), range through 1.5–3 km of sediment, from the oldest coastal mudflat deposits of Downton age to the youngest Siegenian (?Emsian) fluvial rocks anywhere in S Wales and the Welsh Borders. Their greatest development, however, occurred in the early Gedinnian in the form of the so-called *Psammosteus* Limestones.

The *Psammosteus* Limestones can be traced as a facies throughout S Wales and the Welsh Borders (King 1934). In SW Wales, for example, on Caldey Island (Allen & Williams 1979; Wiliams *et al.* 1982) and at Llanstephan (Allen & Williams 1979), and again in the Forest of Dean (Allen & Dineley 1976), the facies comprises some 15–20 generally thick and well developed calcretes closely set in red mudstones with comparatively few sandstones. Northward into the central Welsh Borders the facies expands in thickness as sandstones appear more frequently and lower down amongst the mudstones and calcretes (Allen 1974*a*; Allen & Williams 1979). A thick development of dolomitic calcretes in Anglesey has been correlated with the *Psammosteus* Limestones (Allen 1965). The calcrete profiles range in thickness from a few decimetres up to 12 m and most attain Goudie's (1983) nodular stage of development. Some of the thicker ones show massive to irregularly laminated upper parts, thus resembling Goudie's (1983) hardpan and laminar types. Many of the Lower Old Red Sandstone calcretes have irregularly eroded and in places channelled tops, overlain by calcrete conglomerates (in some instances, almost *in situ* debris) ranging from stringers one clast thick to internally channelled complexes reaching a thickness of a few metres (Allen & Williams 1979). Individual calcretes can be traced for distances of many kilometres (Dineley & Gossage 1959; Allen & Williams 1982). Thus the calcretes for periods must have formed caprocks underlying extensive plains carrying their own internal systems of drainage. Pseudoanticlinal structures (Allen 1973*a*, 1974*b*) occur in profiles of all thicknesses and degrees of development and have some resemblance to modern forms (Watts 1977). The structures in the least calcified profiles may

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be attributable to clay wetting and drying, but in the more calcified ones probably record in addition the displacive introduction of calcite.

Compared with Quaternary calcretes (see above), these late Silurian and early Devonian examples point to a warm and comparatively dry climate, and to periods of geomorphic stability in an area of coastal and near-coastal sedimentation each lasting up to the order of 10^4 years (Allen 1974*b*; Leeder 1975). Their advanced development and unusual frequency in the *Psammosteus* Limestones facies, however, suggests that during the early Gedinnian (i) the rate of sediment production in the distant source-lands was low overall, and (ii) there were rapid and extreme variations in the local availability of sediment at the depositional site. It seems implausible that such a combination of circumstances could have other than an ultimately tectonic control; geomorphic stability on the short term may therefore have been combined with tectonic instability in a longer time scale.

It is significant for the suggestion of a tectonic control that the *Psammosteus* Limestone facies is associated with (i) the appearance and spread to the S and SE of an indubitably fluvial facies, which thereafter prevailed to the end of Lower Old Sandstone times (see below); (ii) a profound change in the geological make-up of the source area feeding southern Britain, but without a change in the general direction of sediment transport, and (iii) a new vertebrate fauna (Allen & Tarlo 1963).

The most detailed and comprehensive evidence for a change of provenance associated with the accumulation of the *Psammosteus* Limestones facies comes from the central Welsh Borders (Allen 1974*a*). The Downton sandstones in the field are conspicuously micaceous, and under the microscope abound in grains of metamorphic rocks. Garnet and other species typical of regionally metamorphosed rocks predominate in the heavy-mineral crops. Although no details are available, it is clear from reconnaissance work that these compositional features mark the Downton rocks throughout the Welsh Borders and S Wales. The source area supplying Downton sediment appears to have been a distant complex largely of regionally metamorphosed rocks lying in NW Britain (Allen 1974*a*). Only in SW Wales, where some Downton sandstones are pebbly and rich in felsic igneous debris, does a more local source appear to have operated additionally (Allen & Williams 1982). The overlying Gedinnian and Siegenian sandstones, on the other hand, are poorly micaceous, rich in fragments of sedimentary and igneous rocks, and no longer overwhelmingly dominated by garnet in the heavy-mineral crops, although the species remains important. The decline in the importance of metamorphic grains and minerals is clearly attributable to the appearance of sedimentary and igneous rocks in the source area, as shown by the suites of pebbles preserved in the younger (Siegenian–?Emsian) beds of the Lower Old Red Sandstone (Allen 1974*a*, *c*; see also Squirrell & White 1978; Allen *et al.* 1982). These pebbles include Ordovician lavas and tuffs with Welsh faunas, greywackés similar to those of Silurian age in Wales, Silurian shelf limestones, and Downton sandstones with typical mineral assemblages and restricted faunas. The *Psammosteus* Limestones facies, and the major change in depositional style and provenance associated with it, may therefore record the commencement of uplift in the former Welsh Basin and Irish Sea Ridge, and the consequent modification (?splitting and diversion) of the long drainage line from the NW established in Downton times (Allen 1974*a*; Allen & Crowley 1983). There is no compelling evidence for the view (Simon & Bluck 1982) either that a single unmodified drainage system fed southern Britain during Lower Old Red Sandstone times, or that the drainage flowed through the Caledonian Basin (Midland Valley of Scotland).

