Fluvial Geomorphology of Great Britain

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Chapter 6

Fluvial geomorphology of central and southern England

K.J. Gregory with contributions from R.J. Davis

FLUVIAL LANDFORMS AND PROCESSES IN CENTRAL AND SOUTHERN ENGLAND

Introduction

The boundary between the 'Highland Zone', of the north and west of Britain, and the 'Lowland Zone', of the south and east, is usually taken as a line running approximately from the mouth of the River Exe to the mouth of the River Tees. The area defined here as central and southern England includes some portions of the Highland Zone. These are principally found in south-west England, where three upland plateau areas of western Cornwall rise above a landscape that is dominated by coastal low plateaux cut across resistant Palaeozoic rocks. Also representing the upland scenery and resistant rock theme of the Highland Zone are the Mendips and the Quantocks, although the limestone features of such areas as the Mendips are included in another volume in this series. The central and southern Pennines also have the relief, scenery and resistant rock outcrops characteristic of the Highland Zone. Such areas have features similar to those already described for Wales and for north-west and north-east England.

Much of the remainder of the area is made up of two types of landscape which are scarplands or lowlands. A cuesta is an asymmetrical feature made up of a steeper scarp slope and a comparatively gentle dip slope that leads into a vale which is usually underlain by clay. This scarpland pattern of alternating scarp, dip slope and vale is a recurrent theme across much of central and southern England and makes up the scarpland landscape of much of southern Britain. The scale of cuestas varies very considerably from one area to another. Overlooking the lower Severn valley, the scarpland of Jurassic rocks at Birdlip Hill is up to 250 m above the Vale of Gloucester. The Cretaceous Chalk outcrops of southern, southeastern and eastern England are also characterized by cuesta landscapes, and there are major scarp slopes bordering the Chalk outcrops in Lincolnshire and in south-east England, where there are scarps bordering the North and South Downs. Although these are the major scarp and dip slopes making up cuesta landscapes, there are many smaller cuestas picking out variations in lithology that make up the diverse geological map of southern and central England.

A second type of area is made up of lowlands, which include the areas of the Fens, the London Basin, the Hampshire Basin, the lower Severn valley and the Trent and Ouse lowlands of eastern England, and also the Vale of Pickering, between the Jurassic rocks of the Yorkshire Moors and the Cretaceous rocks of the Yorkshire Wolds. In these lowland areas the surface rocks and deposits are usually comparatively young and often include sequences of recent Quaternary sediments.

Prevailing themes

Although the contrast between the Highland and Lowland zones, and also the basic character of the cuesta landscape and of the lowlands, owes much to the underlying rock type and the superficial deposits, there are several other distinctive themes that have influenced the pattern and character of river development in central and southern England.

Firstly, there are still traces of the original east-flowing rivers which were thought to have provided the ancestors of the present river system in Britain. Thus the Trent has east-flowing sections that were complemented by a north-flowing section when rock types allowed the development of subsequent streams. The Thames is also dominantly eastward-flowing, and originally there was an east-flowing major river along the line of the present Solent. Many characteristics of rivers in central and southern England reflect the fact that the headwaters of many rivers originate in the Highland Zone. The fluvial characteristics of some rivers of central and southern England need to be seen as depositional counterparts of fluvial systems of the Highland Zone. Thus headwaters of the Severn and Wye in Wales are associated downstream with the characteristics of the middle and lower Severn, draining into the Bristol Channel. Similarly, the Trent, Ouse and Humber receive characteristic Upland rivers from the Pennines and from the North York Moors.

A second major characteristic of central and southern England is that the recent evolution of the landscape is still firmly imprinted on the scenery. This is particularly significant because the most recent glaciation did not extend over much of this area, and the most southerly maximum limit of Quaternary glaciation was north of the present River Thames. This means that, according to the influence of glaciation, three major areas can be distinguished in central and southern England: one north of the Devensian limit, influenced particularly by deposits in the most recent glaciation; a second, to the south, where there are areas influenced by earlier glaciations; and a third, over the southernmost areas and in south-west England, where there are areas which were periglacial and beyond the maximum limits of Quaternary ice sheets. Whereas, in this southern area, the periglacial influence of permanently frozen ground (or 'permafrost') and a much more seasonal climatic regime produced river systems in the past which were different in character, regime and extent from those of today, in formerly glaciated areas there are some thick deposits of till, fluvioglacial sands and gravels, and diversions of drainage such as that of the Severn, which have all influenced characteristics of the present fluvial system. Along rivers such as the Severn, the Trent and the Thames, terrace sequences clearly indicate the stages of river and valley development.

A third theme arises from interaction with the sea. Estuarine influences in the areas adjacent to the Humber, the Bristol Channel, the Severn estuary, the Thames and the Wash arise because river activity and sediment transport and accretion interact with estuarine circulation and sedimentation. The finer sediments that reach estuarine environments may be influenced by different chemical processes involving, for example, flocculation of clays in saline waters. The largest area of alluvium in Britain surrounds the Wash, and sedimentation in this area, and also in the Vale of York, the Thames estuary and the Somerset Levels, has resulted not only from offshore sources from marine sedimentation, but also from peat and salt marsh development (Lewin, 1981).

Human activity is a fourth theme, because it has had a greater effect on the fluvial system of central and southern England than in other parts of Britain. Because the average population density is greater in this part of Britain, there is the potential for the fluvial system, rivers and river channels to be modified much more extensively than in other areas. The impact of human activity has been registered substantially in the past and continues to be significant at the present time. Deforestation is now known to have had a significant influence upon lowland rivers, and considerable amounts of alluvium have accumulated along rivers, including the Severn and the Thames, as a consequence of deforestation over the past 4000 years, which released suspended sediment transported by rivers and accumulated along floodplains. Although human disturbance of woodland may have influenced valley alluviation, Macklin and Lewin (1994) have suggested that climatic controls, including storms and floods, were important. More recent drainage modification, particularly along river valleys in the Chalk areas of central and southeastern England, include water meadows developed for irrigation purposes, which still complicate the pattern of river channels. In some cases the precise impact of human activity has not been so easy to reconstruct and, whereas an early view (Jennings, 1952) regarded the Norfolk Broads as 'natural', it was subsequently realized that they are relics of medieval peat digging (Lambert *et al.*, 1970). At the present time along the Norfolk Broads, the implications of human activity continue, since bank erosion arises from the scour induced by boats, including pleasure cruisers, and this has necessitated protection measures in Norfolk.

In recent years, therefore, many river channels in central and southern England have been modified (Brookes *et al.*, 1988) and the extent of channelization between 1930 and 1980 is demonstrated in Figure 1.6. The density of channelization in England and Wales is, on average, 20 times greater than that in the USA, and in this area of central and southern England, the average density of channelization reaches is greater than 0.06 km^2.

Channelization is a direct modification of river channels, but the indirect consequences of other human activities have also been reconstructed in recent years (Gregory, 1987). The effects of these human activities on river channels and on the processes operating in the drainage basin can lead to the downstream changes. An excellent example is provided by the effects of urban areas, which provide greater impervious surfaces, encourage more runoff to flow directly and more rapidly into streams and rivers, and so potentially can increase the amount of discharge downstream of the urban area. This can lead to increased flooding, which has been a reason for the widespread introduction of flood prevention schemes in many parts of central and southern England. The increased discharges can also cause channel erosion. Because the size, shape and character of the river channel reflect the amount of discharge and the sediment conveyed to a particular cross-section, if the water and sediment delivery change as a result of urbanization, then so the river channel may similarly also change. There are many cases throughout Britain where stream channels have changed in the past century as a consequence of the increase in flows following urbanization. A change in river channel dimensions, often involving a decrease, can occur downstream of dams and reservoirs. The impoundment of reservoirs can reduce the frequency of downstream flooding, and this reduction of the high discharges delivered to a particular river channel means that, over a period of years, the river channels can adjust their morphology and ecological characteristics as a consequence of the changed flows (Petts, 1988b). Adjustments of river channels in central and southern England range from point changes, such as those due to the input of water from stormwater drainage systems from urban areas, to the construction of bridges, the construction of dams across the channel, and the extraction of sediments from the channel. There are also spatial changes that affect the whole or part of the drainage basin, including changes of land use from forest to farmland. It is therefore essential that, in the interpretation and appreciation of rivers and river channels in central and southern England, one recognizes the extent to which human activity has had a continuing role in producing the channels that we see today.

A fifth dominant influence is that the rivers of central and southern England are essentially low-power channels, and in this sense contrast dramatically with those of Wales, northern England and Scotland. Thus Ferguson (1981) showed that the stream power for rivers in this part of England was often less than 0.1 kWm^{-1}, which is much lower than the values characteristic of Wales and of Scotland. Low stream power values in this part of the UK can be thought of in relation to three major characteristics: firstly, because of the low relief in much of central and southern England, which is associated with low river velocities; secondly, because precipitation totals are lower than in other parts of Britain and much of southern England receives less than 1000 mm per year, except those parts of the region that are in the Highland area; and thirdly because the extent of the permeable rocks is very considerable in this area. These permeable rocks can act as aquifers and, as the proportion of groundwater flow is quite considerable, the regimes of the rivers are characterized by high base flows fed by groundwater sources and comparatively low peak flows fed by storm runoff. This means, in turn, that stream power values tend to be relatively low and that channel dimensions also tend to be comparatively small compared with those in the Highland Zone.

GCR SITE SELECTION

Against this background, the selection of GCR sites in central and southern England was quite difficult. Some potential sites were excluded if their features indicated that they were more closely related to other GCR Blocks, for example, river terraces associated with Quaternary development (see, for example, Bridgland, 1994, Allen *et al.*, in prep., Campbell *et al.*, in press) or sites typical of limestone areas (Waltham *et al.*, 1996). Other sites were excluded if they were heavily modified by human activity or at present maintained by human activity. The approach taken initially was to identify all locations which have been cited in the literature and for which there is some published information which could be a useful basis for site selection. These criteria allowed the initial identification of 214 possible sites, which were then classified into 15 classes according to the type of features that they contained. The distribution of the 15 classes throughout central and southern England was analysed, and then the extent to which sites were 'unique', a classic example, or a type example of a particular feature was considered. It was eventually possible to select the representative suite of GCR sites at the locations shown in Figure 1.7, and these can be thought of in five major groups. Firstly, there are those which reflect the significant contribution made by Quaternary development to the fluvial system. This category includes the sediments at Beckford, and at Ashmoor Common, the striking Ironbridge Gorge associated with the Buildwas site, and also examples of channels that are now underfit (Cherwell, Itchen), but which originally carried much higher discharges in the past, as indicated in Dury (1977). A second group represents aspects of contemporary river-channel activity, and this involves not only examples from the Highland Zone at Alport in the southern Pennines and waterfalls and gorges in the Yorkshire Pennines at Aysgarth and the Lynford Gorge on Dartmoor, but also channel planform features on the Axe and Exe, and a channel in Essex to indicate the effect of a catchment overlain by till. Also significant is the influence of organic accumulations in channels, and this is exemplified at the site in the New Forest, which must have been typical of many other sites in central and southern England in the past. A third category includes those landforms produced by changes in channel dimensions and this is exemplified by changes downstream of reservoirs along the Derbyshire Derwent (Petts, 1977). A fourth type results from a particular flood event, the Lynmouth flood in 1974, which had dramatic effects on the catchment area of the River Lyn (Anderson and Calver, 1977). The final category includes an example of a managed channel along the River Lugg.

The selection has resulted in a skewed regional distribution, with more of the sites being in the western part of the area and fringing or within the uplands. This reflects the criteria set out and also the weight of research.

BECKFORD, HEREFORD AND WORCESTER (SO 978361)

Highlights

The deposits associated with a river terrace at Beckford provide an important record of palaeoenvironmental conditions in the Severn basin, based upon reconstruction of past fluvial and slope processes.

Introduction

Beckford is an important site of Late Devensian slope and adjacent floodplain deposits, which are particularly significant in relation to environmental change in the Severn basin. The deposits here include gravels associated with mass movements and slope wash (Dawson, 1986), and sand which may have been imported by aeolian transport (Briggs *et al.*, 1975). These sediments record transport processes operating under different palaeoenvironmental conditions. A number of reactivation surfaces indicate episodic activity. Downslope, the deposits interdigitate with lithofacies deposited by water flow parallel to the valley axis, marking a transition to the terrace deposits.

Description

The Carrant Valley is cut into the Lower Lias Clay, between outliers of the Cotswold scarp: Alderton,

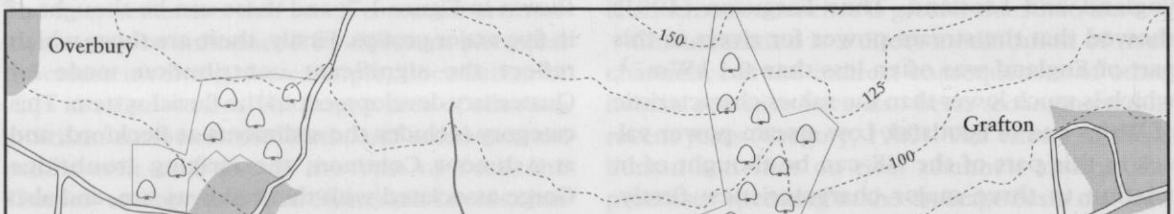

Figure 6.1 The extent of the terrace deposits at Beckford: a map showing the distribution of terrace outcrops in the Beckford area.

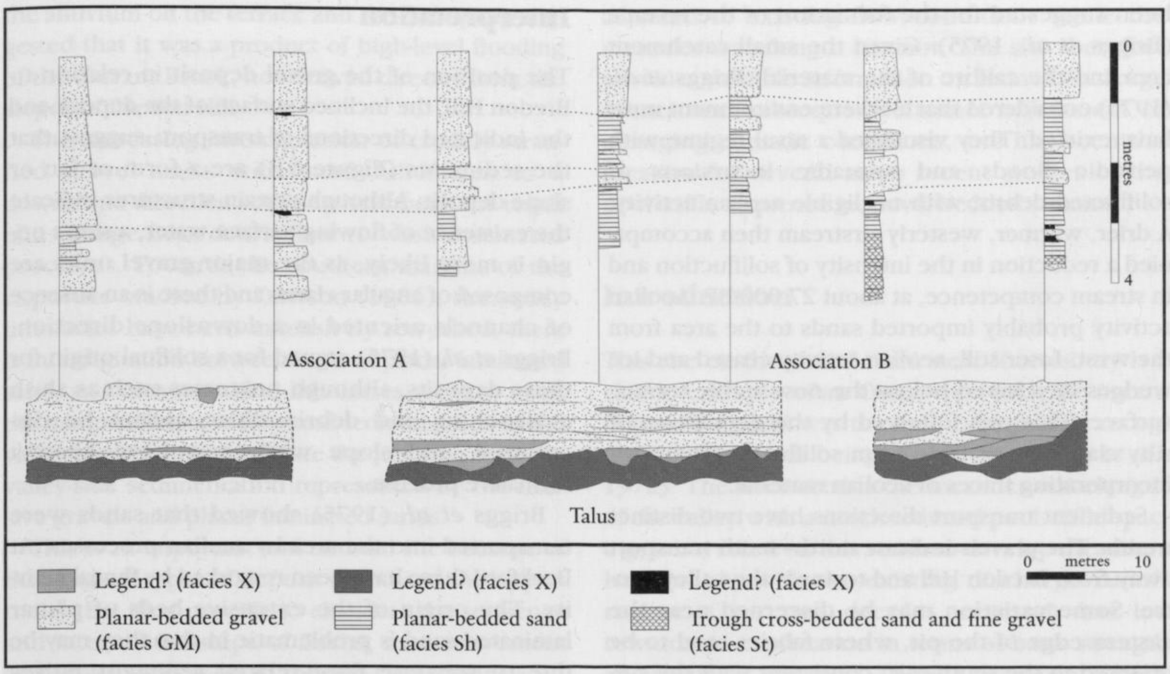

Figure 6.2 Sedimentary sequences at Beckford: vertical sections showing sedimentology and facies associations through terrace. (After Dawson, 1986.)

Oxenton and Crane Hills to the south and Bredon Hill to the north. Richardson (1929) suggested that the Carrant Brook originally extended further eastwards, probably being linked to the Isbourne system. Carrant Brook is a small underfit stream in a valley containing extensive terrace deposits (Figure 6.1), which have been related to the terraces of the Avon valley. The main terrace deposits of the Carrant Brook have been dated at 27 650 ± 250 BP (BIRM-293), and those of the Avon No. 2 terrace at between 38 000 ± 700 BP and 26 000 ± 300 BP, placing both in the latter half of the mid-Devensian (Briggs *et al.*, 1975).

Shotton (1968) suggested that the most recent phase of widespread terrace formation took place in the Late Devensian. Between the Upton Warren and Late Devensian interstadials, the rivers had incised their valleys significantly. It is therefore likely that downcutting was at full glacial rather than interstadial times.

Standing 0–15 m above the present floodplain, the Carrant main terrace extends almost continuously along the right bank from near Grafton to Kinsham. On the left bank, it is widespread near Aston on Carrant, and in the Isbourne valley at 47 m OD. It has been best exposed at Beckford (Briggs *et al.*, 1975). The exposure now remaining, and forming the focus of the site, does not show the underlying terrace deposits but does show slope deposits, the first sequence studied by Dawson (1986), which overlie the Carrant main terrace, forming the second sequence (Figure 6.2).

The exposure of slope deposits which still exists was studied by Dawson (1986), in relation to exposures nearer the valley centre and further deposits below the present ground surface (Figure 6.1), which are now obscured by standing water. The two distinct sedimentary sequences are composed of gravels and sands. In the first sequence (Association A, Figure 6.2), gravels interdigitate with a variety of sandy lithofacies. The gravels on the footslope of Bredon Hill are coarse, poorly sorted and almost wholly of local origin; nearer the river, where they form the basal terrace deposits, they are water-worn, well-bedded, finer gravels and often intercalated with well-sorted sands.

In contrast, the second sequence (Association B, Figure 6.2), comprises sands which often truncate these gravels, and are almost wholly different sediments. Briggs *et al.* (1975) considered these to be foreign material, wind-blown from the west. Within the deposit as a whole there is a general downslope transition from the coarse gravel lithofacies.

Three palaeoenvironments of deposition have been suggested for the formation of the terrace (Briggs *et al.*, 1975). Given the small catchment area and the calibre of the material, Briggs *et al.* (1975) considered that a severe environment must have existed. They visualized a nival regime with periodic floods and sporadic incursions of soliflucted debris, with negligible aeolian activity. A drier, warmer, westerly airstream then accompanied a reduction in the intensity of solifluction and in stream competence, at about 27 000 BP. Aeolian activity probably imported sands to the area from the west. Later still, aeolian activity ceased and ice wedges developed below the now stable terrace surface. This was followed by the deposition of silty clay loam or head from solifluction material, incorporating traces of aeolian material.

Sediment transport directions have two distinct trends. The gravels indicate north–south transport away from Bredon Hill and towards the valley centre. Some variation may be discerned near the eastern edge of the pit, where fabrics tend to be oriented to the south-east, consistent with the surface morphology. Towards the valley centre, a limited number of sand structures show westward dips, aligned parallel to the orientation of the valley axis, and normal to the transport direction indicated by the gravel fabrics.

Interpretation

The position of the gravel deposit in relation to Bredon Hill, the inclined surface of the deposit and the indicated directions of transport, suggest that the sediments (Figure 6.3) are a form of fan or slope deposit. Although certain structures indicate the existence of flowing surface water, a slope origin is more likely, as the major gravel units are composed of angular clasts, and there is an absence of channels oriented in a downslope direction. Briggs *et al.* (1975) argued for a soliflual origin for these deposits, although processes such as slush avalanching and debris slides cannot be discounted, and slope wash may have been a formative process.

Briggs *et al.* (1975) showed that sands were transported into the area by aeolian processes. At Beckford these have been reworked by fluvial activity. The origin of the extensive beds of planar laminated sand is problematic in that they may be due to separate flood-related sediment pulses (Frostick and Reid, 1977); or they may be partly the product of aeolian deflation and deposition. However, a number of characteristics indicate that they may represent reactivation horizons. Brown (1983c) indicated that the textural uniformity of

Figure 6.3 The exposure of terrace deposits at Beckford. (Photo: R.J. Davis.)

the alluvium on the terrace and the floodplain suggested that it was a product of high-level flooding of the Carrant Brook, not derived directly from the slopes of Bredon Hill.

The increasing interdigitation of crudely stratified gravel units with sand lithofacies in a downslope direction seems to have been the result of a change from debris flow to wash-dominated processes. Towards the southern margins of the exposure studied by Dawson (1986), the sandy lithofacies have been extensively reworked, there is interdigitation between beds of planar laminated sand, and cross-stratified lithofacies seem to mark the marginal limits to flow and sedimentation in the valley centre, and to indicate the junction with the valley-side sedimentation represented by the massive gravels and planar laminated sands.

This is an important and unusual site because of the range of types of deposit found and the opportunity that they provide to elucidate the sequence and interrelationships of fluvial, mass-movement and aeolian processes in the mid-Late Devensian.

Conclusions

The sedimentary sequence at Beckford is most recently interpreted as the product of slope processes. These seem to have included both soliflual mass movement and slope wash, although much of the sand may initially have been emplaced by aeolian processes. These slope deposits show a number of major reactivation surfaces, indicating that activity on the slope was episodic, although the causes of this are unclear. At its southern margin the slope deposits interdigitate with lithofacies deposited by flow parallel to the valley axis, marking a transition to the terrace deposits. This spatial relationship is rare evidence of processes that are significant during and following the Late Devensian in the Severn–Avon basin.

RIVER SEVERN AT MONTFORD, SHROPSHIRE (SJ 396153–SJ 432153)

Highlights

The River Severn at Montford represents an excellent example of river-channel underfitness whereby the river-channel system still relates to a previous climatic condition of greater runoff. In this case the pool–riffle sequence and channel size have adjusted to existing flow conditions while the meander wavelength has not. The site therefore provides valuable information on former discharge events and subsequent channel adjustment to changing climatic conditions, and has been the location for several more recent studies of the pool–riffle sequence in gravel-bedded streams.

Introduction

This site on the river Severn near Shrewsbury is a classic example of an Osage-type underfit stream (Figure 6.4), and was the first to be overtly recognized in Britain as being of this type (Dury *et al.*, 1972). The river retains the meander wavelength of the valley with a much reduced pool–riffle spacing; it has failed on this reach to develop meanders harmonious with the existing discharge–runoff regime. Over two decades, research has led towards an explanation in terms of bank strength and sediment transport. The site has recently been studied in relation to the 'velocity-reversal hypothesis' over pool–riffle sequences (Carling, 1991).

From details of the section exposed on the reach at Preston Montford, and other regional evidence, M.D. Jones (1982) concluded that many of the anomalous channel reaches including this straight reach between Shrawardine and Montford appear to have been inherited from the period of deglaciation of the area, and were associated particularly with the formation of glacial lakes and the lower terrace suite of the Upper Severn.

River Channel

Description

This part of the Severn cuts through the glacial and proglacial deposits of the Irish Sea Glacier, including outwash material, ablation till, moraines and lacustrine sediments (Shaw, 1972). Fluvial sediments include floodplain alluvium and the sands and gravels of undated postglacial terraces (Dury, 1983). Most of the stream banks are cut in till and/or alluvium and are rarely deformed; when this is the case, the banks slump at high angles. Part of the right bank of this section is cut in bedrock, in which pools have been excavated by the river. The slope of the valley long profile is low and is determined essentially by deposition rather than by fluvial erosion.

This stretch displays trains of valley meanders,

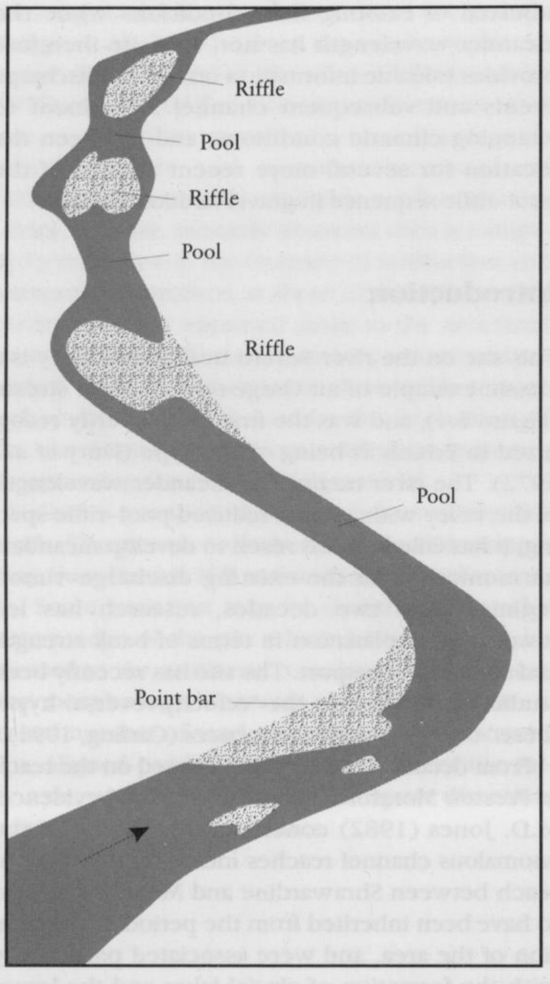

Figure 6.4 The pool–riffle sequence on valley bends of the Osage river, derived from aerial photographs. (After Dury, 1970.)

whereas the existing headwater channels of the Severn display stream meanders. The latter are either very poorly developed or are absent downstream of the Vyrnwy confluence, until a short series of loops near Leighton, a short distance upstream of the Ironbridge Gorge.

Precipitation is seasonal, with monthly maxima in September (78 mm) and December (70 mm), and minima in February (40 mm), recorded between 1956 and 1968. At the Montford Bridge gauging station the channel has a width of 42 m, a mean depth of 4 m, and a maximum depth of 6.4 m. The channel is nearly rectangular in cross-section. Bankfull discharge is approximately 200 cumecs, maximum discharge 473 cumecs, the estimated Manning's n = 0.035, and the drainage area is 2025 km^2 upstream of Montford.

Interpretation

Recent research has examined the sensitivity of alluvial stream channels to climatic change, according to its effects on the characteristics of the drainage basin. Underfit streams are now generally regarded as responses of channels to climatically induced reduction in runoff (Dury, 1983). The Severn drainage was strongly affected by a major climatic change between 12 000 and 8 000 BP; the existing combination of plan, cross-section and profile geometries represents a combination of harmony and disharmony with the present climate (Dury *et al.*, 1972).

After a reduction in discharge, a stream adopts either a shorter wavelength within the long-wavelength valley meanders (manifestly underfit), or retains the wavelength of the valley with a much reduced pool–riffle distance (Osage-type underfit), which reflects the need for smaller channel dimensions. Underfit streams are widespread in central England, although by no means the only types of underfit represented. The Severn here is an Osage-type underfit, which combines a river channel significantly reduced from its former dimensions with a pool and riffle sequence adapted to the existing channel processes, but without stream meanders (Dury, 1966). The identification of these conditions has generated considerable discussion (Kennedy, 1972; Kirkby, 1972; Richards, 1972; Ferguson, 1973). The general question has been (Dury, 1983) stated as: Why should a stream with a discharge sufficiently diminished to have reduced its channel dimensions by a factor of 5–10, and remaining capable of deforming its bed profile, fail to generate stream meanders?

Both the existing and the former, larger, stream channel may be compared to the stereotype meandering stream (Dury, 1977), not only in relation to dimensional characteristics but also in relation to hydraulic characteristics. Dury *et al.* (1972) concluded that the pool–riffle sequence is adapted to the existing river channel width; harmonious with the existing discharge and runoff regime, but disharmonious with existing climate–discharge relationships, in that the meanders of the present channel are not appropriate for the present discharge regime. However, both the size and composition of the riffles seem to depend upon local sediment supply. Movement of gravel within the modern river channel is probably restricted. However, local outcrops of terrace sediment result in large gravel riffles, whereas riffles in the vicinity of loam alluvium tend to be poorly developed. This principle also applies to point bars, the composition of which depends upon sediment supply just upstream.

The conversion of a straight to a meandering channel begins with deformation of the bed, but this need not necessarily lead to deformation of the banks. Dury inferred that this is the case for this part of the Severn. Shrinkage of the stream has been accompanied by deformation of the bed, but by little or no deformation of the banks. A decade of measurements have been made, which allow a provisional explanation of why the former stream could meander, whereas the present stream cannot.

Dury's hypothesis was that the unduly low slope of the study reach so reduces stream power, shear stress and frictional shearing velocity on the banks that little bank deformation can occur. The observed slopes are about half of the stereotype slopes, although the observed velocities through the cross-sections closely resemble the stereotype values; the slope-dependent hydraulic characteristics of stream power, shear stress and frictional shearing velocity at the boundary as a whole are low on the present stream. Some contribution is also made by bank strength, but given the results of the hydraulic calculations, the main influence of bank strength could be thought to be the influence exerted on the channel form ratio. Two measurements have been taken to produce a third value: banktop width and water surface width give the width of batter. All three vary along the channel far more rapidly than would be expected from randomness. Variation in bank strength appears to account for about half the variation in width between banktops.

Bank exposure

Description

The river terraces of the Severn valley between the Vyrnwy confluence and Ironbridge are divisible into three suites. The upper suite terraces (6, 5, 4) are limited to a 6 km valley section. The middle suite terraces (3, 2) extend the length of the valley from the Isle at Shrewsbury to Ironbridge, and are only exceeded in length by the low (1 + floodplain) suite. The lower terraces adopt a meandering course, the dimensions of which indicate a channel of medium sinuosity (1.32) about twice the size of the modern river.

Lake silts shown in the cliff section at Preston Montford, (Figure 6.5) can be traced through stream banks to west of Onslow Hall at 76 m OD. The section has a basal Welsh Ice till unit. Overlying this is a very poorly sorted unit of gravel and cobbles (up to 15 cm across), with a sand/fine gravel matrix. Stone orientation, sorted laminae of sand and openwork fine gravel indicate a general dip of the deposits to the south at 15–20°. The contact of the till with this gravel is obscured by scree, although the superimposition of one over the other is apparent.

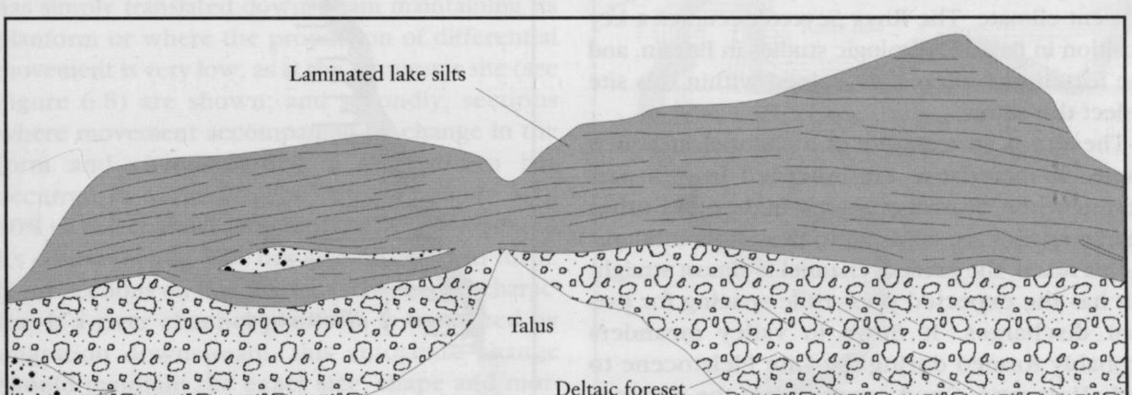

Figure 6.5 A section of lake sediments at Preston Montford. (After Jones, 1982.)

Interpretation

M.D. Jones (1982) inferred that a general reduction in sediment input took place over time. The section (Figure 6.5) is thought to have resulted from the creation of an ice-dammed lake, into which a small 'delta-moraine' developed. Such lake silts cover much of the Severn valley floor. The orientation of the silts and gravels indicates an ice source to the WNW.

At the most, a span of 5000 years covers the deglaciation of the Shropshire Plain; the development of the stream traces now recorded in valley meanders; the ingrowing of these meanders and the concomitant formation of the terraces; and the diminution of the streams to underfitness.

Conclusion

The River Severn at Montford is distinguished as part of the first British stream to be overtly recognized as of Osage-type underfitness. Detailed research has examined the discharge : channel form relationships of the existing and former Severn streams, and has contributed to the evidence explaining the development of underfitness in the deglaciation of the Shropshire Plain. These climatic events have impressed patterns upon the landscape that still survive, out of step with the present climate. The River Severn occupies a key position in palaeohydrologic studies in Britain, and the features of interest contained within this site reflect this status.

The site is an example of a channel in which some characteristics are inherited from a past period when discharges were higher and other characteristics are adjusted to present conditions. In this reach, the present channel is almost straight but has the expected pool–riffle spacing for present discharges. It flows in valley meanders probably formed during the Late Pleistocene to Early Holocene. Good exposures of the stratigraphy in the river terraces allow the chronology of events and interrelationships of processes in that period to be determined.

RIVER AXE AT AXMINSTER AND WHITFORD, DEVON (SY 268961 AND SY 287979)

Highlights

The River Axe provides an excellent example of contemporary river channel planform change for a lowland gravel-bedded river. Cartographic information and data from instrumented channel reaches illustrate how the downstream translation of meanders, changing sinuosity, and decreases in wavelength and in the radius of curvature can be observed.

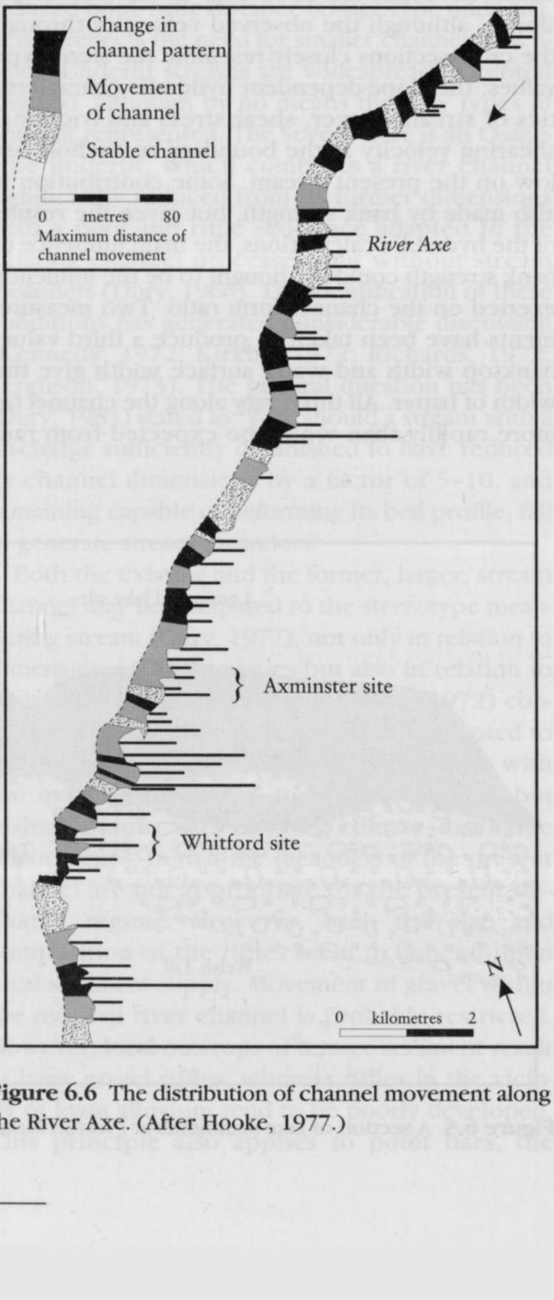

Figure 6.6 The distribution of channel movement along the River Axe. (After Hooke, 1977.)

Introduction

Knowledge of the extent, nature and distribution of changes in channel pattern in Britain was particularly sparse 25 years ago, and the general impression amongst geomorphologists has probably been one of the stability of streams. Since that time much work on channel changes has been carried out in Britain, and this is one of the sites at which detailed documentation and measurement has been undertaken. The migration of meander bends is a characteristic feature of alluvial rivers and such downstream translation of meanders is one of the most conspicuous changes affecting fluvial landscapes. Local changes may occur whenever and wherever streamflow exceeds the threshold for erosion of the materials composing the banks of the channel. It is necessary to know exactly how meanders migrate and change in floodplains typically dominated by agriculture, and subject to capital works within and close to the channel.

Description

From a study of over 400 bends on a number of rivers in Devon, Hooke (1977) found that 55% of the changes involved extension or translation, or their combination. In this study of the distribution and nature of changes in river channel patterns in Devon in a 100 year period, the River Axe illustrated types of distribution of channel movement. In Figure 6.6, firstly, sections where the channel has simply translated downstream maintaining its planform or where the proportion of differential movement is very low, as at the Axminster site (see Figure 6.8) are shown; and secondly, sections where movement accompanied by change in the form and characteristics of the pattern has occurred, as at the Whitford site (see Figure 6.7): 40% of the channel length of the Axe has altered its course, while 17% has also changed in form. Many sections on the Axe have changed in character. If a river channel planform is translated by migration downstream, this autogenic change should maintain the exact size, shape and morphology of the channel pattern. However, there can be changes of size, shape or of composition of the planform, in contrast to the relatively stable pattern of lowland channels characteristic of many rivers in lowland Britain. The Axe is a relatively active river that demonstrates two particular types of mobility particularly well.

Whitford Site

Here, the meanders are migrating down valley but lateral mobility of the river is constrained by the railway embankment (Figure 6.7). Meanders can be described as confined when they impinge against or are partly developed in media that alter the forms or rates of development from those found in local channel materials where such confinement is absent. In this case, maintenance and protection of the embankment prevents natural extension of the secondary lobe developed at the apex of the bend. Where confined, channel movements are typically dominated by down-valley migration. Bank erosion has been estimated as an average rate of 0.37 m yr^{-1} over a two-year period and a maximum average rate of 0.96 m yr^{-1}; this compares with a cartographic estimate of 0.54 m yr^{-1} for the same reach over a longer time period.

Axminster Site

Near Axminster, individual meanders are developing *in situ* through lateral growth in an unconstrained situation (Figure 6.8). Unconfined meandering rivers display a degree of morphological regularity, e.g. meander wavelength is correlated with discharge, frequently being between 10 and 14 channel widths. Indeed, meander width, length, mean radius of curvature and channel width, from a wide variety of locations

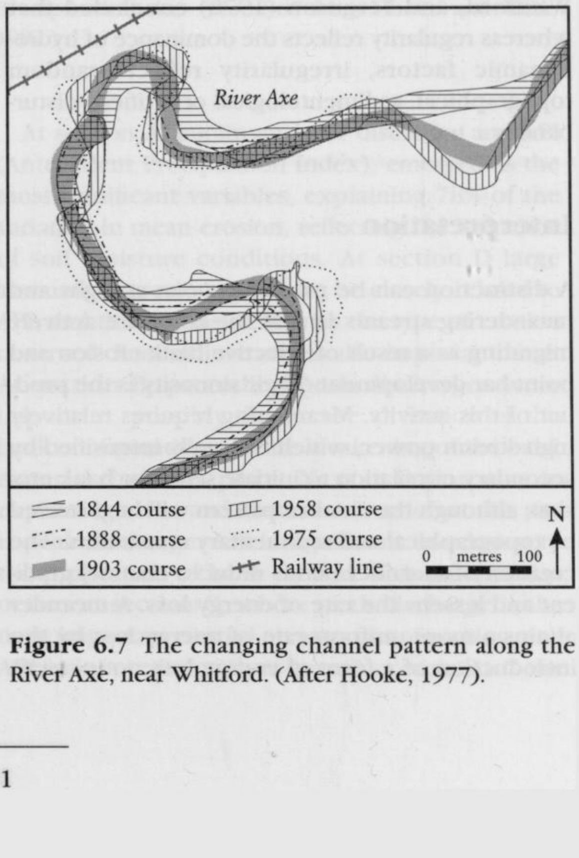

Figure 6.7 The changing channel pattern along the River Axe, near Whitford. (After Hooke, 1977).

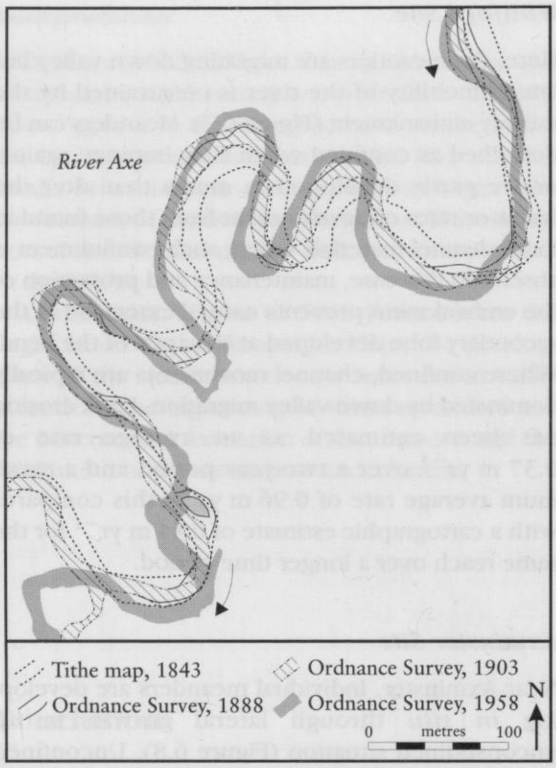

Figure 6.8 The changing channel pattern along the River Axe, near Axminster. (After Hooke, 1977.)

including alluvial rivers and supra-glacial streams, are closely interrelated. However, many meandering rivers are made up of a complex sequence of irregular and compound bends, such as the one at Whitford, and Ferguson (1976) concluded that whereas regularity reflects the dominance of hydrodynamic factors, irregularity reflects random topographical, sedimentological or artificial disturbances.

Interpretation

A distinction can be made between straight and meandering streams in that the latter are actively migrating as a result of selective bank erosion and point bar development. Their sinuosity is the product of this activity. Meandering requires relatively high stream power, which is usually intensified by secondary circulation to initiate selective bank erosion, although the detailed pattern will be distorted by topographical and sedimentary constraints. The creation of meander bends reduces channel gradient and lessens the rate of energy loss. A meander attains a more uniform rate of energy loss by the introduction of a form of energy loss not present in a straight channel, since along a curved path work is required to change the direction of flowing water.

Comparison of these two reaches — one confined and one unconfined — in both of which rates of erosion are relatively high and pattern changes over the past 150 years are quite large, allows variation in the mode and mechanisms of meander change to be examined. Together, the two sites clearly exemplify important features of lowland stream mobility and the effects of one fundamental control on river planform changes. Some examples of the impact of direct interference have been demonstrated on this river, but much of the channel is naturally mobile. The rivers in Devon are generally becoming more sinuous, moving by extension of bends, and decreasing in wavelength and radius of curvature, and sites such as these are of significance in understanding the effects of this development on the landscape, much of which is dominated by agriculture.

Conclusion

This is a site on an active and mobile, lowland meandering river in which historical changes have been documented and processes of bank erosion measured. Two reaches within the site show contrasts in degree of confinement and in change in meander characteristics.

RIVER EXE AT BRAMPFORD SPEKE, DEVON (SX 930986)

Highlights

The River Exe at Brampford Speke illustrates a case in which the dominant channel processes are controlled by river flows that have a return period of once every 2–5 years. Flows of larger magnitude and of higher return period have less erosive impact as floodplain inundation occurs, dissipating excess energy. The site also exhibits river-channel planform change.

Introduction

Bank erosion occurs under a range of discharge conditions but few systematic studies of bank erosion have been made in which a comprehensive set of factors likely to control erosion has been analysed (Wolman, 1959; Twidale, 1964; Walker and Arnborg, 1966; Harrison, 1970). Most studies of river-channel adjustment have stressed the importance of the mean annual flood (Dury, 1961), but Harvey (1975) remarked on the effectiveness of intermediate discharges.

This reach of the River Exe is important in illustrating two mechanisms of floodplain development, both influenced by small-scale, high-frequency discharge events. The floodplain has developed firstly through lateral meander movement and channel accretion, and secondly by overbank sedimentation, including that in abandoned palaeochannels. Detailed study of one compound meander bend has shown that maximum erosion occurred during discharge events with a recurrence interval of 2–5 years. Evidence of large-scale flood erosion is absent.

Description

Hooke's (1977) study of the site was designed to relate the amounts of erosion measured to the meteorological and hydrological conditions in the intervening period. The factors considered in this study were as follows:

1. Parameters of river discharge conditions and hydrograph characteristics;
2. Storm rainfall characteristics;
3. Time between peak flows;
4. Soil moisture conditions;
5. Temperature, especially frost.

At the meander site the drainage area is 620 km^2, and the channel width is 30 m. Here the gravel bed-channel is wide and flows through an open floodplain. This site includes some of the more active reaches available, and exhibits processes and conditions of channel adjustment more clearly and frequently than other reaches of the Exe.

Four sections were instrumented on a compound meander bend (Figure 6.9) which has developed a secondary lobe, a common feature on channels of Devon rivers. The four sections were as follows: A, a long straight section with a low bank and gravel layer at the base; B, a stretch with irregular plan profile due to embayments gouged out by the river; C, a long straight section experiencing a high rate of bank erosion; D, the inside of a sharp bend where erosion takes place at high flows.

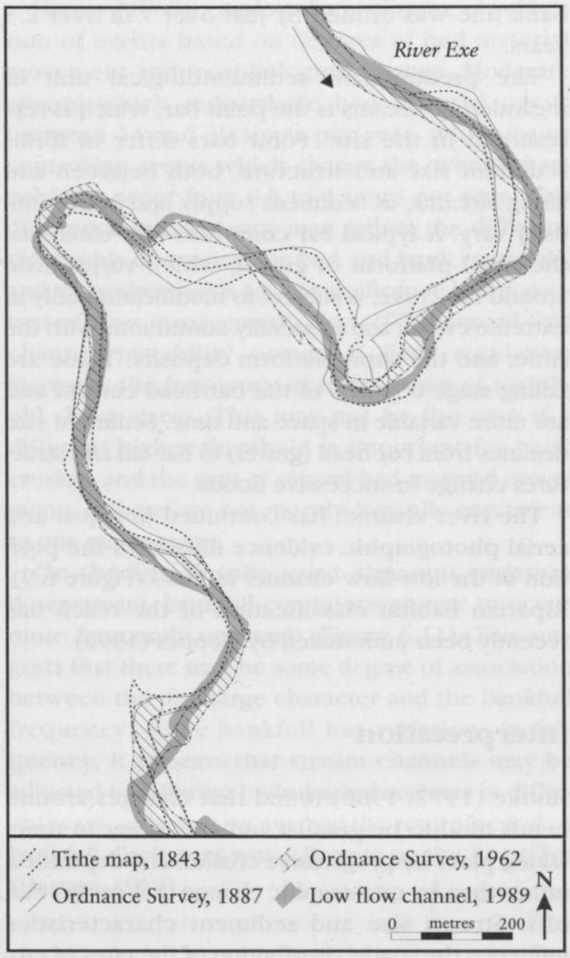

Figure 6.9 River channel change on the River Exe at Brampford Speke. (Modified from Hooke and Kain, 1982.)

At section A, minimum prior discharge and API (Antecedent Precipitation Index), emerged as the most significant variables, explaining 78% of the variation in mean erosion, reflecting the influence of soil moisture conditions. At section D large amounts of erosion appear to be closely related to API, and this is consistent with the type of erosion observed. Rainfall intensity is also important here. At section B, preceding peak discharges were important, possibly due to the alteration of the bank configuration and the weakening of the bank material that took place. Next in importance in explaining mean and maximum erosion at this site is peak discharge. API is significant in relation to the proportion of bank eroded. This pattern also occurs at section C, where peak discharge is the most important control on amounts of erosion, and API influences exactly where it occurs. At C the bank line was eroded by just over 7 m over 2.5 years.

The predominant sedimentological unit in meandering streams is the point bar, which is represented in the site. Point bars differ in form, sediment size and structure, both between and along streams, as sediment supply and bend sinuosity vary. A typical bar comprises two elements: the basal platform of gravel, which varies little around the curve, is subject to modification only in extreme events and is laterally continuous with the riffle; and the supraplatform deposits. These are falling stage deposits of the bar head current and are more variable in space and time. Sediment size declines from bar head (gravel) to bar tail and structures change in successive floods.

The river channel has continued to adjust and aerial photographic evidence illustrates the position of the low-flow channel in 1989 (Figure 6.9). Riparian habitat classification of the reach has recently been undertaken by Hooper (1992).

Interpretation

Hooke (1977, 1980) found that changes around bends tend to be gradual and consistent in time, taking place by progressive erosion and deposition rather than by catastrophic change that causes cut-offs. Stream size and sediment characteristics influence the spatial distribution of the rates of erosion, although not generally the relative amount of erosion in each event. Events on the Exe in which the maximum amounts of erosion occurred (including bank retreat of 3 m at one point) had a 2.5 year return period. The characteristics of the site are important. The magnitude of the erosion rate is most closely related to catchment area (as a surrogate of discharge and width). Two types of erosion were identified: firstly corrasion, influenced by river flow level; and secondly slumping, influenced by antecedent precipitation.

This site is an example of an active piedmont stream in which change in river meanders is progressive. Erosion and deposition take place during events of high frequency. It is a site at which such processes have been monitored in detail, and have been combined with an historical perspective on planform change.

River bank erosion is geomorphologically important as an integral part of changes in the river-channel course and in the development of the floodplain. Also instrumental in floodplain development is the concurrent growth of point bars. Together, these are amongst the most dynamic elements of the fluvial landscape, and thus an understanding of such processes is fundamental to our explanation of the development of fluvial features. As examples detailed in the research literature, the features of this site are significant in this context, and of value in the economic implications for agriculture and the construction of capital works in or near to rivers (Figure 6.10).

Conclusions

The Exe at Brampford Speke comprises a compound meander in an open agricultural floodplain,

Figure 6.10 A temporary sign alongside the Exe. 'This site is part of New Sowdens Farm and is situated in the middle reaches of the Exe Valley, which is of particular geomorphological interest, with good examples of ox bow formations, hence its designation as a site of special scientific interest. Evidence of the former Exe Valley Railway can be seen which cuts through the site on an embankment and was in use until 1963'. (Photo: K.J. Gregory.)

in which the processes of bank erosion have been measured in detail. Historical changes in planform have also been documented. The research shows the dominance of frequent events of erosion and deposition and the progressive nature of change at such sites.

RIVER TER AT LYONS HALL, ESSEX (TL 738108)

Highlights

The River Ter, Essex, illustrates the relationships between river discharge and channel dimensions for a lowland river. The river channels in the Ter catchment have a size and riffle spacing adjusted to relatively frequent flows.

Introduction

This reach of the Ter is representative of a lowland stream with a distinctive, flashy flood regime. Previous work on the frequency of bankfull discharge on alluvial streams in a variety of environments has suggested a recurrence interval for discharge at bankfull condition of somewhere between 1 and 3 years in the annual series. Variations in the frequency of bankfull discharge were found along the Ter in a study by Harvey (1969). Flood regime stream segments appear to exhibit adjustment to the 1–2 year flood but with less frequent flooding downstream. In this study by Harvey, the Ter was compared with the Nar (Norfolk) and the Wallop Brook (Hampshire) to facilitate understanding of the relationship between the size of river channel cross-sections and river discharge.