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8. Gedinnian to Emsian coastal plain of alluviation

Speaking generally, the Gedinnian to Siegenian–Emsian portion of the Lower Old Red Sandstone in the Anglo-Welsh area comprises an upward-coarsening fluvial sequence that ranges from mud-dominated in its older part to sand-dominated and locally even conglomeratic near the unconformity surface beneath the Upper Old Red Sandstone (Upper Devonian). The succession has been studied recently in SW Wales (Allen & Williams 1978; Allen *et al.* 1982; Williams *et al.* 1982), central S Wales (Tunbridge 1980, 1981), the Forest of Dean (Allen & Dineley 1976; Allen 1983), and the central Welsh Borders (Allen 1974*a*). Similar and perhaps slightly younger deposits of mainly Emsian age occur to the E in borings at Faringdon (Falcon & Kent 1960; Mortimer 1967; Chaloner & Richardson 1977), Apley Barn (Poole 1969; Richardson & Rasul 1978, 1979), Beckton Gasworks (Smart *et al.* 1964; Mortimer & Chaloner 1972), and Canvey Island (Smart *et al.* 1964; Chaloner & Richardson 1977). The occurrence of an extensive coastal plain of alluviation is indicated by such a wide spread of deposits (see figure 5). The upward-coarsening evident in the beds, occurring through some 1.5–3 km of rock and culminating in an unconformity surface, testifies to important temporal changes of fluvial style, related to the southward and southeastward march of facies belts across the region and the eventual involvement of the Anglo-Welsh area in general mid Devonian deformation and denudation (Allen 1979).

To judge from the character of the succession, the Gedinnian and early Siegenian rivers were almost as frequently strongly meandering as of low sinuosity. Their deposits are intraformational conglomerates, sandstones and thick calcretized mudstones arranged in sheet-like upward-fining