Description

The River Ter, Essex, drains a low-lying catchment (of area 77.8 km^2) to the gauging station at Crabbs Bridge (TL 876107) which has a cover of glacial till, and has a very low mean discharge but high flood peaks. Daily, monthly and annual discharge variability are also high. The site clearly demonstrates those characteristic features of a lowland stream, including pool–riffle sequences, bank erosion and bedload transport, which are adjusted to the particular pattern of flood frequency.

Harvey (1969), employed a qualitative classification of events based on degrees of bed material movement and morphological change. Moderate events which redistribute bed material occur between 14 and 30 times per year, while major controlling events which change the overall channel form occur from 0.5 to 4 times per year. This difference in frequency may reflect the different thresholds of erosion for bed and bank materials, and these thresholds are of significance in the context of river management. Hey (1976) argued that channel instability occurs if flow regulation increases the frequency of exceedance of threshold shear stress. This may not be the case if a different higher threshold is important for bank erosion, and the sum of several bed material movements is therefore not morphologically equivalent to one major event.

On the Ter, despite great apparent variation downstream, bankfull conditions appear to occur more frequently upstream (Figure 6.11). This suggests that there may be some degree of association between the discharge character and the bankfull frequency. Since bankfull has variations in frequency, it appears that stream channels may be adjusted to differing hydrological regimes in different ways, although on average the return period of bankfull discharges was 1.8 years on the River Ter (Harvey, 1969).

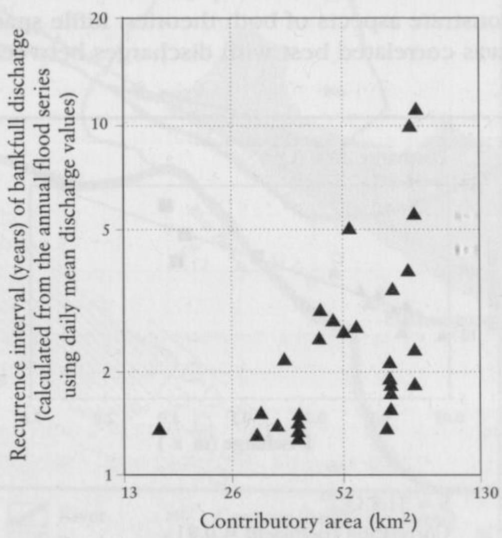

Figure 6.11 The River Ter, showing relationship between the bankfull recurrence interval and the catchment area. (After Harvey, 1969.)

Interpretation

It is an over-simplification to assume a constant return period for bankfull discharge. Harvey (1969) attributed the systematic variation in bankfull frequency to hydrological regime. Flashy streams experience more frequent overbank flooding upstream, which could be related to a downstream increase in flood duration.

It is important to understand the relationship between channel geometry and river discharge, because measurements of channel cross-section area (channel capacity) and channel width can be employed to give estimates of flood frequency at ungauged sites (Wharton *et al.*, 1988). These equations were developed from extensive fieldwork on UK rivers, and allow estimates to be made of discharges at a range of return periods from morphological measurements. Caution may be required when using these channel geometry equations, especially in defining bankfull capacity and in ensuring that the chosen site has a natural cross-section.

The Ter has also been used by Harvey (1975) to test the theories of Carlston (1965) and of Leopold and Wolman (1957, 1960). Leopold and Wolman (1957, 1960) argued that the meander wavelength was related more to channel width than to bankfull discharge while Carlston (1965) countered that, although bankfull discharge was poorly correlated, flows of greater frequency were more significant. By using median riffle spacing as a surrogate for wavelength, Harvey (1975) was able to demonstrate aspects of both theories. Riffle spacing was correlated best with discharges between the mean annual flood and the more frequently occurring Q_{20} (the discharge at which flows are exceeded 20% of the time) (Figure 6.12(a)). However, by assessing the surface water widths at Q_{20} and other flows, even stronger correlation coefficients were obtained (Figure 6.12(b)).

This reach of the Ter is important in the context of understanding how channel geometry is adjusted to discharge. The complexity of this relationship is illustrated by the systematic downstream variation in bankfull frequency on this stream, which exhibits typical features of many lowland streams.

Conclusion

This is a lowland stream in which the relationships between channel form and discharge have been analysed. It is a catchment with a variable discharge regime but relatively rapid response of flows to rainfall. The channel size and characteristics are adjusted to a relatively frequent discharge.

RIVER DERWENT AT HATHERSAGE, DERBYSHIRE (SK 209822 – SK 227807)

Highlights

The River Derwent at Hathersage illustrates the effects of reservoir impoundments on downstream channel morphology. The reduction in discharge

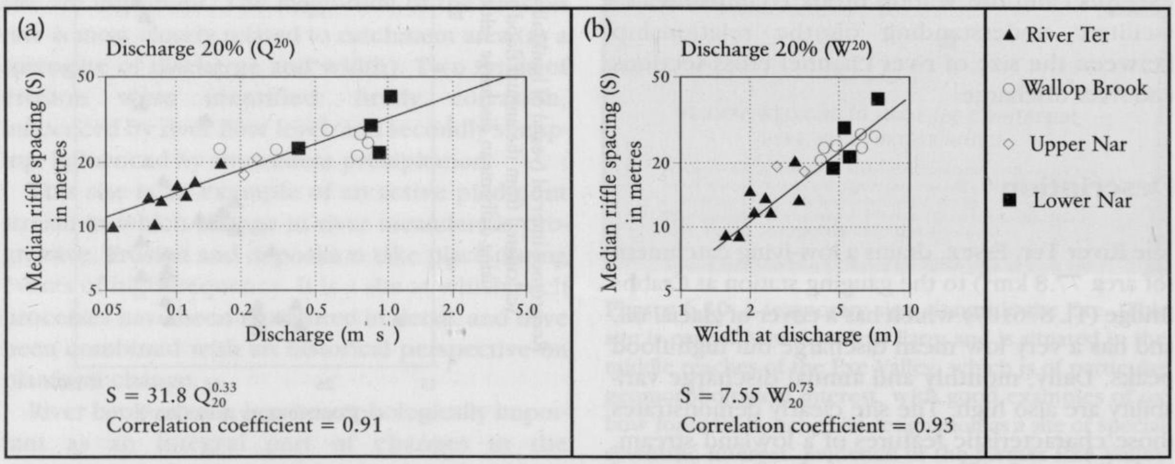

Figure 6.12 The River Ter, showing (a) the correlation of discharge frequency (Q_{20}) to median riffle spacing, and (b) the correlation of channel width at Q_{20} to median riffle spacing. (After Harvey, 1975.)

caused by the flow regulation of Ladybower Reservoir has produced a morphological response in the form of a low-flow bench within the existing channel, although this is only when a downstream sediment supply has become available.

Introduction

This reach of the Derbyshire Derwent flows between terrace remnants of two of the four lower terrace levels (Figure 6.13) identified in the Derwent–Wye catchment (Waters and Johnson, 1958), representing stages in the landscape development of the area. Along the Derwent near Hathersage is a type example of downstream channel adjustment to headwater impoundment, discharge regulation and sediment load reduction. The dominant adjustment of river channels to these changes is demonstrably a reduction in channel capacity (Petts, 1977). Between 4 and 8 km below the Ladybower dam, a flat bench has developed within the main channel, reducing channel capacity by approximately 40%. This adjustment becomes evident only below the river's confluence with the unregulated River Noe, because sediment from this tributary is probably the most significant factor controlling bench formation (Petts, 1977).

Description

The terraces

Erosion benches or flats are discernible throughout these Derbyshire river valleys at different heights above existing floodplains. Attention on the Derbyshire Derwent has been directed to the four better preserved terraces within the valley bottom (Waters and Johnson, 1958). Two are preserved in this part of the valley.

The Hathersage terrace is the oldest of the four, and is preserved around Hathersage. The rejuvenation head of this terrace on the Derwent is under

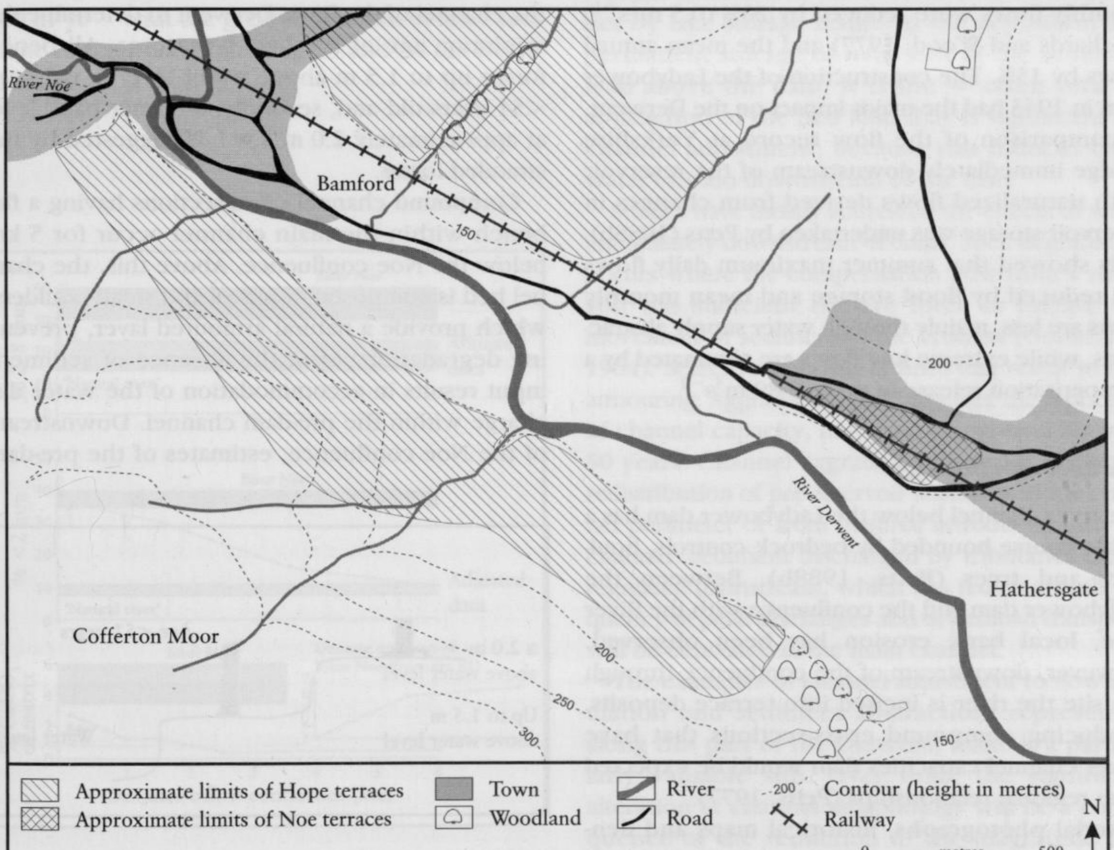

Figure 6.13 The River Derwent at Hathersage: the distribution of river terraces.

the Ladybower Reservoir, and is not determined on the Noe, where it is near the head of the next lower terrace, the Hope. The Hope terrace is also represented on this stretch of the Derwent, where it is being actively undercut by the river. The Hathersage and Hope terraces are composite in places, but the discontinuous upper segments of these two terraces cannot be traced up valley to independent knickpoints on the streams. It would appear that in each case the rejuvenation head responsible for the formation of the lower segment has overrun that related to the upper part of the terrace. The justification for the distinction is in the clear and ubiquitous distinction between upper or lower members of one terrace and representatives of the next, higher or lower, terrace.

Hydrology

The regulation of the Derwent headwaters was authorized in 1899 and involved building the Howden and Derwent Reservoirs, which were completed in 1912 and 1914 respectively (Petts, 1988b). Hence between 1913 and 1933 the gauged monthly flows were reduced by 20% (0.5 m^3s^{-1}) (Richards and Wood, 1977) and the mean annual flows by 15%. The construction of the Ladybower dam in 1943 had the major impact on the Derwent. A comparison of the flow record at Yorkshire Bridge immediately downstream of the reservoir with naturalized flows derived from changes in reservoir storage was undertaken by Petts (1988b). This showed that summer maximum daily flows are reduced by flood storage and mean monthly flows are less, mainly through water supply abstractions, while extreme low flows are eliminated by a compensation release of up to 0.89 m^3s^{-1}.

Channel morphology

The river channel below the Ladybower dam has a stable course bounded by bedrock controls, boulders and trees (Petts, 1988b). Between the Ladybower dam and the confluence with the River Noe, local bank erosion has been observed. However, downstream of the confluence through the site the river is incised into terrace deposits, producing compound cross-sections that have lower channel capacities than would be expected from regional relationships (Petts, 1977).

Aerial photographs, historical maps and dendrochronological evidence suggest that between 4 and 8 km downstream from the Ladybower dam a flat bench, bounded by marked breaks of slope, has developed within the main channel of the Derwent, overtopped at high flows (Figure 6.14). Sampling from the compound sections shows that the depositional benches are formed of coarse sand and gravels, often with a coating of finer sand and silt, contrasting with the terrace materials, which have between 16% and 45% silt clay. The terrace deposits are similar to the bank materials exposed in simple sections both upstream and downstream. The benches, however, are not always depositional; a certain amount of bank collapse is evident, and on occasion this material has become stabilized and incorporated into a bench form (Petts, 1977).

Results of bankfull discharge calculations indicate that a close similarity exists between the calculated bankfull discharge and the 1.5 year flood for the reach. However, sedimentological and vegetational evidence suggest that this channel, having a compound section, is adjusted to a more frequent event, rather than having had insufficient time to reach equilibrium with the new flow regime (Petts, 1977).

The lack of historical records for dating deposits was overcome by dating the numerous trees lining the channel of the River Derwent to determine the minimum age of the bench features. All bench forms up to 1.5 m above water level (a.w.l.) are <51 years old and, secondly, a former bank level at approximately 2.0 m a.w.l. is suggested by the stranded forms.

Compound channel cross-sections having a flat bench within the main channel occur for 5 km below the Noe confluence. Above this, the channel bed is composed of gravel and small boulders, which provide a natural armoured layer, preventing degradation. Also, the absence of sediment input results in accommodation of the water discharge within the pre-dam channel. Downstream of the Noe confluence, estimates of the pre-dam

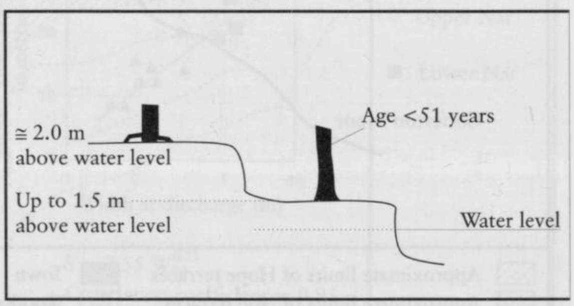

Figure 6.14 A schematic diagram indicating bench formation and the relationship between water level and tree growth. (Derived from data in Petts, 1977.)

channel form indicate that the deposition of a bench has reduced the channel capacity by nearly 40%. The bench is not evident until the tributary has joined the main stream, which implies that the introduction of sediment by the river Noe into the regulated main stream may be the significant factor controlling bench formation. The post-dam channel, adjusted to a flow event of greater frequency than the 1.5 year flood, may transport sediment more efficiently at moderate flows.

Bulk sediment samples taken along the river to just below the Wye confluence (Petts, 1988a) indicate that the D95 (where 95% of the sample is smaller than this value) decreases in size downstream from 47.8 mm on the River Ashope to 47.2 mm on the Derwent at Hathersage and 33.8 at Darley Dale. Freeze coring suggests that the River Noe is an important source of fine sediment, containing five times more material finer than 2 mm than the upper Derwent (Petts, 1988a,b). The influence of the River Noe sediments, indicated by the proportion of limestone in the sample, extends downstream for 3 km (Figure 6.15). The rate of gravel-bed siltation was estimated to be 500 $m^2 yr^{-1}$ with the greatest amount of finer sediment concentrated between 15 and 30 cm below the surface.

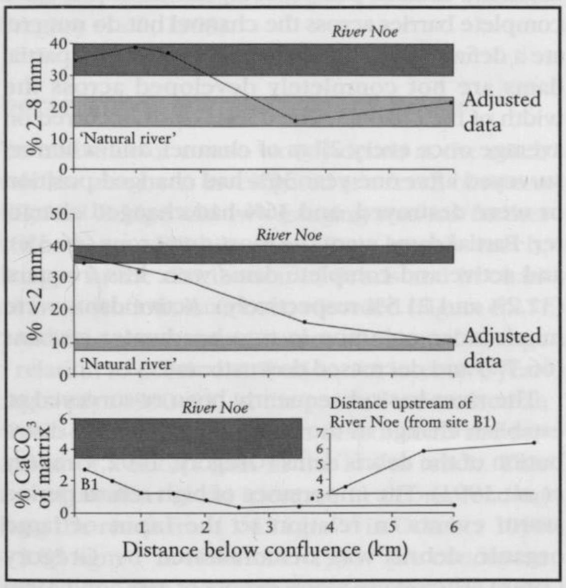

Figure 6.15 The influence of River Noe sediments on channel substrate downstream of the Noe confluence. (After Petts, 1988b.)

Interpretation

The morphology of a river channel reflects the river discharge that flows through the channel, the sediment that is transported by the flow, and the local characteristics, including bed and bank materials, vegetation and slope along the channel. The relation between form and process is an equilibrium or quasi-equilibrium relationship. When changes affect the drainage basin or the river channel upstream of a reach of a particular river channel, then the morphology of the river channel may change. This change can be expressed through adjustment of the degrees of freedom that the river channel possesses. Langbein (1964) suggested five degrees of freedom which may adjust to maintain quasi-equilibrium; namely, roughness, slope, width, depth and planform.

Channel changes can be induced by a number of causes, including the construction of dams and reservoirs. The retention and gradual release of water by a reservoir results in the reduction of peak discharges and the regulation of the flow regime. In addition, the reduction of flow velocities by lake storage results in the deposition and permanent storage of over 95% of the sediment load above the dam. It is the — often total — abstraction of the bed load that is significant for channel adjustment, because this induces clear water erosion downstream of the dam.

Studies have mostly addressed the effects of scour immediately downstream of dams. Such degradation occurs where the compensation flow from a reservoir has sufficient tractive force to initiate the movement of sediment in the channel (Gottschalk, 1964). Selective removal of fines can result in bed armouring. Aggradation, and therefore the reduction of channel capacity, has been recognized for over 50 years. Channel aggradation may occur by the redistribution of pre-reservoir sediment in the channel perimeter or from material introduced into the channel. Sediment discharged by tributaries into a regulated mainstream, which has reduced both frequency of peak discharges and of bedload transport, will be deposited in the main channel.

The examples of channel adjustment to flow regulation and sediment abstraction, represented along this part of the Derwent, make it a particularly valuable site. Over time, the significant alteration of channel morphology will be a consequence of the reduction of the magnitude and frequency distribution of flows and the reduction of sediment load to below-dam reaches. The reduction of channel capacity may adversely affect the efficiency of floodwater transmission downstream. However, although the more frequent floods are markedly reduced, the effect upon rarer events may be negligible. The successful operation of flow regulation schemes, necessary for both flood control and water supply, requires study of the range of geomorphological consequences, one of which is exemplified by this reach of the Derwent at Hathersage. The Hope terrace fragment incorporated is important in its wider context of the suite of terraces that result from the deglacial history of the Derwent valley system. It is also important because the inheritance of that development in the present river is exemplified by the undercutting and reworking of material from the terrace by the regulated and adjusting River Derwent.

Conclusion

This reach of the River Derwent has been affected by construction of reservoirs in the catchment. The reservoirs have reduced peak discharges and have caused channel adjustment downstream. The main feature created has been a low bench along the channel, but this is only present downstream of tributaries, because the latter provided the sediment for the bench formation.

HIGHLAND WATER, HAMPSHIRE (SU 272073–SU 239912)

Highlights

The Highland Water provides an excellent example of the influence of vegetation on river-channel processes in the heathland and forestry context of the New Forest, Hampshire. The presence of coarse woody debris in the river channel is of particular importance in relation to the channel morphology, hydrology and ecology of the catchment, while types of heathland influence the character and amount of runoff.

Introduction

The site has been investigated to establish how the influence of vegetation relates to the processes operating in river channels. This is important for understanding the routing of flood peaks through a fluvial system and the modelling of channel processes at different stages of flow; for understanding patterns of channel and floodplain sedimentation; for the enhanced interpretation of palaeohydrological evidence; and in relation to effective channel management (Gregory, 1992; Gregory and Davis, 1992).

Description

Coarse woody debris: Highland Water

In river channels bordered by trees or flowing through woodland, the presence of organic material in the channels is important in affecting fluvial processes. In these areas trees and branches may fall into the river system through storm events, bank erosion or tree fatality. This coarse woody debris may be redistributed by the river into discrete accumulations known as debris dams; leaf litter may also be stored in these accumulations (Figure 6.16).

The Highland Water drainage area is 11.4 km^2 above the two gauging stations just below Millyford Bridge. A number of studies have focused on the role of organic debris in the river. A classification of debris accumulations was undertaken along the main channel by Gregory *et al.* (1985). Three types of debris dams were recognized: active dams (Figure 6.16) form a complete barrier to water and sediment movement and create a distinct step or fall in the channel profile; complete dams provide a complete barrier across the channel but do not create a definite step in the channel profile; and partial dams are not completely developed across the width of the channel. The debris dams occurred on average once every 27 m of channel, and when resurveyed after one year 36% had changed position or were destroyed, and 36% had changed character. Partial dams were the most numerous (46.3%), and active and complete dams were less frequent (32.2% and 21.5% respectively). Active dams were much more common in two headwater streams (66.7%) and decreased downstream.

The river has subsequently been re-surveyed to establish change in the temporal and spatial distribution of the debris dams (Gregory, 1992; Gregory *et al.*, 1993). The importance of high return period storm events in relation to the input of large organic debris was demonstrated by Gregory (1992). The severe storm events of 1987 and 1990 caused a high degree of blowdown and injected large amounts of debris into the system. A survey of the whole of the Lymington basin (110 km^2),

Figure 6.16 An active debris dam on the Highland Water, New Forest. (Photo: K.J. Gregory.)

which includes the Highland Water, reported debris dam densities of 1.15 per 100 m of channel (Gregory *et al.*, 1993).

Coarse woody debris: Millyford Bridge

The influence of coarse woody debris at the reach scale has also been investigated at a reach between Millyford Bridge and two gauging stations further downstream. Within the 78 m length reach, changes in the distribution of debris and in channel cross-sections have been recorded (Figure 6.17) to indicate the dynamic nature of channel change in relation to debris distribution. An intensive year-long study (1990–91) in the reach documented an increase of debris loading in response to the formation of a debris dam in the lower reach. This was equivalent to an increase in debris loading along the reach from 1.42 kgm^{-2} in October 1990 to 3.85 kgm^{-2} in November 1991. The conditions created by this dam have redirected the flow involving the development of a subsurface cavity after the washout of gravels, which is succeeded by slow subsidence (Davis and Gregory, 1994). This type of bank erosion may occur elsewhere in channels in temperate woodlands.

Withybed subcatchment

The Withybed subcatchment encloses the eastern part of the heathland zone of the Highland Water catchment. It has been surveyed in detail (Figure 6.18) to establish the interrelation of vegetation and groundwater (Gurnell and Gregory, 1984; Gurnell *et al.*, 1985). In contrast to the rest of the catchment, the area contains areas of high soil moisture content. Transects were selected to reflect the range of heath categories in the soil moisture vegetation map produced according to soil moisture variations.

The position of areas of high soil moisture content, such as the moist heath category, in relation to the main drainage network is an important factor determining runoff production. These areas are often quite remote from the main drainage network and so may produce a complex response to different temporal and spatial patterns of rainfall. Such a relationship is of considerable

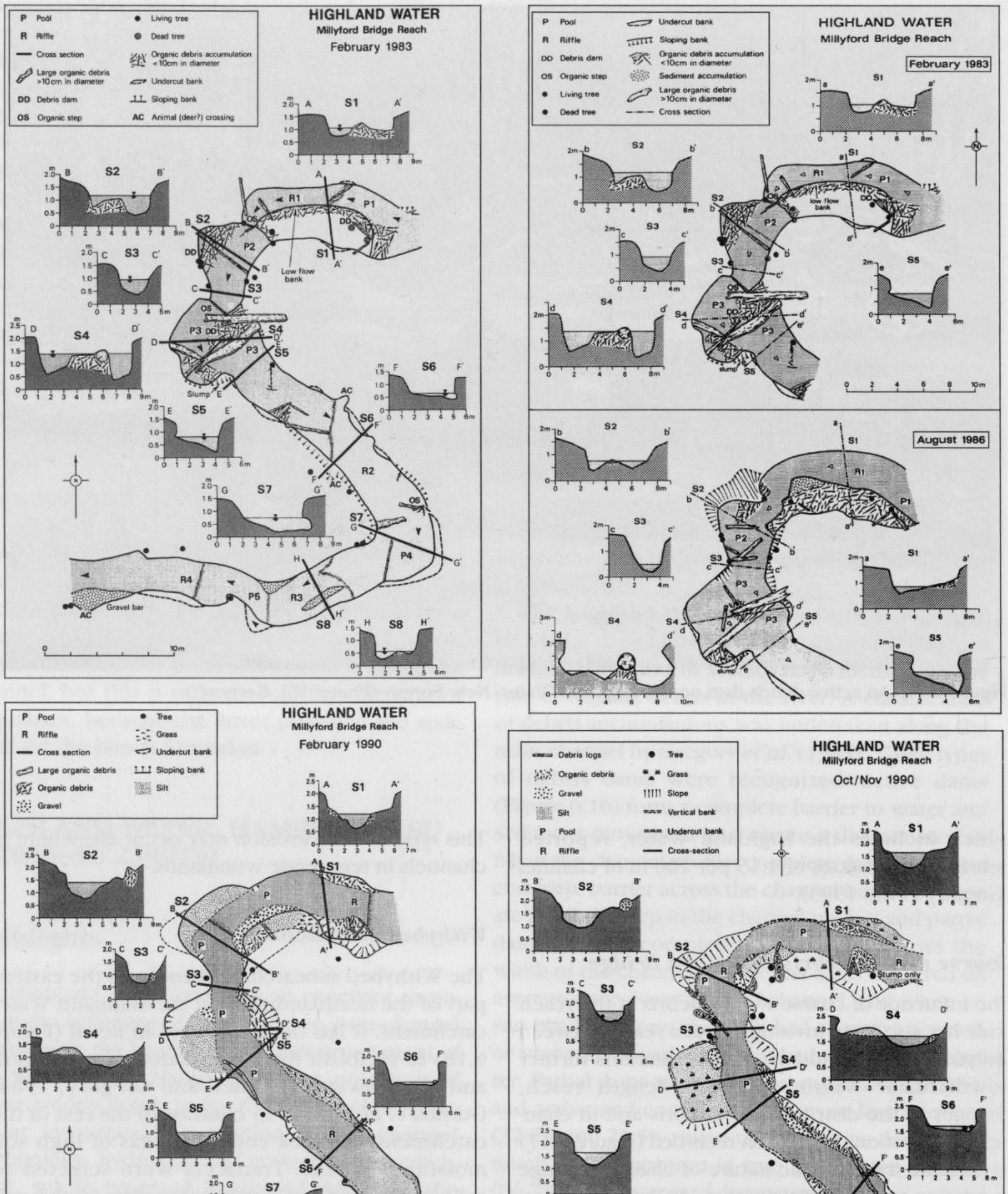

Figure 6.17 Millyford Bridge reach, illustrating changes in debris location and channel morphology.

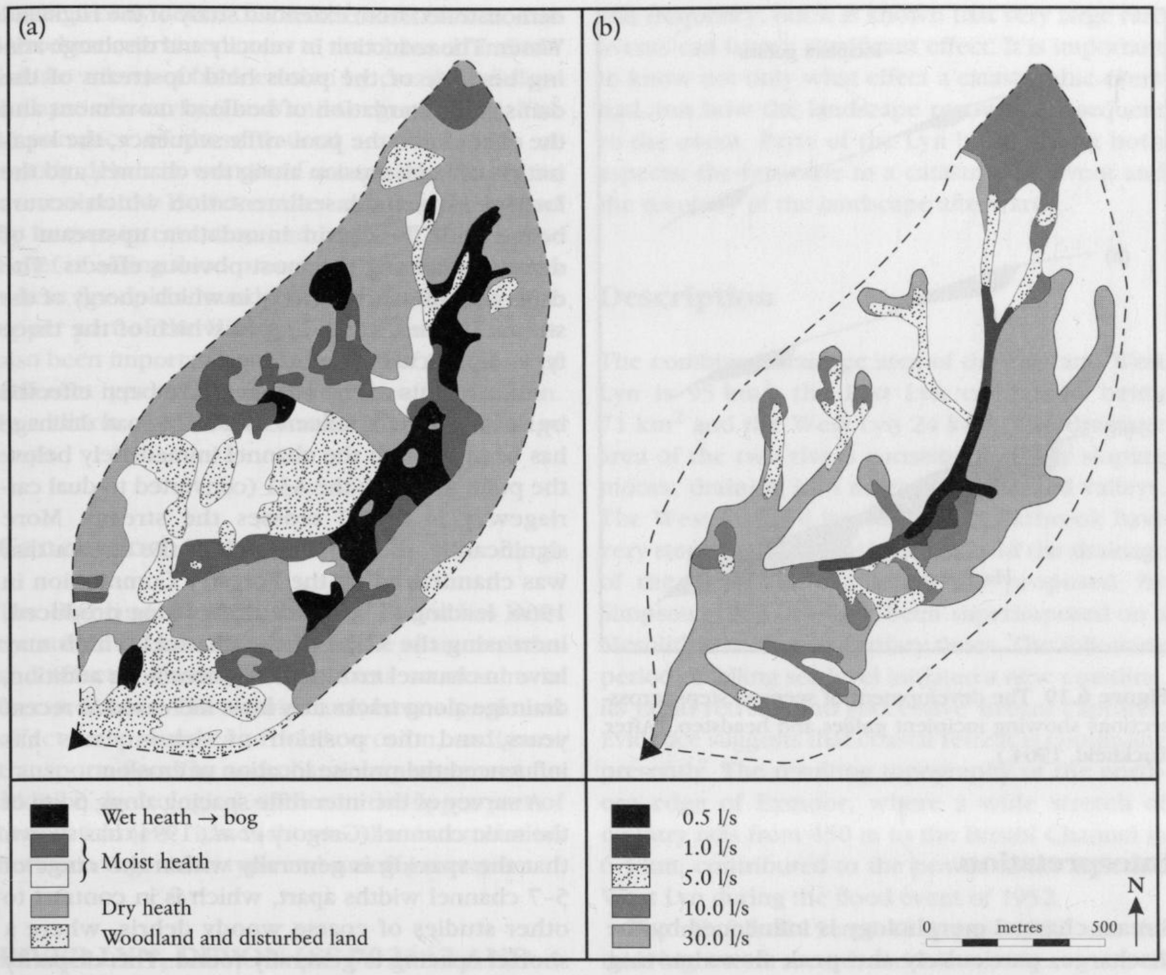

Figure 6.18 The distribution of vegetation — soil moisture categories (a) and runoff contributing areas at different levels of baseflow (b) in the Withybed subcatchment. (After Gurnell *et al.*, 1985.)

significance for linking the character of the vegetation composition to the catchment runoff response.

Seepage steps

The heathland areas of the catchment are podsolized, and leaching has greatly reduced the cohesion of the sandy matrix of the gravel and so decreased its resistance to erosion once the vegetation cover is removed. Gullies have developed mainly in the gravel or other soliflucted material. The example incorporated in the site limits is on the south-west slope of Stoney Cross Plain, at SU 252107. More than one incipient gully typically forms on the same slope, where the local increase in gradient is greater than the general increase for that slope. In these cases, the separate gullies grow by deepening and headward erosion until they coalesce, and the junction between the two sections is marked by a prominent seepage step, often 0.75–1.00 m in height (Figure 6.19). These tend to persist, migrating slowly upslope.

The headward extension of a gully by headstep migration seems to be a very slow process, although headward growth can be achieved much more rapidly by the development of small incipient gullies immediately above the main one. The rapid longitudinal growth of gullies on hillslopes, up to 65 m in two years in this case (Tuckfield, 1964), is achieved by the coalescence of discontinuous sections rather than by headward erosion by upward step migration.

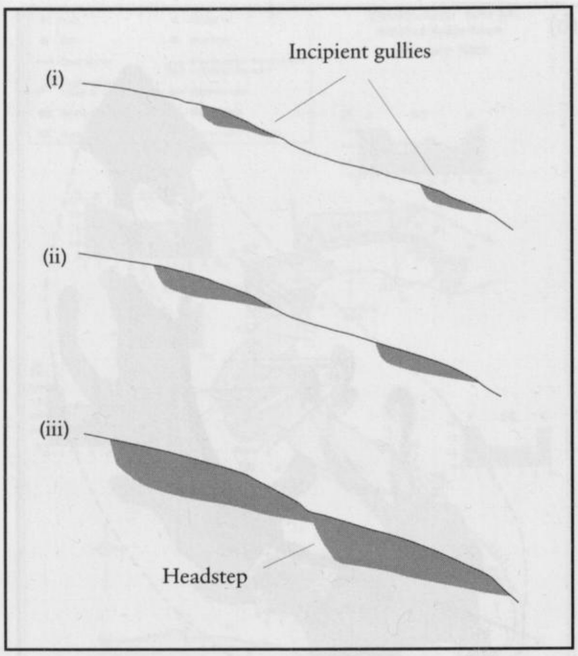

Figure 6.19 The development of seepage steps: cross-sections showing incipient gullies and headstep. (After Tuckfield, 1964.)

Interpretation

Stream channel morphology is influenced by the discharge, particularly the peak flows moving down the channel; the sediment that is transported; and the local channel-forming characteristics, including the vegetation. In lowland British stream channels the riparian vegetation is a significant influence determining the stream channel morphology. This is very well demonstrated along the course of the Highland Water. The stream channel processes are influenced by the material from blowdown and leaf litter fall that each year contribute to organic accumulations in the stream channel.

Debris dams are particularly significant in relation to the distribution of stream channel processes, and particularly in relation to the distribution of flooding. It was originally thought that flooding might take place extensively across the floodplain of a small stream or river, but it is now appreciated that the debris dams can lead to localization of flooding because one major dam causes flooding upstream, whereas immediately downstream of the dam the flood discharge is contained within the river channel.

A number of major implications have been demonstrated from extended study of the Highland Water. The reduction in velocity and discharge arising because of the pools held upstream of the dams, the retardation of bedload movement and the effect upon the pool–riffle sequence, the localization of bank erosion along the channel, and the localized overbank sedimentation which occurs because of floodplain inundation upstream of dams, are among the most obvious effects. The dams also influence the way in which energy of the stream is used, according to which of the three types a particular dam belongs.

Adjustments of the channel have been effected by three groups of influences. Firstly, road drainage has been fed into the channel immediately below the point at which the A31 (converted to dual carriageway in 1980) crosses the stream. More significantly, there is a stretch of the stream that was channelized by the Forestry Commission in 1966, leading to several cutoffs being produced, increasing the slope of the channel which may have in-channel erosion downstream. In addition, drainage along tracks may have increased in recent years, and the position of debris dams has influenced the precise location of erosion.

A survey of the inter-riffle spacing along 6 km of the main channel (Gregory *et al.*, 1994) has shown that the spacing is generally within the range of 5–7 channel widths apart, which is in contrast to other studies of coarse woody debris, where a shorter spacing is generally found. This disparity could be due to the human impact of stream clearance preventing the establishment of debris dams and also to the inherent instability of the smaller pieces of coarse woody debris.

The Highland Water is significant because of the way in which it reflects the fluvial processes that are active in the relationship between stream channel processes in a woodland catchment. The accumulations of coarse woody debris are particularly important in the way in which they control sediment storage and transport, the formation of pools and riffles, and the processing of organic matter, as well as increasing the diversity of the river channel. In addition, certain parts of the stream channel are showing adjustment to changing processes through the existence of enlarging channel cross-sections, which can be identified by undercut banks, bent trees and exposed roots, when they occur on opposite sides of the channel. In these cases this indicates that the channel is enlarging rather than simply shifting its position autogenically (Gregory, 1992). On the reach scale, it has recently been shown that accumulations of coarse woody debris can initiate bank erosion, can introduce coarse sediment to the channel, and can initiate a form of bank erosion by creation of slow collapse of a cavity (Davis and Gregory, 1994). The importance of these natural accumulations of woody debris in woodland ecosystems cannot be overestimated. However, they are often threatened by human intervention and removal, often in the belief of aiding flood prevention in a forest system where floodplain inundation occurs naturally. The upper part of the Highland Water catchment has also been important for assessment of the effect of different types of vegetation on runoff generation. In addition, the area has some interesting seepage steps.

Conclusion

The Highland Water in the New Forest has been the subject of intensive study of the characteristics and effects of woody debris dams on streams in forested areas. The debris dams have a profound effect on processes of bank erosion, sediment transport, flooding and channel enlargement. In addition, hydrological studies in the upper part of the catchment have shown the interrelation between vegetation type and runoff generation.

RIVER LYN, DEVON (SS 702442 AND SS 721492)

Highlights

The River Lyn, draining the northern slopes of Exmoor, north Devon, illustrates the impact that rare, high return period flood flows can have on the morphology of a steep catchment. The 1952 flood, with a return period of at least 150 years, transported massive amounts of sediment and boulders, the impact of which is still persistent in the present-day channel morphology.

Introduction

The Lyn catchment was the scene of a unique storm rainfall sequence and flood event in Britain, which demonstrated the effects of short-term processes and long-term changes in sea level on river development. Fluvial landforms have often been attributed to events of moderate magnitude and frequency, but it is known that very large rare events can have a significant effect. It is important to know not only what effect a catastrophic event had, but how the landscape recovers subsequent to the event. Parts of the Lyn basin reflect both aspects: the response to a catastrophic event and the recovery of the landscape afterwards.

Description

The combined drainage area of the East and West Lyn is 95 km^2, the East Lyn catchment being 71 km^2 and the West Lyn 24 km^2. The drainage area of the two rivers consists of gently sloping moors, draining into narrow steep-sided valleys. The West Lyn and its tributary at Barbrook have very steep profiles. The broad lines of the drainage of the Lyn river catchment are proposed, by Simpson (1953), to have been superimposed on a Mesolithic surface in Tertiary times. The following period of falling sea level initiated a new coastline, its rapid retreat, and successive stream captures. Evidence suggests that coastal retreat is continuing presently. The resulting topography of the northern edge of Exmoor, where a wide stretch of country falls from 450 m to the Bristol Channel in 6.4 km, contributed to the power of the East and West Lyn during the flood event of 1952.

The rainfall of 15 and 16 August 1952 was the third highest on record, and followed heavy rain on 13 of the previous 14 days. The rainfall observer at Longstone Barrow recorded 228 mm from 11.30 hours on the 15th to 09.00 hours on the 16th, with 178 mm estimated between 17.00 hours and midnight. Bleasdale and Douglas (1952) considered that significant percolation was minimal due to antecedent rainfall conditions, and subsoil drainage contributed a low proportion of the total runoff. Reports of eyewitnesses describe field slopes being 150–200 mm deep with surface runoff during the torrential periods. It seems probable that this happened in many parts of the Moor, with sheet flow over extensive surfaces. By the time the rain reached its greatest intensity, the rate of runoff was probably almost equivalent to the rate of rainfall, and had been maintained for some hours.

The upper site is in the upper Cannon Hill valley (Figure 6.20), where interference by man and animal has been minimal and features resulting from the event persist (Anderson and Calver, 1977). The volumes of water and sediment moved were enormous; 50 000 tonnes of boulders (Kidson, 1953), some more than 10 tonnes each (Delderfield,

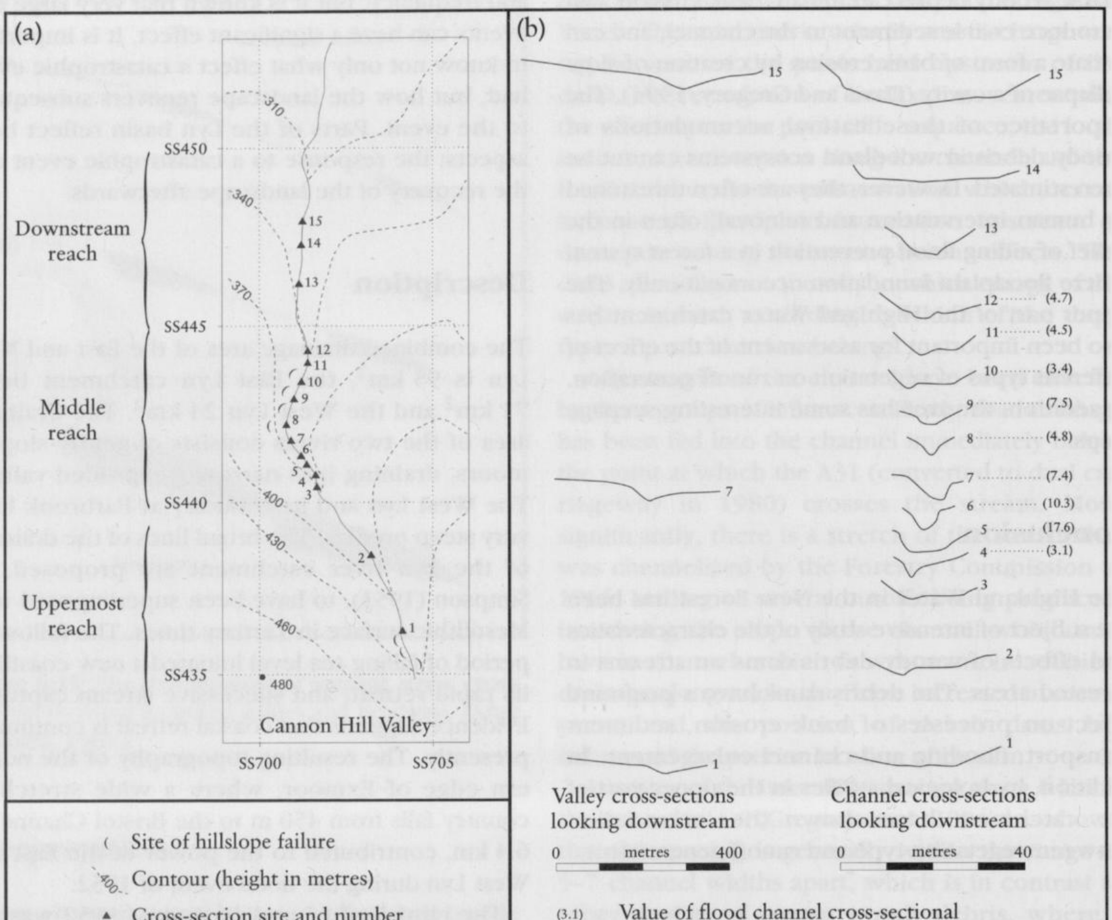

Figure 6.20 Topographical features of Cannon Hill Valley, Exmoor. (After Anderson and Calver, 1977.)

1981), were moved by a volume of water in the order of twice the average daily discharge of the Thames (Kidson, 1953). The lower site in Glen Lyn Gorge contains a suite of huge boulders lying where they came to rest, and demonstrating the magnitude and power of the flood event. If the event is measured according to the power expenditure per unit area of channel (Wm^{-2}), then the event had a greater unit power than the Mississippi or the Amazon in flood. The event was exacerbated by the bridges crossing the river. Debris accumulated upstream, causing temporary dams that gave even larger flood flows when they burst successively down the tributary valleys to Lynmouth.

Interpretation

In their study of the recovery of the landscape from the flood event, Anderson and Calver (1977) considered three sections (Figure 6.20): the upper, middle and lower reaches.

Upper valley reach

Before the event, the wide flat valley floor was apparently largely vegetated and contained several small channels incised into peat deposits (Figure 6.21). During the flood, these channels were overtopped and widespread overland flow occurred over most of the valley floor. The combination of dense vegetation and shallow overland flow effected little surface erosion. However, two shallow slides occurred, and the debris and extent of the slip are still in evidence.

Middle reach

The passage of the valley on to grits in this reach results in steeper valley sides (30–35° slope) which restrict channel width. The event formed a flood channel by removing bed material which was coarse in places. The high flood stage left continuous small bank-side scars which are still in evidence, although slumped in places. Vegetation has grown below the scars, revealing the reduced dimensions of the present channel. Channel infilling since 1952 seems to have been effected by discrete bank-side failures accompanied by slower, more continuous degradation of flood and slumped forms. Channel debris is minimal and in places bedrock forms the channel bed: any debris found is likely to be from slumping. Little evidence exists of channel pattern change during the flood, although severe undercutting has taken place.

Figure 6.21 A channel (flowing top right to bottom left) incised into peat deposits, Cannon Hill Valley. (Photo: R.J. Davis.)

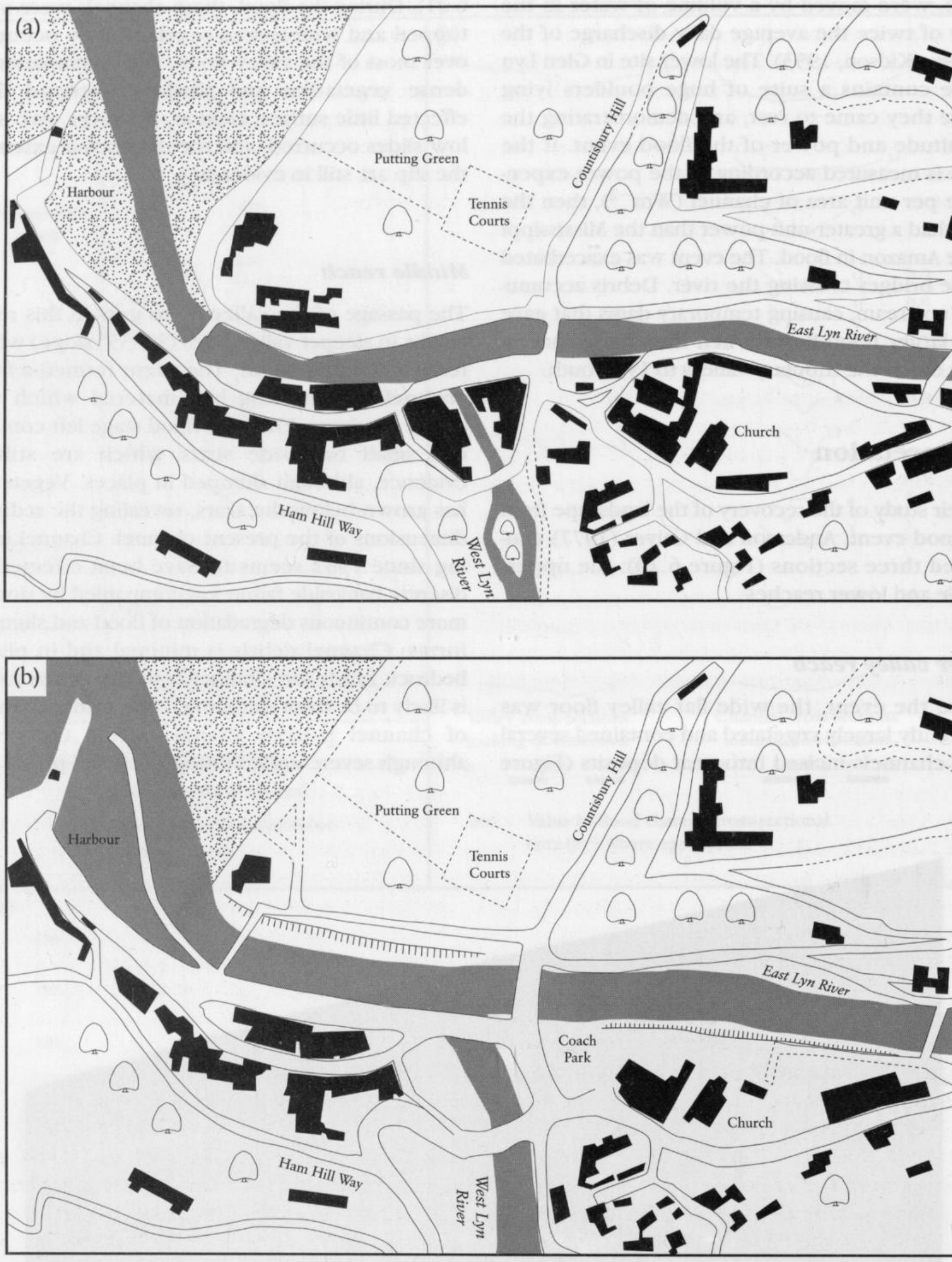

Figure 6.22 Channel changes on the River Lyn after the 1952 flood: plans of Lynmouth: (a) before the flood, (b) today. (After Dobbie and Wolf, 1953.)

Downstream reach

This reach has a similar wide valley floor and relatively shallow valley sides to the upper reach. The flood dispersed over the floor, giving rise to widespread deposition. The boulder fields deposited where the valley widens out during the flood are now largely covered with vegetation, but within them is evidence of former water courses. The receding flood revealed a discrete channel through the boulder fields, which has suffered some measure of subsequent infill by downhill movement of fines and small-scale slumping of the banks.

The contrast between the valleys of the lower and middle reaches arises from a recent strong rejuvenation which has extended only a limited way upstream. The lower reaches are V-shaped gorges, so narrow there is no room for a road, and Lynmouth has been built on the river bed. The middle reaches are relatively broad and open, and there is always a floodplain. The valley sides reveal a valley within a valley resulting from an earlier rejuvenation, more extensive than the recent one that created the lower reach gorges.

Under natural conditions, the erosional features of the 1952 flood have since degraded. Where erosion was dominant, the effects are still apparent. Depositional features are also very much in evidence and have, in general, suffered less degradation, although vegetation is obscuring their extent. The 150-year flood has produced some landscape modification that is likely to last at least for the mean recurrence interval of the event. However, the relationship of such events to landscape form is complicated by the fact that recurrence intervals vary, and that some of the persisting features are less recognizable than the overall aggradation or degradation of parts of the valley.

The floods in the Exmoor region, their effects in the Cannon Hill valley, and the suite of boulders in the Glen Lyn Gorge, re-emphasize both the nature of the processes by which rivers shape their valleys, and the significance of rare and powerful floods. It is important to have a record of the type of features that can be produced by such floods, and to know how long they survive in the landscape. The detailed investigations of such features and their subsequently monitored geomorphological significance make the upper site unusually valuable. The change of course of the West Lyn in Lynmouth (Figure 6.22), by the transport of immense quantities of boulders and smaller material, is a spectacular characteristic feature of valley development in the broader context of fluvial system adjustments to sea-level changes and coastal retreat.

Conclusion

This is the site of one of the most extreme flood events in recent British history. The large flow had devastating effects on the valley of the River Lyn, including the formation of slope scars and the movement of very large boulders. Not only are the effects of this flood, which vary in different parts of the river system, well recorded, but the long-term persistence of the changes has been assessed. This has become a classic site, exemplifying the significance of large events in small upland catchments.

RIVER ITCHEN NEAR KNIGHTCOTE, WARWICKSHIRE (SP 404558)

Highlights

The River Itchen at Knightcote represents an example of stream underfitness where the present channel cuts through the deposited alluvium of the former channel and is wholly contained within it. The site was the first of its kind to be distinguished.

Introduction

The Warwickshire Itchen was the first site to have been deliberately tested for the presence of a former channel, and has revealed more details than other sites. It is cut in impermeable alluvium underlain by impermeable solid rock, has not been extended or reduced by divide migration since the last glaciation, and is tectonically stable. The type features of winding valleys are well-developed throughout — large-scale sinuosity, steep slopes at the outside of valley bends, with gentle slopes on the inside, together with the small meanders of the existing stream.