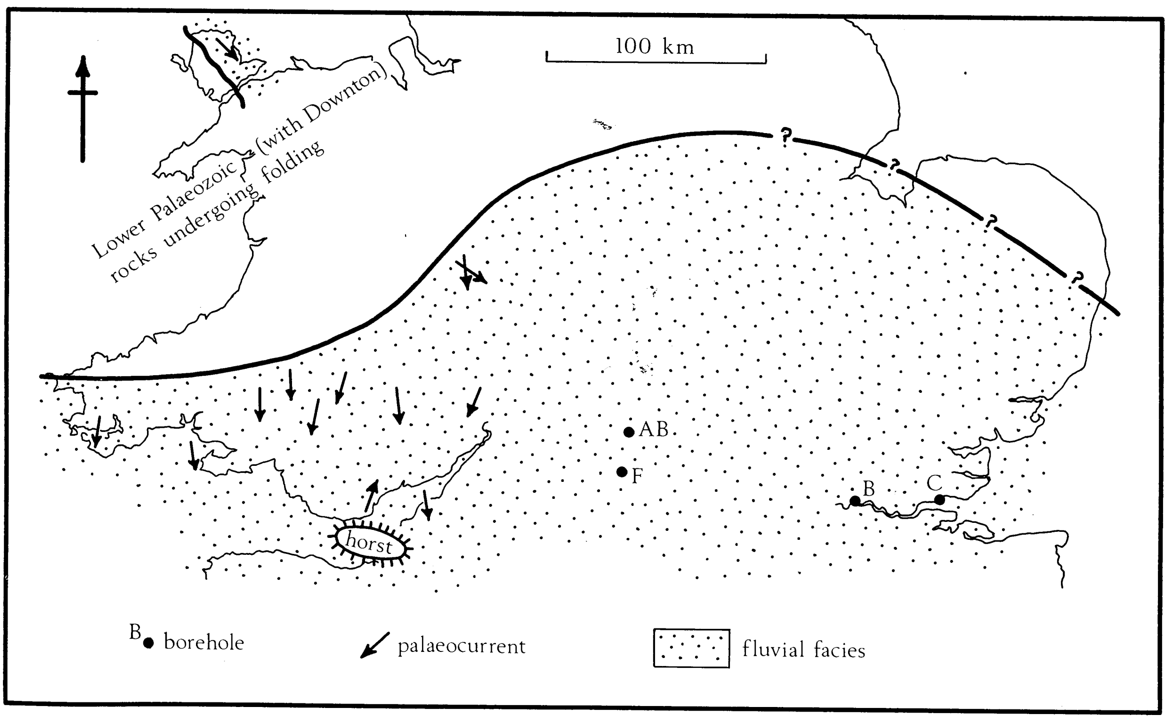

Figure 5. Gedinnian to Siegenian–Emsian fluvial sedimentation in southern Britain. Palaeocurrents mainly after Allen (1974*a*), Tunbridge (1981), and Allen & Williams (1982). Boreholes: F, Faringdon; AB, Apley Barn; B, Beckton; C, Canvey Island.

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sequences with erosional bases. In such widely separated localities as Anglesey (Allen 1965) and SW Wales (Allen 1974*d*; Williams *et al.* 1982) many of the conglomerate–sandstone complexes preserve clear evidence of their lateral accretion in high-sinuosity streams. Other complexes, however, lack signs of lateral accretion (Allen 1974*d*) and therefore may record either channelized low-sinuosity rivers or sheet floods. The vertebrate fauna introduced with the Gedinnian alluvium is strikingly different from that found in the Ludlow and Downton rocks (Allen & Tarlo 1963; Turner 1973). The remains are locally abundant but mainly water-sorted and only rarely articulated (Ball & Dineley 1961; Allen & Tarlo 1963; Miles 1973; Turner 1982; White & Toombs 1983). Vascular plants occur locally (Edwards & Richardson 1974) and many of the mudstones are strongly bioturbated. High-density but low-diversity invertebrate faunas occur at several horizons (Allen & Tarlo 1963; Allen 1973*b*), suggesting short-lived lakes or brackish-water incursions. The Gedinnian and early Siegenian rivers flowed toward the SE in Anglesey (Allen 1965), where deposition may soon have ceased, toward the SE in the central Welsh Borders (Allen 1974*a*), and to the S in SW Wales (Allen & Williams 1982).

The denudation of the Lower Palaeozoic fill of the nearby Welsh Basin is unequivocally demonstrated by the character of the pebbles preserved in the later Siegenian and (?)Emsian fluvial deposits (Allen 1974*a*, *c*; Allen *et al.* 1982). The succession, of northerly provenance (see figure 5), is now sand-dominated and toward the top lacks calcretized mudstones altogether, except as a ghost facies preserved in intraformational conglomerates. Although plant fossils are locally abundant though seldom varied (Smart *et al.* 1964; Edwards 1968, 1969, 1970*b*, 1980, 1982; Poole 1969), suggesting riverside stands, and trace fossils are also plentiful, invertebrate body fossils are unknown. Vertebrate remains are likewise scarce, but continue lineages of Gedinnian establishment (Allen *et al.* 1968; Loeffler & Thomas 1980). The sandstones and intraformational conglomerates form laterally extensive sheet-like complexes, described as multistorey in central S Wales (Tunbridge 1981) and as upward-fining with signs of local lateral accretion in the northern Forest of Dean (Allen 1983). Powerful currents are indicated by the sedimentary structures and coarse textures of these deposits. Although there are differences of detail and interpretation, it is clear that low-sinuosity sand-bedded streams, some deep and multiply braided and others of a shallow distributary type, now dominated the extensive coastal plain (Tunbridge 1981; Allen 1983). Their fall-line to the NW would have been clearly visible to an observer in either S Wales or the Welsh Borders.

There are signs of instability in the Bristol Channel area during the long terminal fluvial episode, for the Llanishen Conglomerate around Cardiff (Squirrell & Downing 1969) originated nearby to the S (Allen 1975), perhaps as an alluvial fan spread from a horst or thrust-block. Another similar local source may have operated in central S Wales (Tunbridge 1980).

9. Discussion and conclusion

The events which either interrupt or punctuate the marine to fresh water transition, that is, the end-Ludlow movements and the early Gedinnian drainage rearrangement, allow the division of the transition into three main phases.