Description

The Warwickshire Itchen is a tributary to the Leam which, in turn, flows into the Avon near Stratford. The valley was selected by Dury (1954) for investigation for its almost impermeable alluvium underlain by solid rock; for the fact that it has not been extended or reduced by divide separation since the last local glaciation; and because it is tectonically stable. In part the valley is incised and winding, while in part it is open and with low relief but still with valley windings (Figure 6.23).

The windings are best displayed where the valley is cut into a limestone shale series near the base of the Jurassic succession. The ribbon of alluvium

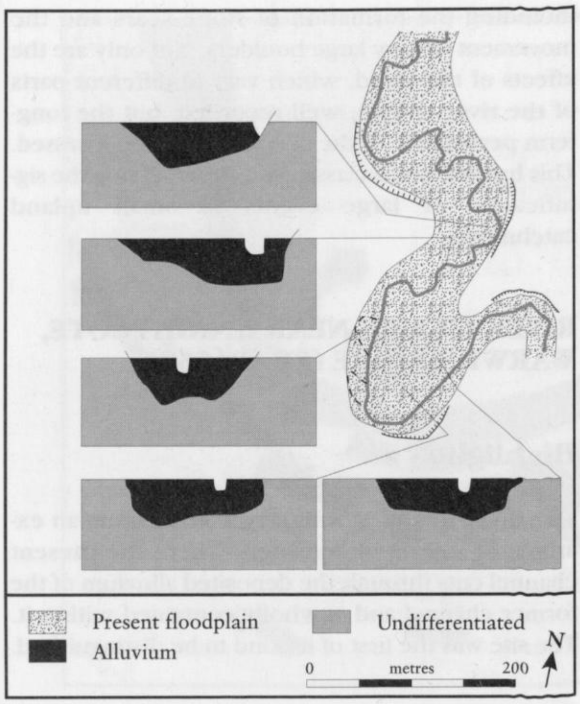

Figure 6.23 Alluvium-filled palaeochannels on the River Itchen, Warwickshire. (After Dury, 1954.)

that lines the valley bottom is for the most part sharply bounded by the valley walls. The type features of winding valleys are well-developed throughout; large-scale sinuosity, and steep slopes on the inside, together with the small meanders of the existing stream. The latter, as is commonly observed, are mostly not in contact with the valley walls at many points but instead are described in the alluvium.

It is important to note that three different reaches of the Itchen were investigated, necessarily at different distances from the present source of the stream. It was found that the dimensions of the infilled channel diminish headwards in about the same proportion as the dimensions of the existing channel. Hence the ratio between the respective widths of the two channels is the same throughout. The same ratio is given by a comparison of the radius of curvature of the valley bends and the existing stream meanders in a group of streams in the English Midlands.

First trial augurings in the valley of the Itchen revealed an abrupt transition from moist gleyed alluvium to dry shale or marl. The valley fills contain little material coarser than sand, consisting mainly of silt and clay, varied in places by peat, sapropel, tufa and malm (earthy, amorphous calcium carbonate).

On the Itchen, the sole consistent change in the vertical succession of alluvium is from dull-brown cloddy columnar or prismatic clayey silt at the top to moist, blue-grey, gleyed clayey silt in the lower part. When the first records were made, the significance of gleying was not fully taken into account, and the two types of deposit were recorded as distinct. Subsequently, it became clear that the difference is one between permanent waterlogging and seasonal waterlogging. In the absence of lithologic change it is impossible to define the base of the present floodplain, although the base of the large channel lies well below the maximum depth of scour of present conditions. The bed is not scoured severely enough to permit the stream to reach the base of the underlying filled channel; nor does the present channel widen sufficiently to provide the width to depth ratio which — on any reasonable view of channel form — would be needed if scour were to go down to the underlying bedrock.

Interpretation

The reduction of a stream to a manifestly underfit condition necessarily lengthens the trace and thus tends to reduce the slope. Lengthening must occur when stream meanders are added to the trace of valley meanders. To compensate fully for the lengthening, the vertical difference between source and mouth would need to increase in the same proportion. There is no scope for degradation at the downstream ends where levels are controlled by confluence or sea level, and all compensation would therefore need to be performed by infilling on the upper reaches. The depths of fill required by full compensation for lengthening of trace are great. However, no such deep fills are evident in the field. Indeed, the fills in some Cotswold head valleys are distinctly shallow. The fills on the Itchen increase in depth with distance downstream, and appear to do so regularly.

This stream was the first of a number which have been used to establish the widespread occurrence of underfit streams, to demonstrate that not all underfit streams need possess meandering channels at the present time, and to show that rearrangements of drainage do not constitute a general hypothesis to explain the facts of distribution and chronology of the stream shrinkage.

Subsurface exploration shows that manifestly underfit streams on the English Plain are characteristically underlain by large channels that wind round the bends of the valleys (e.g. Figure 6.25). They reach their greatest depths at or near the extremities of valley bends. The average ratio of width between the large channels and the present channels is 11.5 : 1, (Dury, 1954).

Records for the Itchen River valley include this site and three other reaches that have been used to make a graphic comparison of bed width and drainage area. These indicate that the large channels can be traced far up the valleys. They also suggest that the disparity between bedwidth increases headwards, as the disparity between wavelengths will later be seen to do. The bedwidth ratio falls from 15 : 1 at 3 miles to 9 : 1 at 40 miles.

Three periods are critical for consideration of the development of underfit streams:

1. initiation of large meanders or large channels;
2. the onset of underfitness and the abandonment of large meanders or channels; and
3. the duration between times of initiation and abandonment.

Detailed studies of fossil pollen in southern England not only serve to confirm that reduction to underfitness post-dates the last glacial maximum but also show that channelling and filling went on long after 10 000–11 000 BP. Infilling of some channels in southern England during Zone V is referable to increasing dryness. As the floristic record demonstrates, this was a time of increasing cold which necessarily reduced evapotranspiration. Therefore the reduction of precipitation was great enough not merely to compensate for the influence on runoff of reduced temperatures, but also to promote stream shrinkage.

Data confirm that wavelength varies with the square root of discharge and support the contention that bankfull discharge of manifestly underfit streams has been reduced in proportion to the square of the reduction shown by meander wavelengths.

Dury began research involving subsurface exploration of the bottoms of valleys occupied by former manifest underfits. The first stream selected was the Warwickshire Itchen, which is very suitable, being on impermeable alluvium and impermeable solid rock, tectonically stable and not extended or reduced by divide migration. Discharge of meltwater cannot adequately explain the valley meanders of the Itchen. Not only does the stream flow towards the direction of the local ice front, but the valley meanders were still being shaped after the last local ice had disappeared, and after a whole interglacial and succeeding glacial had intervened. These conditions make the Itchen a classic representative of underfitness.

Conclusion

Many examples exist in southern England where the present stream and its meanders are disproportionately small compared with the size of the valley. The valleys themselves exhibit meandering. The Itchen is a classic example of this underfitness and was the first to be investigated in the 1950s.

RIVER CHERWELL AT TRAFFORD HOUSE, NORTHAMPTONSHIRE (SP 528487)

Highlights

The River Cherwell and the Eydon Brook (Figure 6.24) are both examples of underfit streams where morphological and sedimentary evidence suggests that the streams conveyed greater discharges in relation to palaeohydrological conditions during glacial periods.

Introduction

The Upper Cherwell and the Eydon Brook are excellent examples of underfit streams, their channel dimensions being much reduced when compared with those described by the subsurface contours. At Trafford House (Figure 6.25), the deposits occupy an old channel that is still visible on the surface and is abandoned by undercutting. They include alluvial material, shell debris and tufa from an adjacent limestone face, confirming the condition of the streams as underfits.

Description

The source of the Cherwell lies in a belt of dissected country where hills of Upper Lias, capped by Northampton Sand, overtop the broad structural bench of Middle Lias Marlstone. Drift deposits of importance in the present context are found in the

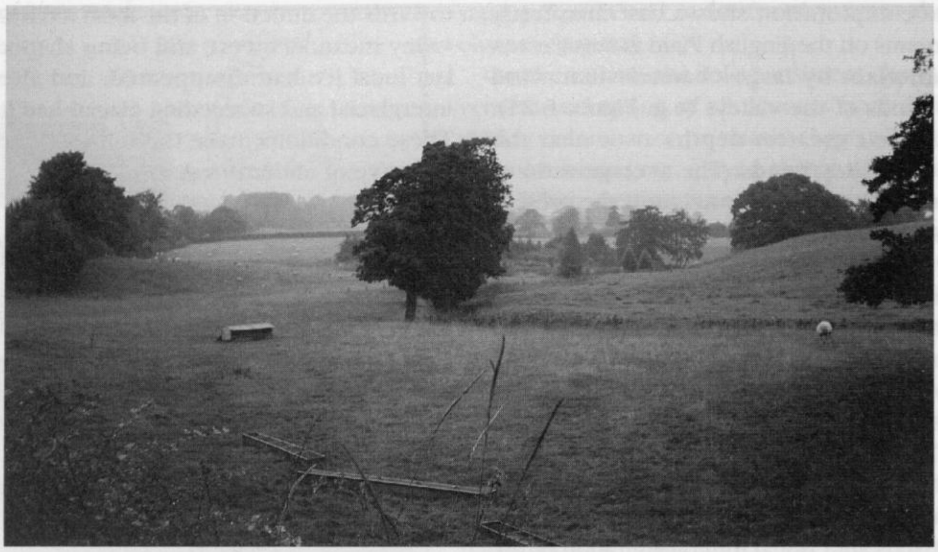

Figure 6.24 The palaeochannel of the Eydon Brook, looking upstream to section C (of Figure 6.25). (Photo: R.J. Davis.)

two left-hand tributaries; one being Eydon Brook at Trafford House (SJ 527486).

The two streams are excellent examples of misfit, or more properly underfit, streams. The underfit condition is recognized by the fact that the meanders of the stream are described within the limits of the much more ample meanders of the valley.

Borehole investigations prove the presence beneath each floodplain of a channel much wider and deeper than the existing stream channel. The upper deposits consist of cloddy or moist and tenacious clay, while the lower deposits, confined largely to the deeper portions of the filled channels, are made up of black or dark grey silt, moist, very soft, and containing plant debris. The upper clay is interpreted as material which settled out of floodwater, while the lower silt may represent the former bottom load in the large channels, together with the remains of plants washed or fallen in.

In general, the greatest depths are recorded at the outside of valley bends. The width varies somewhat, tending to be greater at bends than in the intermediate reaches, but is roughly ten times that of the present stream channel. It seems beyond doubt that the filled channels were formerly the beds of large streams, much more powerful than the Cherwell and Eydon Brook of today. In Figure 6.25 is illustrated the present-day Eydon Brook flowing between the banks of the former larger capacity channel.

The ratio between valley meander wavelength is approximately ten times as great as the interval between the meanders of the present stream. The meandering valley is in part cut into, and through, the deposits of Chalky Boulder Clay, so it follows that the loss of volume must post-date the glaciation.

The Trafford House site is particularly interesting because it contains certain deposits unlike anything so far discovered elsewhere — tufa, shell marl, peat and peaty clay occur here. These deposits occupy an abandoned channel that curves gently round a valley bend. The hollow formed by the enclosed ends of the abandoned channel received a variety of materials, but is not yet entirely filled.

Interpretation

A close relationship is postulated by Dury (1953) between an ice-front and the Cherwell valley, and the eastern divide of the Cherwell drainage has been termed morainic. However, the valley had been fashioned before the onset of the Chalky Boulder Clay ice.

The line of the Eydon Brook appears to have been determined during or before glacial times. Remnants of the sheet of Chalky Boulder Clay with associated underlying gravels occur on the high ground to the north and south of the valley of the Eydon Brook, whereas within it small patches of gravel alone remain. It seems probable that this

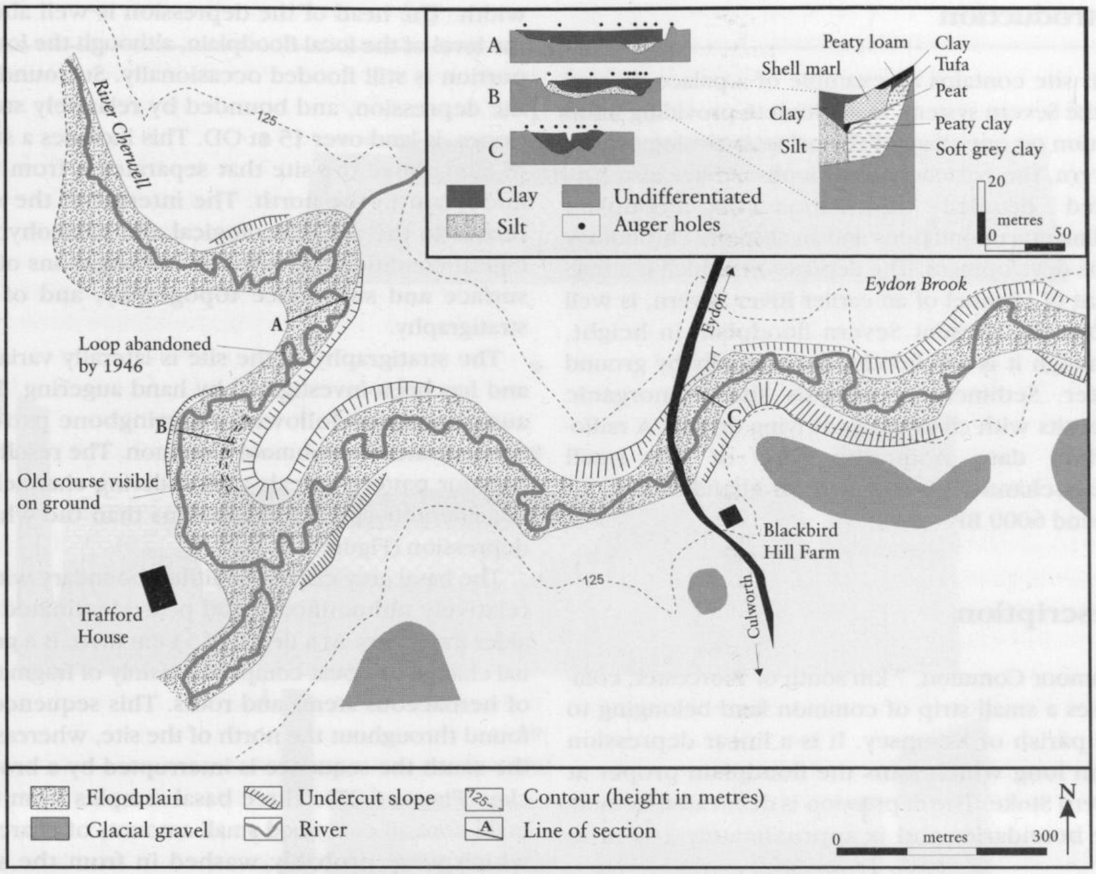

Figure 6.25 Palaeohydrological features of the River Cherwell and Eydon Brook. (After Dury, 1953.)

valley received quantities of glacial gravel, most of which has now been cleared, and that the valley floor has been cut below the level reached when the gravel was deposited.

In certain localities of the upper Cherwell valley there is abundant scope for study of drifts, but it now seems well established that the valley system existed much as it is now before the arrival of the Chalky Boulder Clay ice. The conclusion that the size of the headwater catchment has changed very little since deglaciation, whether by capture or by divide migration, has a direct bearing on the problem of formation of valley meanders and cause of the present underfitness.

Identification of underfit streams and inherited valley meanders was an important development in fluvial geomorphology. These features remain valuable for determination of Quaternary chronologies and for palaeohydrological investigations. This site is particularly significant because of the unusual deposits in the abandoned valley meanders as well as its classic features of underfitness.

Conclusion

These streams are classic underfits. Their present stream meanders are well-developed within the valley meanders cut into the Chalky Boulder Clay. Two features of particular interest are the abandoned channel visible on the surface, and the deposits which have been deposited in it. The site has both intrinsic and potential value.

ASHMOOR COMMON, HEREFORD AND WORCESTER (SO 855465)

Highlights

The Ashmoor Common site provides an excellent example of a palaeochannel of the River Severn. The site, abandoned around 6000 BP, contains evidence of former sedimentary evolution of the floodplain.

Introduction

This site contains an example of a palaeochannel of the Severn system. In addition to providing information on one stage of the palaeohydrology of the Severn, the palaeochannel deposits have also provided detailed information on floodplain sedimentary conditions and an absolute chronology of its development. The depression, which is a fragment of channel of an earlier River Severn, is well above the present Severn floodplain in height, although it is still flooded occasionally by ground water. Sedimentary analysis shows inorganic deposits with channel sands lying on top. A radiocarbon date from the base of the small palaeochannel gives a date of abandonment of around 6000 BP.

Description

Ashmoor Common, 7 km south of Worcester, comprises a small strip of common land belonging to the parish of Kempsey. It is a linear depression 4 km long which joins the floodplain proper at Severn Stoke. The depression is delineated by land-use boundaries and is approximately 160 m in width. The head of the depression is well above the level of the local floodplain, although the lower portion is still flooded occasionally. Surrounding the depression, and bounded by relatively steep slopes, is land over 15 m OD. This includes a strip of land above the site that separates it from the floodplain to the north. The interest of the site relates to the palaeoecological and palaeohydrological conditions revealed by investigations of its surface and subsurface topography, and of its stratigraphy.

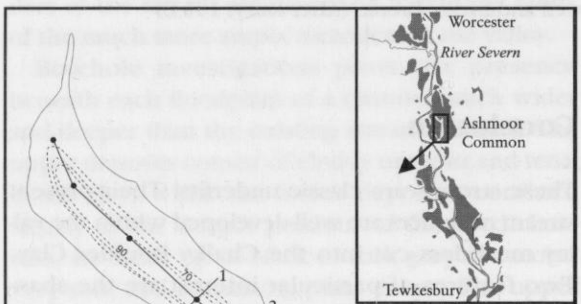

Figure 6.26 Ashmoor Common: location and sub-peat contours. (After Brown, 1982.)

The stratigraphy of the site is laterally variable and has been investigated by hand augering. The auger locations followed a herringbone pattern concentrated on Ashmoor Common. The resulting contour pattern reveals a meandering channel of considerably smaller dimensions than the whole depression (Figure 6.26).

The basal grey clay has a diffuse boundary with a relatively unhumified wood peat, dominated by alder fragments. At a depth of 54 cm there is a gradual change to a peat composed mainly of fragments of herbaceous stems and roots. This sequence is found throughout the north of the site, whereas to the south the sequence is interrupted by a brown clay (Figure 6.27). Three basal samples from the main core all contained small amounts of charcoal which were probably washed in from the surrounding slopes along with the clay.

Interpretation

Five samples were sieved from the inorganic deposits underlying the peat, which varied from sand and gravel to silty sands. These can be compared to a sample from the main terrace and the top of the Worcester terrace, suggesting that at least part could be terrace derived. The sand and gravels are interpreted as being channel and/or locally inwashed sediments, while the clay is typical backswamp sediment. From the grain size analysis alone, it is difficult to separate the underlying sand and gravels from terrace and modified terrace deposits. However, the depth of occurrence, relative lack of thickness, lack of cohesion and relatively greater proportion of fines suggest they are reworked terrace channel deposits, with channel sands, also probably locally derived, lying on top. The large Ashmoor channel post-dates the Worcester terrace which it dissects, and it is probably of Late Glacial age.

The site is a relict tract of floodplain which has filled with autochthonous peat over the past 6000

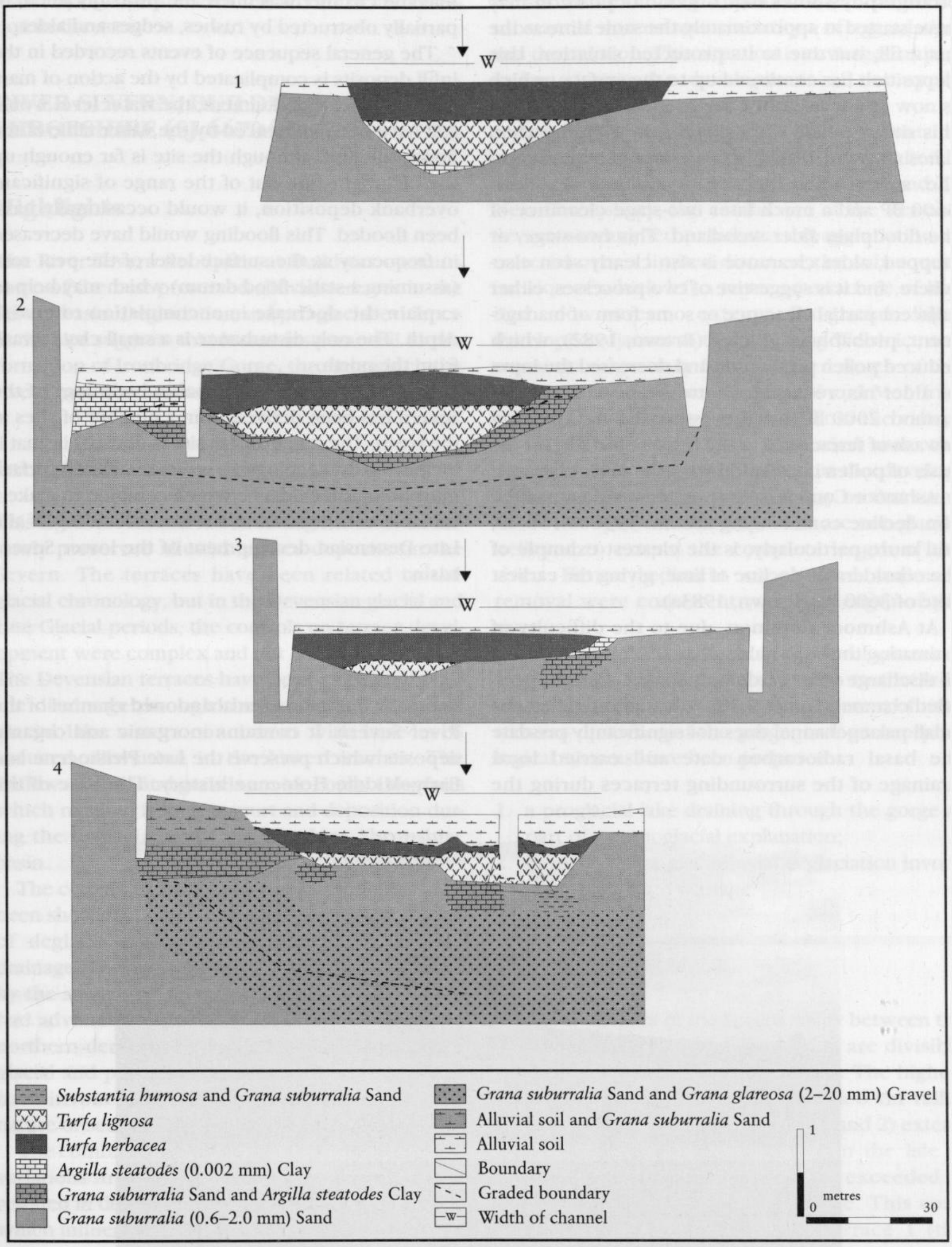

Figure 6.27 Ashmoor Common: cross-sections. (After Brown, 1982.)

years (basal dated 5930 ± 70 BP, HAR4350). The stratigraphy shows that organic deposition may have started at approximately the same time as the main fill, but due to its protected situation, this deposition has continued up to the surface, which is now at a level above the main floodplain. From this site, a pollen diagram with four radiocarbon dates (Brown, 1983a) shows the deforestation of the surrounding terrace woodland at about 3600 BP and a much later two-stage clearance of the floodplain alder woodland. This two-stage, or stepped, alder clearance is also clearly seen elsewhere, and it is suggestive of two processes, either adjacent partial clearance or some form of management, probably coppicing (Brown, 1982), which reduced pollen production and decreased the input of alder macro-remains into the peat. A date of around 2000 BP can be estimated for the clay inwash of terrace soil at the head of the bog on the basis of pollen correlation with the main core.

Ashmoor Common also provides evidence of the elm decline commencing around 5930 ± 100 BP, and more particularly, is the clearest example of the considerable decline of lime, giving the earliest date of 3600 BP (Brown, 1983a).

At Ashmoor Common, due to the difficulty of estimating the large palaeochannel depth, estimates of discharge were made only for the smaller, peat-filled channel (Figure 6.28). It is suggested that the small palaeochannel does not significantly pre-date the basal radiocarbon date and carried local drainage of the surrounding terraces during the pre-Boreal and Boreal. Pollen evidence suggests that the channel was lined and probably filled, or partially obstructed by rushes, sedges and alder.

The general sequence of events recorded in the infill deposits is complicated by the action of man. After channel abandonment, the water level would still have been influenced by the water table of the floodplain and, although the site is far enough up the channel to be out of the range of significant overbank deposition, it would occasionally have been flooded. This flooding would have decreased in frequency as the surface level of the peat rose (assuming a static flood datum) which may help to explain the decrease in accumulation rate with depth. The only disturbance is a small clay inwash from the north.

Evidence relating to the palaeohydrology of the Severn has been derived from a number of sites in the basin. The range at this site is unusual in that it includes palaeoecological, sedimentary and surface morphological evidence which combine to make a valuable contribution to the understanding of the Late Devensian development of the lower Severn basin.