The first, lasting some 8 Ma, is the Ludlow infilling of the Welsh Basin. Assuming that the early Ludlow turbidites accumulated in water depths similar to those of Japanese fore-arc basins (Okada & Smith 1980), an average rate of shoaling of the Ludlow Welsh Basin of the order of

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5×10^{-4} m a^{-1} is indicated. An average deposition rate (consolidated sediment basis) of roughly 2.5×10^{-4} m a^{-1} may be suggested, as the Ludlow rocks are nowhere thicker than about 2 km. This may seem high, but it implies no great total supply of sediment to the Welsh Basin, as the thicker sequences are restricted to the comparatively narrow Montgomery and Denbigh troughs (see figure 2*a*). These inferences required no drastic modification when including the Midland Microcraton in the calculation, on which there is almost an order of magnitude reduction in the thickness of Ludlow. By the end of the Ludlow, the sea may have been retracted to central Wales and had certainly become very shallow.

In the second phase, covered by the Downton, lasting possibly 2 Ma, there is a rapid expansion of the marine-influenced area and the development of sand shoals followed by an extensive and relatively featureless coastal mudflat complex (see figures 3 and 4). This complex, as now preserved in the Welsh Borders and S Wales, experienced frequent and rapid changes from a marine to a freshwater influence. Calcretized mudstones and *Lingula*-bearing sandstone–mudstone sequences, for example, alternate several times within the clearly marine-influenced earlier Downton sediments above the sand–shoal facies. The higher Downton rocks are broadly similar to these, but are less easily diagnosed environmentally, as they lack lingulids and other clearly marine body fossils. The Downton beds are approximately 500 m thick, which points to an average deposition rate similar to that which prevailed during the Ludlow. However, the Anglo-Welsh area had by now become connected to a distant metamorphic source-land, and the total of sediment received by it probably increased by at least an order of magnitude, in view of the scale of the region it is necessary to think of as having originally been covered by Downton deposits.

Carbonate-bearing soils indicative of a warm and comparatively dry climate are present throughout the Lower Old Red Sandstone, but in the *Psammosteus* Limestones (early Gedinnian) achieve a quite remarkable frequency and degree of individual development. This unusual concentration is associated with a major change of both sediment provenance and depositional facies. There is a concomitant change in the vertebrate fauna from 'Silurian' to 'Devonian' forms (Allen & Tarlo 1963). In the more southerly parts of the Anglo-Welsh area, the *Psammosteus* Limestones have the aspects of a condensed sequence; calcrete profiles are piled on top of each other with little intervening in the way of mudstones, and with rare sandstones. Northward the facies is somewhat thicker and here sustained fluvial deposition appears to have begun a little earlier. The denial of sediment to the area suggested by these features, and the observed change of provenance to a nearer source, is consistent with a rearrangement of drainage lines over a substantial part of the British Isles. The lengthy Downton system, reaching far to the NW, may have become split, with the more northerly portion becoming diverted westward, leaving the now-isolated southern part to spread its tributaries over the upwarped Lower Palaeozoic (and Downton) rocks beginning to emerge in the region of the Irish Sea and N Wales. As the *Psammosteus* Limestones facies has a characteristic thickness of a few tens of metres, this major geomorphological revolution may have been completed in as little as 10^5 years.

The rearrangement of drainage suddenly introduced to the Anglo-Welsh area a series of fluvial deposits which spread progressively further south without any perceptible large-scale interfingering with marginal-marine sediments. These constitute the third phase of the transition. The area from the Gedinnian to the Siegenian and Emsian was part of an extensive coastal plain of alluviation. The plain received huge volumes of sediment from the source-lands to the NW, but the Downton average deposition rate (2.5×10^{-4} m a^{-1}) was approximately maintained, allowing 10 Ma for the Devonian portion of the Lower Old Red Sandstone.

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Speaking generally, the rock succession discussed above presents few problems of environmental interpretation, provided that a broad brush will suffice.

In purely physical terms, the early Ludlow turbidites and slide deposits are quite representative of their class, and speak of a deep-water environment which, in the context of the Phanerozoic, is unlikely to have been other than oceanic or perioceanic. The turbidites comprise a monotonous alternation of sharp-based graded sandstones with mudstones on a scale of centimetres to decimetres. In addition to being graded from coarse up to fine, the sandstones show an orderly upward sequence of sedimentary structures indicative of sediment transport within a relatively narrow range of directions both locally and regionally. Thus these rocks agree in all essential respects with modern deep-sea sand deposits (abyssal plains, outer parts of deep-sea fans) attributable to turbidity current action below wave base. The Ludlow slide deposits, with their slope-related internal folds and pull-aparts and their upward truncations, can also be matched with modern oceanic sediments, in this case attributable to mass movements on submarine slopes. The invertebrate fauna, which unusually for such facies is relatively plentiful, is therefore not the leading criterion for the marine origin of the Ludlow beds.