Conclusion

Ashmoor Common is an abandoned channel of the River Severn. It contains inorganic and organic deposits which preserve the Late Pleistocene and Early–Middle Holocene history. The base of the

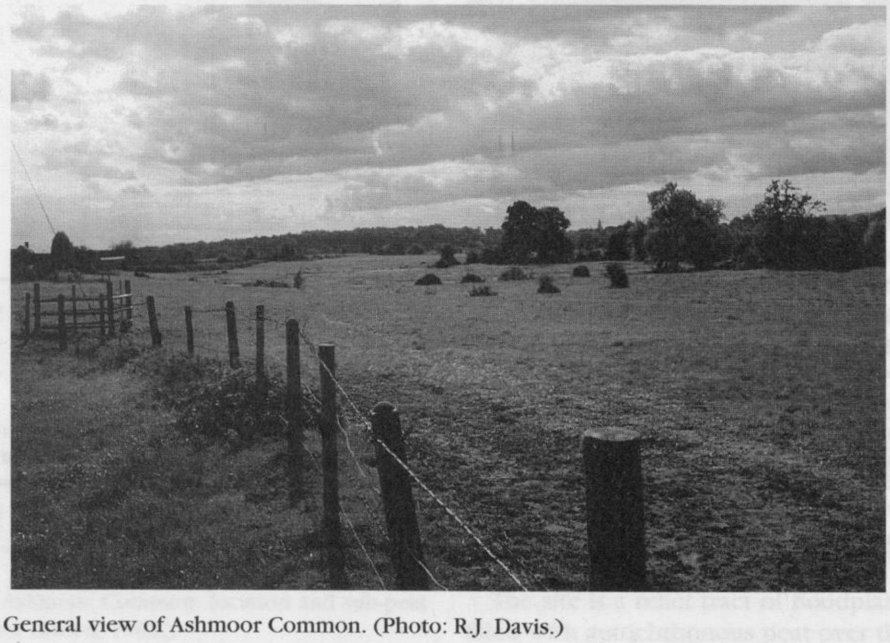

Figure 6.28 General view of Ashmoor Common. (Photo: R.J. Davis.)

peat has been dated as *c.* 6000 BP. Pollen analysis of overlying layers reveals the record of deforestation in the area.

RIVER SEVERN, BUILDWAS, SHROPSHIRE (SJ 647042)

Highlights

The River Severn floodplain at Buildwas contains some of the best preserved alluvial terraces in the Severn basin. Deglaciation and pro-glacial evidence at the site suggests that it was instrumental in the formation of Ironbridge Gorge, through which the river flows downstream of the site.

Introduction

This site is set in an area containing the most extensively preserved alluvial terrace deposits on the Severn. The terraces have been related to the glacial chronology, but in the Devensian glacial and Late Glacial periods, the controls on terrace development were complex and not easily decyphered. The Devensian terraces have been associated with the discharge through the Ironbridge Gorge of glacial lake waters concurrent with base-level-induced alluviation in the lower reaches. The Buildwas sands and gravels are ice-contact deposits which resulted from transport and deposition during the deglaciation of the Cheshire–Shropshire basin.

The courses of the Severn and its tributaries have been shown to have been determined by the form of deglaciation and the course of meltwater drainage. The area has been profoundly influenced by the stagnation of an Irish Sea ice sheet, which had advanced over the stagnating blocks of the northern derived ice mass. The nature of the proglacial and paraglacial environment is believed to have invoked considerable local control on the nature of sedimentation and erosion.

The combination of the deglacial history and the variations in stream dynamics is believed to have resulted in downstream variations in stream activity which influenced both the degree of floodplain formation and the extent to which palaeofloodplains can now be recognized as terrace remnants. Thus, with the Welsh ice near its Shrewsbury limit, a sandur developed initially over an ice-free surface. It is proposed that it then entered the stagnating Irish Sea ice mass in the proximity of a buried valley, within which the local controls above introduced substantial downstream variations of aggradation and degradation. This situation is envisaged as prevailing during the development of the upper terrace suite (terraces 6, 5 and 4).

As ice retreated towards the Welsh mountains and the remaining blocks of Irish Sea ice melted, a middle terrace suite developed (terraces 3 and 2). Near Shrewsbury, the confluence of the Rea and Severn sandurs led to substantial aggradation of material over a limited distance, the subsequent incision of which left a terrace (terrace 3) of limited downstream extent. Pro-glacial lakes developed adjacent to the retreating Welsh ice and morainic ridges may have covered much of the Severn valley between Welshpool and Montford Bridge. Outwash fans emanating from the valleys of the Welsh Mountains prograded over the drained lake basins and resulted in the development of the low terrace suite. It may be that some of the floodplain terrace fragments are non-glacial, believed to extend the complete length of the river. Hence sediment supply and sediment removal were consistent with the length of the river.

The most critical problem in the development of Ironbridge Gorge is its relationship to the sediments upstream. Prior to the development of river terraces upstream of the gorge, two geomorphological systems have been proposed:

1. a proglacial lake draining through the gorge as part of a monoglacial explanation;
2. more complex glaciofluvial deglaciation involving stagnant ice wastage.

Description

The river terraces of the Severn valley between the Vyrnwy confluence and Ironbridge are divisible into three suites (M.D. Jones, 1982). The highest terraces (6, 5 and 4) are limited to a 6 km valley section. The middle suite terraces (3 and 2) extend along the length of the valley from the Isle at Shrewsbury to Ironbridge, and are exceeded in length only by the low terrace suite. This comprises the terrace fragments of terrace 1 (the Cressage terrace), at 2–4 m above the floodplain, and the floodplain terrace fragments. The latter are gravel outcrops within the modern floodplain, varying from 1 m above the present silty alluvium surface to 1 m below. The scattered nature of these terrace fragments may reflect more the adjustments of the river to its own aggradation than to a meandering river system.

A kettle hole extends beneath the projected terrace 2 level west of Buildwas Abbey, and suggests the persistence of isolated blocks of Irish Sea ice during the formation of this terrace. The surface that surrounds this hole is 2–3 m above the expected level of terrace 2, and is not thought to be part of it. Rather, it is believed to represent a tributary fan emanating from Farley Brook into the main Severn Valley very similar to that at the top of terrace 2, evidence for which is derived from Buildwas Quarry.

Buildwas Quarry deposits: description and interpretation

The deposits at Buildwas are in a critical position, since they are just upstream of the entrance to Ironbridge Gorge and downstream of the terrace sequence of the Shropshire Plain. The deposits comprise massive sands and gravels at the base, overlain by a fluvial gravel unit (Figure 6.29).

The deposits at Buildwas illustrate the facies changes which are typical of the zone between the gravel-dominated and sand-dominated units. Shaw (1972) has argued that these sediments represent the ice-walled fluvial sedimentation of meltwater draining to the south-east, as indicated by palaeocurrent measurements. The indicator is that ice-walled sedimentation was characterized by a zone of gravel bars, inferred from horizontal imbricate and foreset gravel units. Flanking this facies assemblage are finer sediments deposited as low flow regime horizontal stratification.

At the top of the quarry, reaching an altitude of 65 m OD, is a gravel unit (Figure 6.29) the provenance of which indicates that 90% of the clasts originate from within the present catchment of the tributary Farley Brook. The gravel unit is up to 2 m thick and consists of a lower well-stratified gravel–sand sequence and an upper poorly sorted, often massive gravel unit. Sand lenses often define gravel sets of 30–50 cm thickness, which are poorly imbricated. The altitude of the gravel indicates that fluviatile drainage was taking place towards Ironbridge Gorge after the cessation of ice-contact sedimentation. Also derived from the quarry is evidence of the avalanche front of a gravel bar. The proximal foresets are of medium to fine gravel with openwork-matrix filled co-sets defining the cross-stratification. Downvalley, the sediment contains more coarse–medium sand. The high percentages of medium sand found in the gravel derived from the Farley Brook are probably reworked from the underlying sands and gravels.

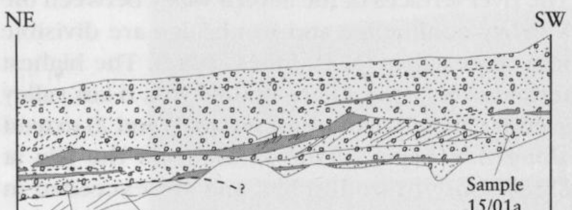

Figure 6.29 Fan gravels at the top of the Buildwas quarry section.

Interpretation

There is no evidence that Ironbridge Gorge was initiated before the Middle Devensian. Two major theories have been proposed.

The pro-glacial lake

Following Howson's (1898) realization of the work of land ice in the creation of the drift deposits of the Midlands, Lapworth (1898) — and more fully Harmer (1907) — proposed that diversion of the Severn took place after the damming of a lake formed by advancing ice. This proposal was recognized by Wills (1924), who mapped deposits and overflow channels. Wills proposed two lakes, glacial lake Newport (Dixon, 1921) and glacial lake Buildwas, which later coalesced to form glacial lake Lapworth; this was then suggested to have drained through a col at what is now Ironbridge Gorge.

This monoglacial explanation was followed by Pocock *et al.* (1938), who stated that although a reddish, loamy, sandy clay, and bedded sands and gravels were distinguished, it was often not possible to prove their chronological relationship. These deposits were thought to have been laid down by the contemporaneous Irish Sea and Welsh ice sheets. As the ice sheets waned, meltwater deposited sand and gravel as kame deltas and fans, and was eventually ponded up by the ice margins and the rim of the Shropshire–Cheshire basin, to form the 300 ft (91 m) stage of Lake Lapworth with its overflow channel at Ironbridge gorge. It was suggested that, at about this time, the Welsh ice-sheet readvanced into the western part of the Shrewsbury district and deposited boulderclay on the outwash sand and gravels of the Irish Sea ice.

Worsley's (1985) reappraisal concluded that 'it is inadvisable to reject the concept of a proglacial lake on the basis of limited sedimentological evidence, especially when this is amenable to interpretation in various ways, none of which necessitates the elimination of glacial Lake Lapworth'.

Stagnant ice wastage

M.D. Jones (1982) summarized the three main lines of evidence for stagnant ice wastage as the explanation of the deposits and features. The gorge is envisaged as a natural continuation of a subglacial drainage system and indicates the presence of ice during the formation of the gorge. Sediment palaeocurrent directions point to the presence of ice at the northern end of the gorge, while water entered at or near its base. Faulting of sediments, incorporated till bodies and their esker-like or kamiform morphology all point to an ice-contact environment for the sediments that flow into the gorge.

A more likely explanation is that the retreating ice margin ceased and melted ice *in situ*, possibly forming a small pro-glacial Lake Buildwas, and overflowed in the gorge (Hamblin, 1986). This, in combination with the cessation of sub-glacial flow through the Lightmoor channel (to the north-east of Ironbridge Gorge) represents the shift of drainage from the Stour valley to the Severn (Hamblin, 1986). The lack of lacustrine sediments between those around Lilleshall and those possibly at Buildwas suggests that Lake Newport and Lake Buildwas formed adjacent to their respective overflow channels and may not have coalesced, with Lake Buildwas already having drained through Ironbridge Gorge before Lake Newport drained into the River Tern (Hamblin, 1986).

The most significant development of the River Severn, associated with a decrease in slope, is from an unstable multi-thread proximal pro-glacial river to a stable low-sinuosity single-thread river. This increased stability is reflected in a change from sediments representing migrating mid-channel bars to diagonal bar–riffle development. This trend has been attributed to a change in sediment load of the river from a gravel-bedload river to a silt-suspended-load river.

The present river has adopted 'meander' wavelengths that correspond with those of the terrace 1 meanders. Since these channel meanders are not consistent with the modern fluvial regime, the Severn is a classic Osage type underfit. Many of the anomalous channel reaches, such as the straight reach between Shrawardine and Montford, appear likely to have been inherited from the deglacial evolution of the area. The river is now very stable for much of its length, and very little migration has taken place since Roman times.

In character and location, the deposits are important in relation to the interpretation of the late-Devensian history of environmental change in the Severn Basin, in which the Ironbridge Gorge plays a fundamental role. The quarry sediments are essential to the understanding of the origin and development of Ironbridge Gorge, which are subject to continuing debate in the literature. The importance of local controls on sedimentation and erosion in the pro-glacial and para-glacial environments of the Late Devensian give these deposits a unique value in this context.

Conclusions

Competing theories of explanation of the changes in the drainage of the Severn and the formation of the Ironbridge Gorge in the Late Glacial are still debated. Sand and gravel deposits and a river terrace at Buildwas stand at a crucial position in relation to the gorge and the upstream terraces. A complex stratigraphy and key sedimentary features are exposed in the quarry at Buildwas and provide evidence of the sequence of development.

ALPORT VALLEY, DERBYSHIRE (SK 118938)

Highlights

The Alport valley, draining the eastern slopes of Bleaklow, illustrates the morphology formed in upland channels where, with limited sediment supply, the river cuts into the underlying bedrock forming step–pool features, rapids and waterfalls.

Introduction

The origin of meandering bedrock valleys has intrigued and frustrated geomorphologists. A major difficulty is the inability to observe and measure the processes and rates of erosion in bedrock channels, or to monitor long-term changes in their position in response to the slow rate of operation of the relevant processes.

Description

Bedrock channels occur wherever potential rates of removal of material exceed sediment supply: in upland areas with steep slopes, in glaciated hard-rock regions, and in areas undergoing active tectonic uplift. Alport Valley (Figure 6.30) is cut in sandstone with very steep slopes which have contributed to the spectacular landslips of Alport Castle downvalley from the site. Parts of the stream channel are cut in bedrock, while elsewhere discrete concentrations of coarse material form the bed. The valley has waterfalls and rapids usually associated with bedrock channels, which are more likely to be caused by resistant strata on the stream course than by rejuvenation (Petts and Foster, 1985).

Parts of the stream channel also possess the step and pool form characteristic of mountain and upland rivers. The step–pool structure gives rise to a distinctive tumbling flow pattern. Bedload transport processes in such streams are strongly linked to the bed form. Steps are formed of large bed elements, often of the order of the depth of flow, or even the width of the channel. The coarse sediment is transported from discrete sites and stored in pools.

Alport Valley also has the typical narrow, discontinuous floodplain of steep channels. An analogy with the headwater tributaries of lowland streams highlights several important differences in behaviour between upland and lowland rivers (Lisle, 1987). Direct inputs of sediment from hillslopes to upland streams cause sediment transport to be episodic. Resistant, non-alluvial boundaries inhibit lateral migration and can control the position of large-scale bedforms. Where the stream channel is unconfined, it forms a freely meandering planform in alluvial material.

Interpretation

Dominant channel-forming discharges tend to be less frequent in upland channels than in lowland ones. High-magnitude, low-frequency events can be more effective in shaping upland channels. The steep valley slopes, in parts of the site forming vertical bare rock cliffs, and the locally constrained valley bottom, can generate extreme unit stream power. Bed or bank materials are generally too

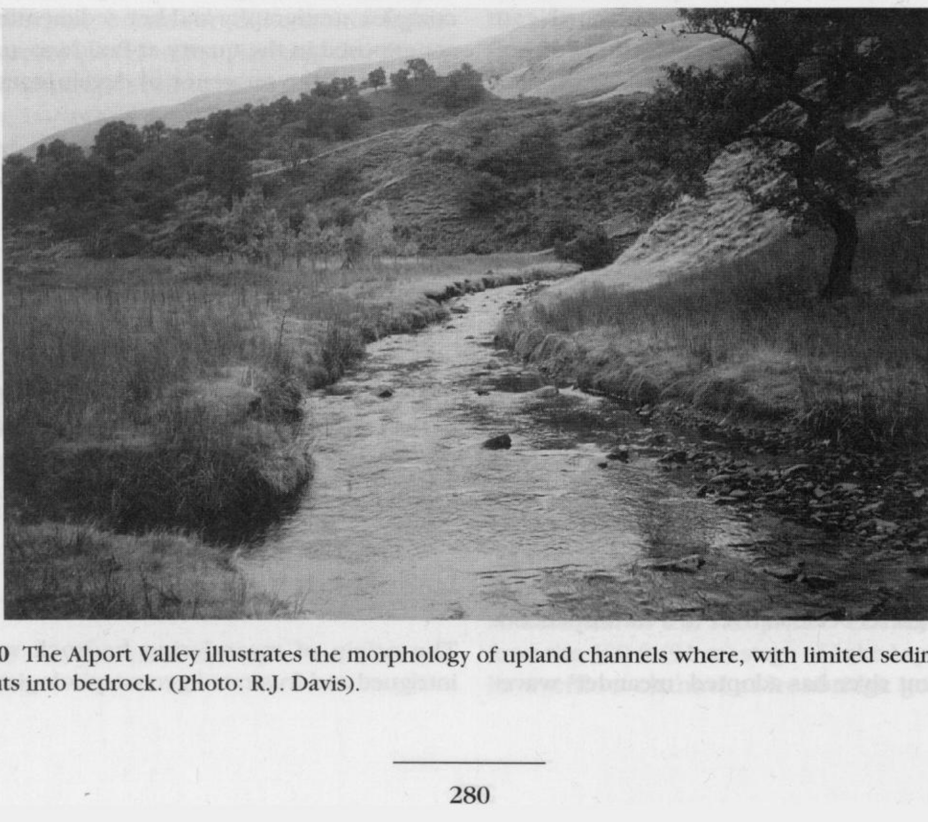

Figure 6.30 The Alport Valley illustrates the morphology of upland channels where, with limited sediment supply, the river cuts into bedrock. (Photo: R.J. Davis).

large or too resistant to deform in all but the most extreme events, and post-flood flows may be unable to modify the resulting forms of large floods.

Alport Valley is important for its assemblage of fluvial landforms typical of upland rock-bed and coarse bedload channels, in addition to possessing meandering reaches in unconfined situations. The concentration of channel landforms such as caprock waterfalls, rapids, steps and pools, and bedload features, together with valley landforms including hanging tributaries, discontinuous floodplains and terrace surfaces, make it a classic representative of this type of stream. Although little specific research has been undertaken in the valley, it has the potential to help understand the formation of bedrock meanders and bedrock channel features.

Conclusion

The Alport Valley contains excellent examples of the features typical of an upland channel cut in bedrock. These include waterfalls, steps and pools, a narrow floodplain, steep valley sides, coarse bedload and direct sediment input from the slopes.

BLEAKLOW, DERBYSHIRE (SJ 183865)

Highlights

The high moorland areas of the southern Pennines have some of the highest rates of peat erosion in the UK. Erosional activity, which dates from the formation of the peat blanket, has produced a number of different morphological responses. While climatic adjustments may have been responsible for historical degradation, human impacts may have increased recent activity.

Introduction

Many of the peat deposits of the southern Pennines are undergoing recent erosion, creating a number of morphological features (Figure 6.31). A number of factors has been suggested as causing the erosion, including: the natural instability of rapidly growing peat under deteriorating climatic conditions, (Conway, 1954); climatic variability (Tallis, 1985); the influence of biotic factors (Bower, 1962; Radley, 1962); the effects of slope (Radley, 1962); inheritance of stream characteristics developed

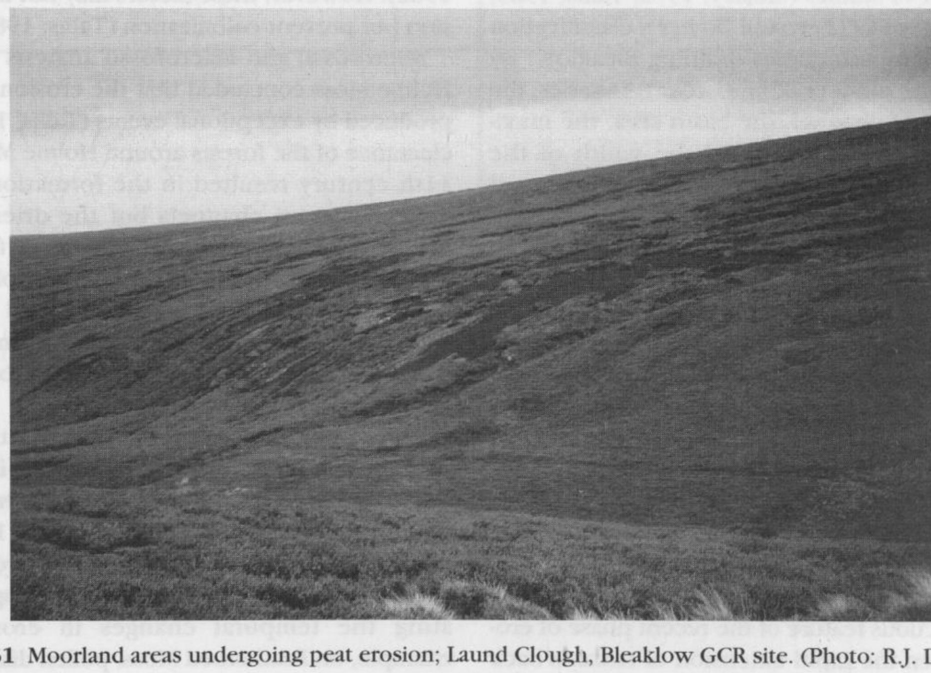

Figure 6.31 Moorland areas undergoing peat erosion; Laund Clough, Bleaklow GCR site. (Photo: R.J. Davis.)

under different climatic and ground surface conditions (Johnson, 1957); air pollution (Tallis, 1965); and moorland fires (Tallis, 1987).

Description

A number of morphological classifications of peat erosion have been put forward. Bower (1960, 1961) suggested two types of gully erosion on the peat, Type I dissection by gully and sheet erosion of flat areas in a close network, and Type II being less intensive with gullies occurring on sloping ground, forming sub-parallel trenches. Radley (1962) proposed an alternative classification of grains and groughs, peripheral erosion, sub-peat erosion and summit erosion. Grains and groughs are tributary channels and gullies that have cut between peat mounds isolated by erosion and causing stream head retreat. Peripheral erosion at the edge of the bog is caused by sheetwash, although Conway (1954) and Bower (1960) have attributed this form to bog-burst and to slump respectively. Sub-peat erosion occupies tunnels in the peat and open joints in the gritstone and may be downstream of minor grains.

Finally, summit erosion is characteristic of deeply dissected plateau areas. Tallis (1965) combined these typologies into stream erosion (which included U- and V-shaped cross-sectional gullies), summit erosion and marginal recession.

Bower's classification has been debated and used in a number of studies (Mosley, 1972; Tallis, 1985, 1987). Mosley (1972) tested Bower's classification on the streams and gullies draining Bleaklow. By measuring the number of first order tributaries, the total length of channel, the basin area, the maximum straight-line length and the width of the basin, Mosley (1972) concluded that while good examples of each type can be found, these tend to be at opposite ends of the spectrum. Difference in patterns can be attributed to natural variance, as only when there are gross differences in slope can the flow patterns be superficially distinctive. However, random simulations of straight, concave and convex slopes produce patterns similar to the actual stream system on Bleaklow (Figure 6.32). Tallis (1985, 1987) identified Type I and II gullies on Featherbed Moss (at the headwaters of the River Ashop) and Holme Moss. However, these different gully systems may have developed at different times.

A conspicuous feature of the recent phase of erosion has been the rapid extension of streams back into the peat blanket (Radley, 1962). Moss (1913) calculated a general 3/4 mile extension between the Ordnance Survey editions of 1830 and 1870, and a further 1/4 mile by the 1912 re-survey. There are few areas where the opposing headwaters have not reached the summits and dissected the interfluve. Howden Moor on Bleaklow is typical of this general condition and, in such an active environment, headwater captures have been identified (Pugh, undated); for example, the headwaters of the Derwent have been captured from those of the River Little Don.

Interpretation

The peat erosion on the high peaks has been attributed to a number of factors, including natural processes, human activity, biotic factors, air pollution and climatic change (Moss, 1913; Conway, 1954). The natural instability of the peat has been considered to be instrumental in causing erosion, as in some areas erosion may be an intrinsic property of the peat (Bower, 1961). Once exposed, lateral erosion by wind action may occur on exposed ridges.

Biotic factors have also been proposed for initiating peat erosion where destruction of plant cover exposes the peat blanket to the elements (Radley, 1962). Overgrazing by sheep, development of pack horse trails, and World War One manoeuvres also have been blamed for exposing the peat (Radley, 1962). However, these factors may not initiate erosion but prevent colonization (Tallis, 1985).

Microfossil and macrofossil analysis at nearby Holme Moss concluded that the erosion had been produced by exceptional events (Tallis, 1987). The clearance of the forests around Holme Moss in the 11th century resulted in the formation of well-defined stream channels but the drier climatic conditions early in the Middle Ages led to the subsequent lowering of the water table followed by climatic deterioration in the 18th and 19th centuries. Moorland fires and a cloudburst in July 1777 may both have accentuated existing erosional features (Tallis, 1987).

Atmospheric pollution may have contributed to recent erosion (Tallis, 1964). However, in the case of Holme Moss it was already established by the time of the industrial revolution (Tallis, 1987).