The Gedinnian and younger fluvial rocks are also highly distinctive in physical terms alone, and present no problems of broad interpretation. Like the deposits of modern rivers, with which they may be closely compared, the rocks occur in erosively based upward-fining sequences on a scale of a few to many metres. The range of sedimentary structures found in the sandstones is the same as that encountered in modern sand-bed streams, and the palaeocurrents indicated, although dispersed, are invariably unidirectional, as is observed of rivers today. The occurrence of frequent desiccation cracks in this facies points to atmospheric exposure on a scale of hours to days, while the abundant presence of fossilized soil materials suggests that atmospheric influences prevailed at times over periods individually as long as 10^4 years. The presence of vascular plants and vertebrates within the beds is therefore not an essential criterion of origin, although of course consistent with the interpretation offered above.

The main difficulties of even broad interpretation on the basis of physical evidence arise in the case of the later Ludlow and earliest Downton sediments, presenting a combination of features which could arise in large lakes as well as in shallow seas, and with the equally ambiguous later Downton deposits which are largely unfossiliferous. In the case of the former, the presence of a high-diversity and high-density marine invertebrate fauna is crucial to their (shallow-marine) environmental interpretation. In the case of the later Downton rocks, the intimate association of fossil soils with mudstones and sandstones containing a low-diversity but commonly high-density marine fauna (lingulids, ostracodes, some bivalves and gastropods) can only point to a rapid and frequent alternation of terrestrial with marine influences in a complex of high-stress marginal environments. Precisely what these environments were, and to what extent we may assign to them the unfossiliferous rocks making up the greater part of the Downton, remain questions for future research. At present one can assign the later Downton rocks to a marginal or mixed environment, on only the most general grounds.

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Discussion

J. B. Richardson (*British Museum (Natural History), Cromwell Road, London SW7 5BD, U.K.*). Palynomorphs and palynological debris can provide a great deal of information on the environment of deposition of marine and fresh water sediments. In association with Dr S. M. Rasul I have worked on a number of sequences covering the Upper Ludfordian and Downtonian. Miospores, plant cuticles and 'tubes' increase in abundance towards the land and studies of Recent sediments off the Orinoco delta (Muller 1959) show that plant cuticles in particular are most abundant near the mouths of rivers and their distributaries. In contrast acritarchs are marine phytoplankton and their diversity increases offshore in a similar fashion to modern dinoflagellate cysts. No Palaeozoic acritarchs have been found in continental sediments but the nature of acritarch assemblages found by us in near shore sediments is very different from assemblages farther from the shore (as indicated by miospore abundance and palynodebris). Graphs for two sections at Downton and at Long Mountain showing miospore abundance, an inshore index and a marine influence index (based on the relative abundance of acritarchs in the major groups) show a parallel series of fluctuations in the two sections across The Ludfordian–Downtonian boundary and in the overlying Lower Downtonian. But the Long Mountain sections provide evidence that a more open marine situation existed, above and below the basal Downtonian boundary, than in the comparable section in the Downton area.

Evidence that lower Palaeozoic sediments were being eroded in early Downtonian times is provided by the presence of reworked acritarchs in the assemblages from the Downton area. In contrast the sedimentological evidence indicates that only metamorphic terrains were providing sediments at that time.

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J. R. L. Allen. The micropalaeontological studies of Dr Richardson and Dr Rasul add significantly to our understanding of events around the Ludlow–Downton boundary and give added weight to the existing evidence that the Ludlow seas lingered longer in central Wales (Clun, Kerry, Long Mountain) than in areas to the northwest, east and south. The richly fossiliferous early Downton succession in Central Wales is much muddier than elsewhere, presents a comparatively long interbedding of sand and mud lithologies and, in lacking the Ludlow Bone Bed, arguably represents a more continuous deposition than in the Welsh Borders and central South Wales. However, it is only in the Downton beds higher than the Downton Castle Sandstone and its correlatives that metamorphic debris from the far north or northwest is overwhelmingly predominant. While the influence of metamorphic source rocks is already evident in the Downton Castle Sandstone and its correlatives, the petrographic and mineralogical evidence allows for contributions from other types of source. Indeed, palaeogeographical reconstruction for earliest Downton time suggests that contributions could also have come from Lower Palaeozoic rocks in the Bristol Channel area, the Irish Sea Ridge and, possibly, from a folded belt of early Lower Palaeozoic rocks in eastern England. Contributions from these sources would have become progressively less important later through the Downton as the area of sedimentation, fuelled from the northern metamorphic source, spread outward from its initial locus in central Wales.