Climatic conditions represent a control on rates of erosion and can be used as a backdrop to evaluating the temporal changes in erosion. For example, on Featherbed Moss, pollen diagrams and

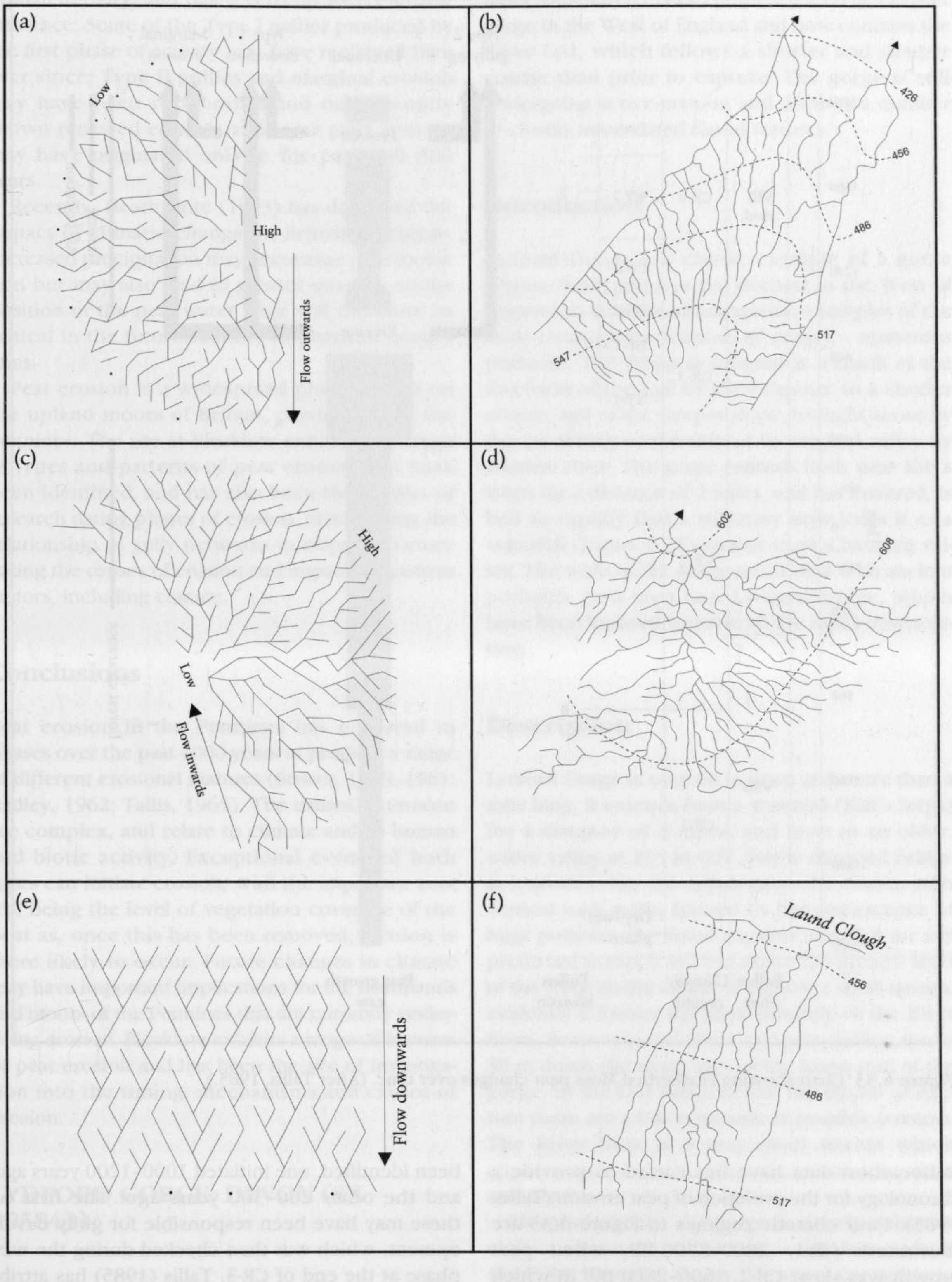

Figure 6.32 Simulated and actual streams on Bleaklow: (a,b) convex slopes; (c,d) concave slopes; (e,f) straight slopes. (After Mosley, 1972.)

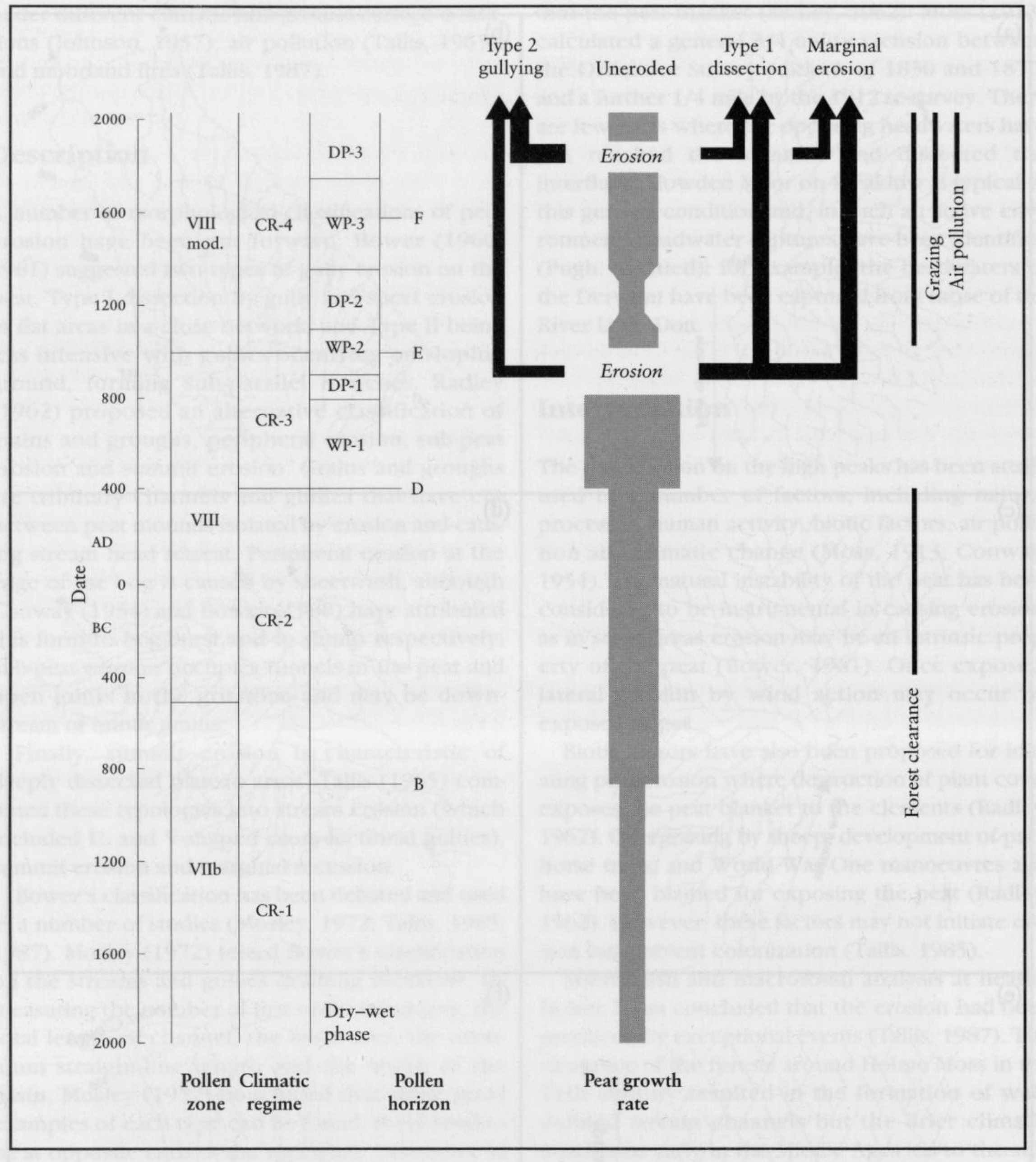

Figure 6.33 Chart showing Featherbed Moss peat changes over time. (After Tallis, 1985.)

radiocarbon data have been used to provide a chronology for the evolution of peat erosion (Tallis, 1985). Four climatic regimes in Figure 6.33 are illustrated: CR-1, 2800–5700 BP, when peat growth was slow; CR-2, 1600–2800 BP, in which peat accumulation increased in the wetter climate; CR-3, 400–1000 AD, associated with rapid peat growth; and CR-4 initiated since 1000 AD, a slower growth phase. Two periods of peat erosion have been identified, one initiated 1000–1200 years ago and the other 200–300 years ago. The first of these may have been responsible for gully development, which was then checked during the wet phase at the end of CR-3. Tallis (1985) has attributed this activity to naturally induced mass movement, with marginal areas of the peat blanket drying out. The second phase of erosion is more recent and currently active through biotic and human activity, and has less to do with climatic influence. Some of the Type I gullies produced by the first phase of activity may have remained bare ever since; Type II gullies and marginal erosion may have been recolonized and only recently shown renewed erosion; and some gully systems may have originated only in the past 200–300 years.

Recently, Heathwaite (1993) has discussed the impact of climatic change on British peatlands. Increased precipitation may encourage peat formation but may also lead to greater erosion, so the position of the peat water table will therefore be critical in the future stability of peatland ecosystems.

Peat erosion is a widespread phenomenon on the upland moors of Britain, particularly in the Pennines. The site at Bleaklow exhibits the range of types and patterns of peat erosion that have been identified, and has also been the subject of research dating phases of erosion, investigating the relationship of gully networks to slope, and elucidating the causes of erosion and impacts of various factors, including climate.

Conclusions

Peat erosion in the Pennines has occurred in phases over the past 4000 years to produce a range of different erosional features (Bower, 1960, 1961; Radley, 1962; Tallis, 1965). The causes of erosion are complex, and relate to climate and to human and biotic activity. Exceptional events of both types can initiate erosion, with the important control being the level of vegetation coverage of the peat as, once this has been removed, erosion is more likely to occur. Future changes in climate may have important implications for the heathlands and moors of the Pennines that are currently undergoing erosion. Bleaklow exhibits a range of features of peat erosion and has been the site of investigation into the timing, mechanisms and causes of erosion.

LYDFORD GORGE, DEVON (SX 505843)

Highlights

Lydford Gorge is a classic example of the effects of river capture in creating a characteristic gorge with numerous waterfalls and potholes. It is the deepest gorge in the West of England and now contains the River Lyd, which follows a shorter and steeper course than prior to capture. The gorge is still undergoing active erosion and displays a number of closely interrelated fluvial features.

Introduction

Lydford Gorge is a classic example of a gorge (Figure 6.34) and it is the deepest in the West of England. It contains characteristic examples of the most conspicuous features of gorges – numerous potholes. The gorge developed as a result of the diversion of the Lyd by river capture to a shorter course, and of the steeper slope, brought about by the breaching of the side of its original valley by another river. The gorge extends from near Kitt's Steps for a distance of 2 miles, and has lowered its bed so rapidly that a tributary now joins it as a waterfall (Figure 6.35) issuing from a hanging valley. The walls of the gorge are riddled with ancient potholes, best seen near Lydford Bridge, which have been isolated because of the rapid downcutting.

Description

Lydford Gorge is over 60 m deep and more than a mile long. It extends from a waterfall (Kitt's Steps) for a distance of 2 miles, and rests in an older, wider valley at 215 m OD. Below the road bridge at Lydford it has the appearance of a chasm, with vertical rock walls, formed by the coalescence of huge potholes, the fluted imprints of which are still preserved as much as 15 m above the present level of the river. At the elbow of capture a small stream, evidently a former left-bank tributary of the River Burn, flows into the River Lyd after falling freely 30 m down the steep side of the lower end of the gorge. In the Lyd valley below the elbow of capture there are a few remnants of possible terraces. The River Burn is a very small stream which appears underfit in its valley of considerable width and depth.

Morphologically, the gorge is merely a narrow cleft or chasm, in places only a few feet wide in the Culm Measures Shales. The most conspicuous features of these gorges are the numerous potholes (Figure 6.36), which are instrumental in drilling out the chasms through which the river now flows.

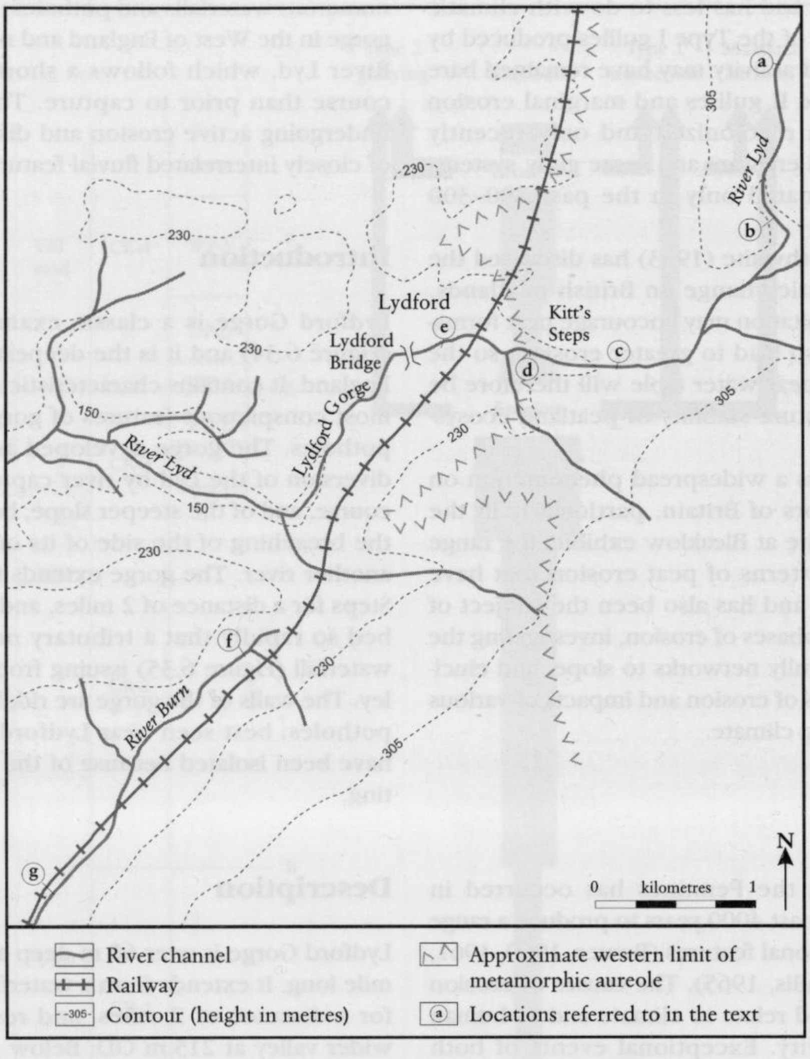

Figure 6.34 Lydford Gorge: the site of the Lyd–Burn river capture.

Interpretation

The clarity of evidence for river capture depends on the length of time that has elapsed since diversion took place. In many cases this has been sufficiently long for all evidence except the drainage pattern to have been removed. The capture of the headwaters of the River Burn by the River Lyd was first described by Dewey in 1916. The Lyd flows westwards from the western flanks of Dartmoor, across, and incised into, a series of levels usually described as high-level erosion surfaces. It falls from 500 m OD at its source to about 45 m OD where it joins the River Tamar. The river flows from its source on the granite, across the metamorphic aureole, and on to the unaltered Carboniferous slates before entering the spectacular Lydford Gorge in the section upstream from the elbow of capture. The River Burn flowed southwards, eventually joining the Tavy and the ria estuary at Plymouth.

The gorge was created by the diversion of the Lyd to a shorter course and a steeper gradient, brought about by the breaching of the side of its original valley by another river. The watershed divide between the Lyd and the Burn is near Lydford Junction station, and since its diversion the Lyd has cut a ravine exceeding 60 m in depth at this locality.

Sherlock in 1912 ascribed the gorge to the contrasting resistant powers of unaltered and metamorphosed rocks. Dewey (1916) regarded Kitt's Steps as being the head of the gorge, but whether e or d (Figure 6.34) be regarded as the

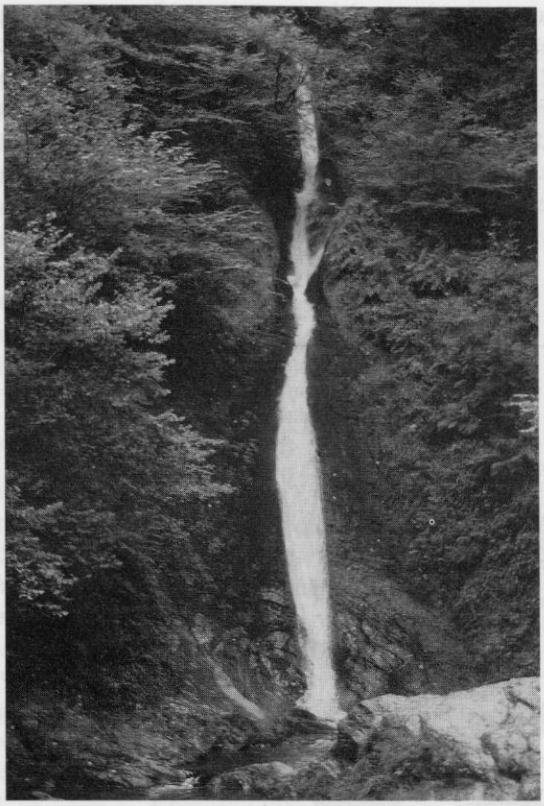

Figure 6.35 Lyd Gorge waterfall. (Photo: K.J. Gregory.)

Figure 6.36 Potholes in Lydford Gorge. (Photo: K.J. Gregory.)

true head, there is no doubt that it does lie in this zone and that it is quite distinct and removed from the knickpoint at b. The paradox then arises in which it seems possible that the gorge, apparently the most impressive evidence of capture, is in fact completely unrelated to it, being formed subsequently, since the knickpoint now at b had begun its retreat upstream from the elbow of capture. It is certain that the gorge is being actively deepened at the present time. The contours of the area lend support to the contention that the gorge rests in an older deeper valley which Dewey considered to have been eroded to a depth of about 15 m in what he described as the 750 ft (230 m) plateau. This raises the possibility that the thalweg of the River Burn before capture was represented by the line abcfg. Since c is at a higher level than f, there is no logical reason why this should not be so, in which case there is no evidence to demonstrate conclusively that the gorge is not the immediate outcome of capture.

The area is a classic example of a river gorge formed by river capture, and is of particular interest because it contains a number of closely interrelated fluvial features which can be interpreted in various ways. As evidence for the recession of the knickpoints and the stage at which capture took place, the Rivers Lyd and Burn can be studied in a wider context, in relation to the high-level erosion surfaces.

Conclusion

Lydford Gorge is the deepest gorge in the West of England and comprises spectacular examples of waterfalls, potholes and chasms. The gorge was formed by river capture, but shows evidence of the progressive development of the gorge and the relationship to the surrounding topography and drainage.

MIMMSHALL BROOK AT WATER END, HERTFORDSHIRE (TL 230043)

Highlights

The swallow holes at Water End are of national significance, being the only major sinkholes which are a permanent feature of the landscape, draining the largest enclosed karstic basin in England. Changes in water and sediment fluxes entering the river system have occurred through catchment developments, which have led to the environmental deterioration of the swallow holes, but these effects have now been ameliorated.

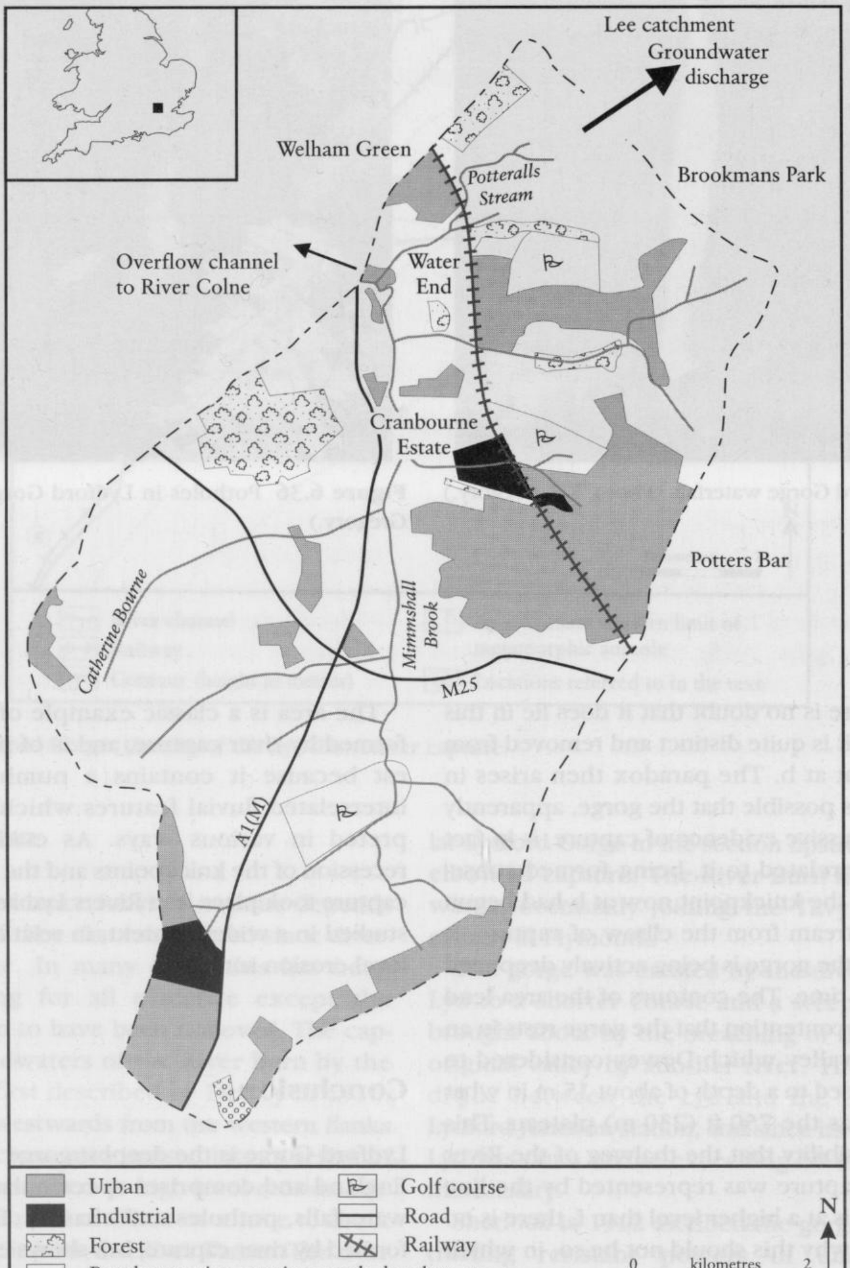

Figure 6.37 Land use in the Mimmshall Brook catchment. (After Darby and Thorne, 1992.)

Introduction

The swallow holes at Water End have fascinated researchers for over a century. Often a feature of upland limestone areas, the swallow holes at Water End are some of the most accessible in southern England. The Mimmshall Brook, flowing northward, drains a catchment of 52 km^2 which has been progressively developed (Figure 6.37). Increases in water and sediment fluxes have increased the incidence of flooding at the site, which also boasts unique flora and fauna.

Description

The swallow holes at Water End are the result of a unique combination of geological and hydrological factors (Roberts, 1989). The Mimmshall Brook catchment lies on the boundary of the Chalk in Hertfordshire and the London Clay and Reading Beds of the London Basin (Darby and Thorne, 1992). In zones of low resistivity in the Chalk, jointing and fissures caused by earth movements develop into swallow holes (Walsh and Ockenden, 1982). Solution processes and turbulent flow processes enlarge the fissures into channels, and eventually the channel may collapse, forming a swallow hole (Harold, 1937; Wooldridge and Kirkaldy, 1937).

Fluorescein tests indicate that water absorbed into the swallow holes reappears in the adjacent Lee catchment at springs on the New River north-east of Water End, flowing over a distance of 16 km in approximately three days (Harold, 1937). However, once the swallow holes reach their capacity, estimated to be 0.4 m^3s^{-1} by Roberts (1989), a lake forms over the whole area and spills through a channel to the west of Water End, directly into the surface drainage of the River Colne (Figure 6.37). In Figure 6.38 is shown an active hole in September 1994, which consumed all of the low-flow discharge at that time. The funnel-shaped depression, approximately 6 m across and 3 m deep had a fissure hole, offset from the centre, measuring approximately 1 m by 0.6 m. When the capacity of this hole had been reached, an upstream overflow channel led to a much smaller hole of diameter 0.35 m, beyond which the channel linked up to the overflow channel leading into the Colne catchment.

Figure 6.38 Water End swallow hole. (Photo: R.J. Davis.)

Interpretation

The number and location of swallow holes have been well documented and shown to fluctuate. Holes have also been observed in the channel bed of the Mimmshall Brook upstream of Water End, and in the Potteralls Stream and Catherine Brook tributaries. Historical maps indicate that between 1898 and 1919 the river channel shifted eastward away from an area which currently comprises three large 30 m diameter depressions (Roberts, 1989). Channel sinuosity between 1898 and 1973 has increased, with the development of a new swallow hole system between 1938 and 1973. These changes can be attributed to an increase in discharge. Historical evidence also suggests that the swallow holes were larger and more extensive than at present. Hopkinson (1892; cited in Whitaker, 1921) described the holes as 'large enough to carry a man', while Wooldridge and Kirkaldy (1937) reported 28 upstream of Water End and 14 at Water End (pools in the river were also counted in this figure, but even excluding these the number of holes would have been greater than today).

The observed increase in the incidence of flooding at Water End can be attributed to an increase in runoff in the catchment and an increase in sediment transport which is threatening the existence of the current hole formations. Flow into the Colne catchment has become associated with more frequent discharge events, compared with exceptional ones in the past. The increases in water and sediment have resulted from the changing nature of the catchment. From previously being a dominantly rural catchment, urban expansion, industrial development, road and motorway construction, land drainage and channelization have all contributed to increasing the sediment and water fluxes (Roberts, 1989). X-ray diffraction has attributed the fine sediments accumulating at Water End to those derived from bank erosion upstream (Darby and Thorne, 1992). Rubbish and litter are also catchment-wide concerns, as they are threatening to contribute to blockage in the swallow hole system.

Sear *et al.* (1994) have advocated a catchment management approach to address the sedimentation problems at Water End. Measures suggested include regrading banks and the installation of drop structures and gravel traps, while a trash screen immediately upstream of the site may reduce the litter problem (Roberts, 1989; Darby and Thorne, 1992). Such a restoration scheme of the Mimms wash would function as a silt trap and improve the water quality of the motorway runoff entering the brook. Also, the restoration of a meander belt in the channelized reach would increase water and sediment storage potential.

The swallow holes at Water End represent a unique site in lowland southern England on the interface of the Chiltern Chalk and the London Clay and Reading Beds. Evidence suggests that the swallow hole complex may be contracting in response to changing sediment and water fluxes from the catchment. Mitigation measures to prevent the blockage of remaining holes through environmental deterioration are needed at the catchment scale to preserve this important site.

Conclusion

Swallow holes are solutional features which develop in limestone rocks. These are good examples of such features but they are also in a unique geological setting in southern England. The swallow holes have developed in the bed of the Mimmshall Brook, but recent changes in the water and sediment flow from the catchment threaten to block them. A restoration scheme will help to prevent this.

AYSGARTH, NORTH YORKSHIRE (SE 014887)

Highlights

Aysgarth Falls, in the Ure valley of north Yorkshire, is a spectacular example of a knickpoint in the long profile of a river. Glaciofluvial processes occurring during deglaciation have eroded less resistant rock to leave steep stream gradients, creating a series of cascades, Aysgarth Force, High Force and Middle Force.

Introduction

Aysgarth represents a classic example of a stepped fall. It developed where a knickpoint was held up by relatively resistant bedrock. The site is also important in the context of the deglaciation of the area. The Devensian deposits at Bear Park, originally described as a moraine, have also been interpreted as a complex of glacio-fluvial deposits associated with glacial meltwaters, which may also have contributed to the formation of the present morphology of the falls.

Description

The valley of the Ure lies geologically in a general downwarp between an anticlinal area to the north of the Swale and the half dome to the south. Aysgarth represents one of the many conspicuous waterfalls that occur where knickpoints have been held up by hard rocks. Well-developed drumlins occur in Upper Wensleydale as far down the main valley as Aysgarth, where till deposits become less evident, except on the spurs where tributary valleys enter Wensleydale. Over 80 large drumlins can be seen in the 20 km west of Aysgarth.

The river Ure cuts through a broad belt of mounds of debris crossing the valley at Aysgarth at about 215 m OD, and now flows through a deep rock gorge, above the sides of which are a few exposures of the glacial deposits.

Above Bear Park is a possible lake flat 5 km long and 1 km wide, with the present surface at 195 m OD, and some of the mounds reach between 4 m and 12 m above this level. This part of the valley is a true rock basin, and the gorge that drains it is one of the finest gorges in the district. The river descends 60 m in less than a mile by the cascades

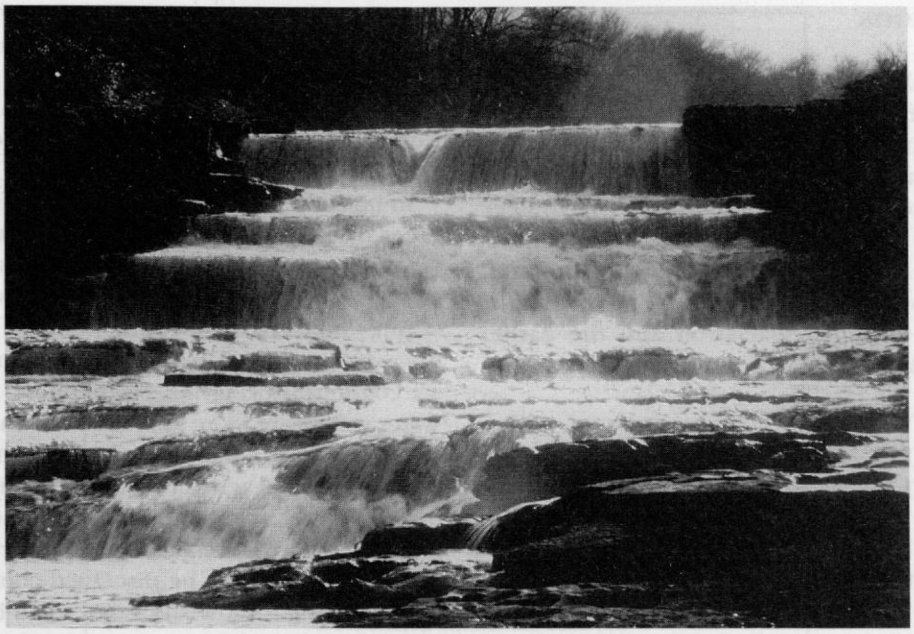

Figure 6.39 Lower Force, Aysgarth Falls. (Photo: I.D. Hooper.)

of Aysgarth Force, High Force and Middle Force (Figure 6.39). The valley of Upper Wensleydale hangs considerably above Bishopdale, the tributary valley from the south.

Interpretation

The development of Aysgarth Falls has to be seen in the context of the deglaciation of Wensleydale. During the last glaciation this dale, like other Yorkshire dales, was occupied by local glacial ice. At an early stage in the erosional history, it was suggested by King (1935) that at the location of Aysgarth Falls there was a steeper part of the valley long profile. It may be that this fall and the constriction of the valley facilitated the separation of masses of ice and led to the further development of Aysgarth Falls.

Originally, it was proposed by Raistrick (1926) that the stages of deglaciation of Wensleydale could be interpreted in relation to an active ice margin that retreated so that a series of moraines could be identified across Wensleydale. Raistrick proposed how each of these moraines could be related to lateral moraines and ice marginal drainage channels that he identified on the sides of the dale. He suggested that the Bear Park deposits were a major recessional moraine related to one of the stages of recession of the ice margin. He argued that the size of the moraine indicated a long pause in ice margin recession, during which the ice in the upper valley must have been eroding little, if at all. In contrast, Bishopdale and the lower Ure valley were being scoured by the meltwaters of the upper area.

However, in the valley downstream from Aysgarth Falls, there is a complex of glacio-fluvial landforms. This is made up of eskers and kettles which combine to produce a pattern of features which indicate that, when they were produced, the ice in the valley floor must have been stagnant. Subsequently, downstream near Wensley, these glacio-fluvial features are succeeded by a series of low terraces.

More recent investigations of these glacio-fluvial features, (e.g. by Cullingford and Gregory, 1978) have proposed that the pattern of deglaciation was characterized by the stagnation of the valley glaciers. The development of stagnant ice was influenced by the stepped morphology of the sides of Wensleydale, which led to the progressive isolation of masses of ice that were comparatively thin. Aysgarth Falls is significant in this pattern of deglaciation because the water draining within and beneath the stagnant ice must have flowed over the present location of the falls. It therefore seems likely that erosion by meltwater during deglaciation contributed to the morphology of Aysgarth Falls.

Conclusions

Aysgarth Falls is an important feature in the interpretation of the deglaciation of Wensleydale. The deposits within which the falls are set have been explained in relation to a major recessional stage of the local ice; and more recently they have been interpreted in the context of the decay of stagnant ice blocks. In the development of such features, the contribution of erosion by large volumes of meltwater during deglaciation and the earlier effects of glacial erosion must be considered.

DOVEDALE, NORTH YORKSHIRE (SE 872911)

Highlights

The valley fill deposits along Dovedale Griff and Jugger Howe Beck provide important evidence of alluviation and terrace formation that are likely to have been caused by accelerated slope erosion. Wetter conditions during the Atlantic period and the effects of forest clearance since the Mesolithic have been suggested to explain sedimentation in these small headwater catchments of the North York Moors.

Introduction

Together with other palynological and pedological evidence from the North York Moors, the valley fill deposits at these two sites record environmental conditions during the Atlantic period, coincident with known human vegetation disturbance. The sedimentary deposits, palaeochannels and terrace and fan sequences of the two sites exemplify Holocene valley alluviation and the difficulties involved in the explanation of the causes of alluviation. Support for interpretation is derived from organic material from a number of locations within the sites.

Description

The North York Moors constitute a broadly anticlinal Jurassic upland with mean annual rainfall of 700 - 1200 mm and mean annual temperature of 8 - 9°C. At the southern edge is an escarpment at 200 - 350 m in the south-dipping upper Jurassic Corallian Sandstone and Oolitic Limestone Series. Northwards is a central core of heather-covered moorland at 300 - 450 m underlain by middle Jurassic Estuarine Sandstones. Dovedale is in the Corallian Series and Jugger Howe Beck in the Estuarine Series.

In upland catchments, gullies, fans and low terraces of sand and gravel are often attributed to vegetation disturbance (e.g. Crampton, 1969; Harvey *et al.*, 1981). However, upland valleys constitute high-energy sediment sinks in which reworking, removal and replacement of sediment are continually occurring. Furthermore, small headwater valleys are more susceptible to the effects of localized meteorological events, and the marginal nature of upland agriculture may result in more frequent abandonment and re-use. Further data are required before a common pattern of response to human influence can be discerned comparable to that emerging for lowland valleys, and before conceptual models of that response can be evaluated. Although Late Holocene hillslope destabilization may be a common feature in the history of upland valleys, its precise timing and cause may be more variable than is the case in lowland valleys. For example, Iron Age forest burning is invoked as the cause of aggradation in south-east Wales (Crampton, 1969), while overgrazing by sheep is thought to have triggered gully-and-fan development in the Howgill fells (Harvey *et al.*, 1981).

Atherden (1976) has correlated palynological evidence from several bog sites within 10 km of both Dovedale and Jugger Howe Beck, to produce a regional picture dominated by three phases of woodland clearance, each separated by forest regrowth: (1) 2280 ± 120 BP; (2) 1060 ± 160 BP; (3) 390 ± 100 BP.

On the moors, surface erosion is presently very sensitive to the state of the heather biomass. There is evidence of culturally induced hillslope instability and erosion, which might be expected to have influenced the evolution of valley floor sediment storage. Inevitably, however, the history of sediment accumulation and erosion in a relatively high-energy valley floor environment is difficult both to date and to relate unambiguously to causal influences.

Interpretation

Dovedale is a deeply incised north-south-trending valley, cut in the Corallian Series. The 1 m wide Dovedale Griff drains the valley, on the eastern

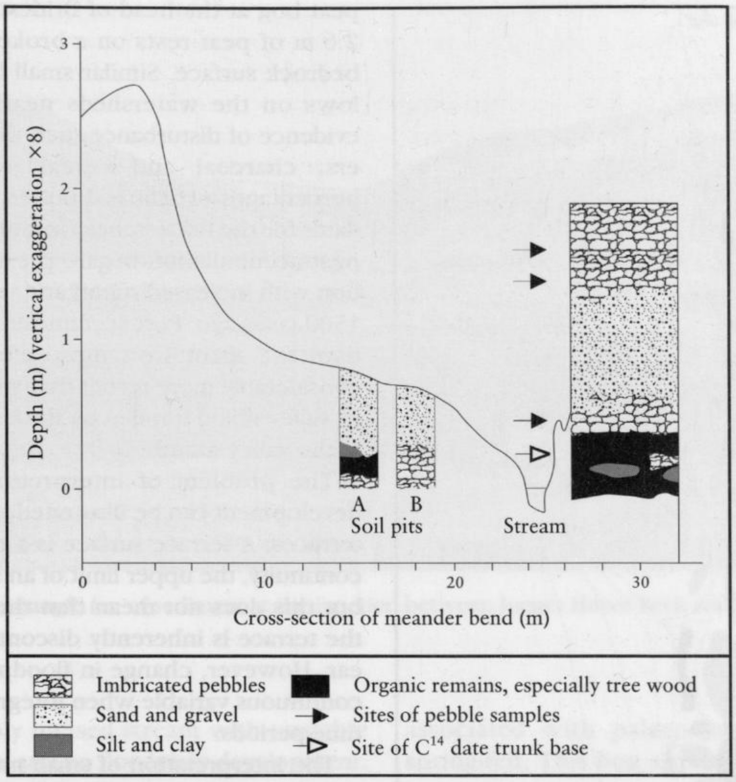

Figure 6.40 A cross-section of the meander bend on Dovedale Griff, and the schematic stratigraphy of the cut bank deposits. (After Richards, 1981.)

slopes of which are found the tors and free faces of the Bridestones (Palmer, 1956). It joins the east–west flowing stream at Staindale at SE 87259046, on the southern side of the main valley after crossing valley fill sediments. The amplitude of the meander bend 30 m upstream from this junction is increasing as the Dovedale Griff incises into the fill, and migrates laterally; bank collapse is evident at the apex (Richards, 1981). Here it has exposed sub-fossil wood at the base of the bank. The wood includes oak driftwood and, apparently, *in situ* stumps of willow and alder. An alder stump is dated at 6200 BP. Fluvial gravels approximately 1.5 m thick have accumulated above the wood (Richards, 1981).

The stratigraphy (Figure 6.40) consists broadly of two units. At the base, at the present stream elevation, is a unit consisting of the sub-fossil wood in a mixture of sand, pebbles and clay lenses which appears to be a slough deposit. Three samples of 50 sandstone pebbles from the upper unit show significant differences between the two coarser outcrops in this unit: the median intermediate axis lengths were 37, 38 and 43 mm. The fluviatile (rather than colluvial or solifluvial) origin of the pebbles is suggested by their rounded edges and imbricated structures. The observed directional properties suggest that deposition by a prior Dovedale Griff is more probable than by Staindale stream. The Holocene sediments burying the dated wood correlate with the terrace within Dovedale, which itself represents the remnants of alluviation probably associated with the slope erosion which formed the Bridestones (Palmer, 1956).

The valley floor sediments are fluvial sands and gravels underlying three distinct small-scale terraces with traces of palaeochannel on their surfaces (Figure 6.41). At the junction of Dovedale and Bridestone Griffs, a west-bank gully has deposited a pronounced alluvial fan which has experienced dissection and renewed deposition. The outer fan correlates morphologically with the upper terrace surface, and the inner fan of coarser sediment grades to the middle terrace. The multiple terrace sequence and hypothesis of basal stream rejuvenation together suggest that complex response mechanisms may have been significant in Dovedale, in addition to environmental disturbance.

Although the available evidence indicates that

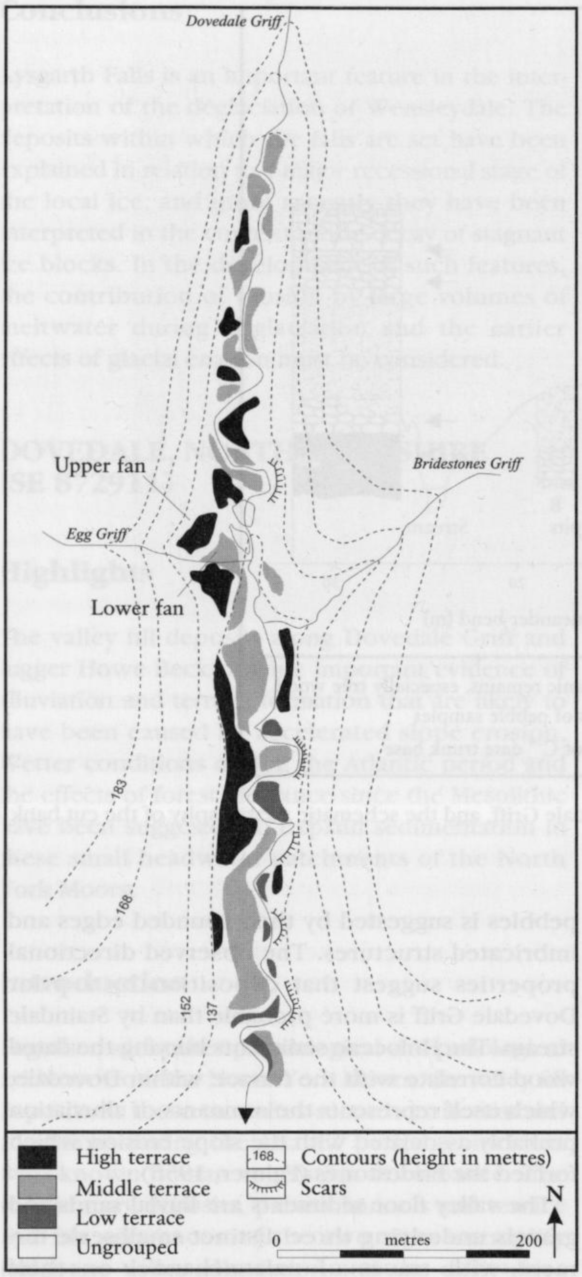

Figure 6.41 Dovedale: a morphological map of the terrace and fan fragments based on air photograph analysis and field survey. (After Richards, 1981.)

sedimentation in Dovedale occurred in the Late Holocene, it is sufficient to associate this unambiguously with human impact, to pinpoint the ages of the three terraces, to relate them to successive causal influences, or to define the cessation of the valley filling.

Further evidence of environmental disturbance and catchment instability is provided by a small peat bog at the head of Bridestones Griff, where 2.6 m of peat rests on a broken and weathered bedrock surface. Similar small bogs occur in hollows on the watersheds nearby, and all show evidence of disturbance such as sandy inwash layers, charcoal and cereal pollen, with low percentages of arboreal pollen at the base. Basal dates for the Bridestones Griff bog suggest that the peat accumulation began, presumably in association with increased runoff and waterlogging, about 1500 years ago. Peat accumulation appears to postdate the main Iron Age deforestation, and is considerably more recent than the date of initiation of valley filling implied by the buried alder stumps at the valley mouth.

The problem of interpreting morphological development can be illustrated using the Dovedale terraces. A terrace surface is a morphological discontinuity, the upper limit of an aggradation phase, but this does not mean that the process creating the terrace is inherently discontinuous or non-linear. However, change in floodplain elevation is a continuous variable when integrated over sensible time periods.

The interpretation of small multiple terraces, as the product of a complex response to a single external stimulus, usually involves downstream control by base-level change (Born and Ritter, 1978), as envisaged by Palmer (1956). This is then followed by coupling of main and tributary streams, with tributary sediment supply steepening the main valley fill gradient, until renewed incision is triggered. However, the response of the valley floor to hillslope erosion, which is evidently more important in Dovedale with its active hillslope scars and upper slope tors, requires a different conceptual model. If sediment supply is delivered directly to narrow headwater valley floors from adjacent hillslopes, areal rather than linear sources, then after basal rejuvenation the downvalley reaches receive accelerated sediment input before the upvalley ones, and the mechanism for floodplain steepening which then triggers renewed downcutting is less obvious. Under such circumstances, and more generally when there is catchment-wide control of the erosion processes, it may be more appropriate to interpret valley fill development as a continuous response to varying hillslope conditions.

Jugger Howe Beck, description and interpretation

The reach of Jugger Howe Beck is at and below the entrance of the right bank tributary, Hollin Gill, a

Figure 6.42 The low-angle fan of sediment at the junction between Jugger Howe Beck and Hollin Gill. (Photo: R.J. Davis.)

narrow and deeply incised stream with unstable bare hillslope scars and no floodplain development. A low-angle fan of coarse, flaggy, angular sediment exists at the junction, and at least two palaeochannel alignments are visible on its surface (Figure 6.42). The fan is clearly a compound feature, and intensely podzolized soils on its surface suggest that its general structure may be Late Devensian or Early Holocene in age. The main valley floor has a well-defined floodplain, with degraded marginal terraces and meander scars at the base of the hillside.

One palaeochannel alignment on the Hollin Gill fan is now intersected obliquely by the present channel, and organic-rich clays succeeded by angular gravels have been deposited in the palaeochannel. The dates correspond with that defining the Viking–early Medieval vegetation disturbance seen in the fen bogs, 5 km to the west, and the nature of the Hollin Gill catchment suggests that rapid response to such disturbance was likely. Thus there is evidence of a phase of sedimentation in an abandoned channel on the existing fan surface, during a period of known vegetation disturbance. Nevertheless, evidence of more continuous floodplain sedimentation rates would be more useful than such a discrete episode.

Two hundred metres downstream from the Hollin Gill junction there is a small peat bog on the right-bank floodplain surface of Jugger Howe Beck. This is one of several such bogs, some of which are associated with palaeochannels and are now spring-fed. This bog appears to have been more extensive, since peat outcrops as a continuous layer in the river bank for about 10 m downstream, having been exposed by lateral migration, bank undercutting, and compensatory bank-attached bar growth opposite.

Evidence from Jugger Howe Beck supports and complements the evidence from Dovedale. Maximum ages of 1150 years BP and 260 years BP are indicated for two soils in Jugger Howe and Hollin Gill Becks. These suggest that the middle Dovedale Terrace is likely to have a maximum age of around 1500 years BP.

Episodes of rapid sedimentation evident in the bank exposure (Figure 6.40, units B and D) are replicated in the stratigraphy of a core taken from the centre of the bog, as are the peats forming units A and C. The site therefore provides evidence of more general floodplain accretion. Floodplain sediments and soils exposed in the river bank upstream towards Hollin Gill junction display similar buried organic horizons and immature surface soils.

The evidence for rapid phases of sediment accumulation and floodplain development from this site appears to coincide with the last and very rapid vegetation disturbance evident in the regional pollen diagrams at about 250 years BP. This is associated with significant development of moorland management practices involving heather burning, which Imeson (1974) has shown strongly accentuate sediment delivery from the hillslopes today. Results indicate that 1 m of floodplain accretion occurred over about 50 – 100 years.

Evidence of slope erosion correlates with the indication of alluviation provided by the Dovedale site, although these deposits do not permit inference of the time span over which sedimentation occurred.

Conclusions

Until recently there have been comparatively few sites at which the Holocene history of landform development has been related to phases of human activity. At these two sites, valley fill morphology and sedimentology have been dated and related to archaeological evidence to link human vegetation disturbance with sediment movements within the catchments. They represent rare cases in which the human impact on upland valley sedimentation has been investigated and show similar patterns of response to those that have been found in lowland Britain. The geomorphological features of interest in the sites also include alluvial fans, palaeochannels and multiple terrace sequences.

RIVER CULM AT REWE, DEVON (SX 951995)

Highlights

The River Culm, at Rewe, Devon is an active depositional environment in which overbank river flows deposit large amounts of fine sediment on the floodplain. Evidence indicates that this is an important mechanism in floodplain development, with sediment deposition rates varying spatially and temporally.

Introduction

The floodplains of lowland rivers are regions of considerable geomorphological, environmental and economic importance (Nicholas and Walling, 1995). Well-developed floodplains may be an important sink for suspended sediment and this can be investigated through an understanding of contemporary rates of floodplain deposition (Walling and Bradley, 1989). New techniques have been developed to assess contemporary overbank sedimentation rates and patterns which, together with the use of hydraulic and sediment transport models, have allowed the gap between form and process studies to be narrowed. The River Culm is an important site for contemporary floodplain deposition, with its spatial variations in sedimentation rates.

The River Culm, which joins the Exe about 3 km north of Exeter, has a catchment area of 276 m^2. Between Cullompton and Stoke Canon, the river meanders across a floodplain (Figure 6.43) averaging 450 m in width (Lambert and Walling, 1987). The channel has a gravel bed and is approximately 12 m wide, while the banks are up to 1 m high and composed of fine alluvial material (Walling and Bradley, 1989). Overbank flooding is relatively frequent during winter months, significant inundation occurring generally seven times a year.

The depth of flooding varies from 40 cm during mean annual flood to 70 cm during a 50-year flood (Walling and Bradley, 1989).

The site at Rewe is part of a 13 km reach which

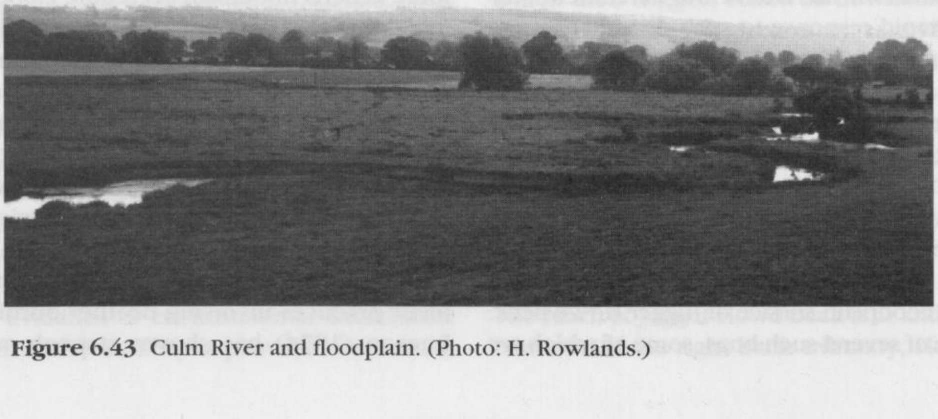

Figure 6.43 Culm River and floodplain. (Photo: H. Rowlands.)

is instrumented by two gauging stations, at Woodmill and Rewe (Lambert and Walling, 1987), providing continuous monitoring of river stage (for discharge) and suspended sediment concentrations. A number of techniques have been used to assess rates of temporal and spatial overbank sedimentation, including the analysis of sediment load data (Lambert and Walling, 1987), the use of sediment traps (Lambert and Walling, 1987; Walling and Bradley, 1989) and the use of caesium-137 concentrations as an indicator of fluvial activity (Lambert and Walling, 1987).

Description

Analysis of the gauging station records shows that an average conveyance loss of 28% occurs between Woodmill and Rewe (Lambert and Walling, 1987). In Figure 6.44 is illustrated a typical flood event whereby floodplain inundation between the gauging stations has attenuated the flood peak and resulted in a net loss of suspended sediment, on this occasion estimated as at least 374 tonnes. By assessing sediment load data, bank erosion within the reach and tributary input from the River Weaver, estimates of conveyance loss can be made. Hooke (1977) described the Culm as a relatively stable stream with slow but steady bank erosion with rates up to 0.2 m yr^{-1}. On Hooke's (1977) study reach about 18% had changed its course over a 100 year period of which 40% had changed in characteristics, sediment composition was considered a major factor affecting the spatial distribution of erosion. Sediment supply from this source has been estimated at 150 t yr^{-1} and inputs from the River Weaver at 600 t yr^{-1}, which for the period November 1982 to May 1994 gives an average annual sediment conveyance of 2500 t yr^{-1} (Lambert and Walling, 1987). Assuming an average area of inundation of 5 km^2 this approximates to 0.5 mm yr^{-1} deposition (Walling and Bradley, 1989).

Deposition rates can also be measured directly using sediment traps both for individual storms (Lambert and Walling, 1987) and for longer periods (Walling and Bradley, 1989). Both experimental design strategies have demonstrated temporal and spatial variation at individual sites. For example, for the period December 1986 to February 1988, total sedimentation rates vary by an order of magnitude between different sites (Park Farm 17 278 gm^{-2}) and within sites (sites at Rewe had a range of 286 - 1319 gm^{-2}) (Lambert and Walling, 1989). Micro-scale variations in sedimentation rates relate to surface topography, and to the incidence of depressions.

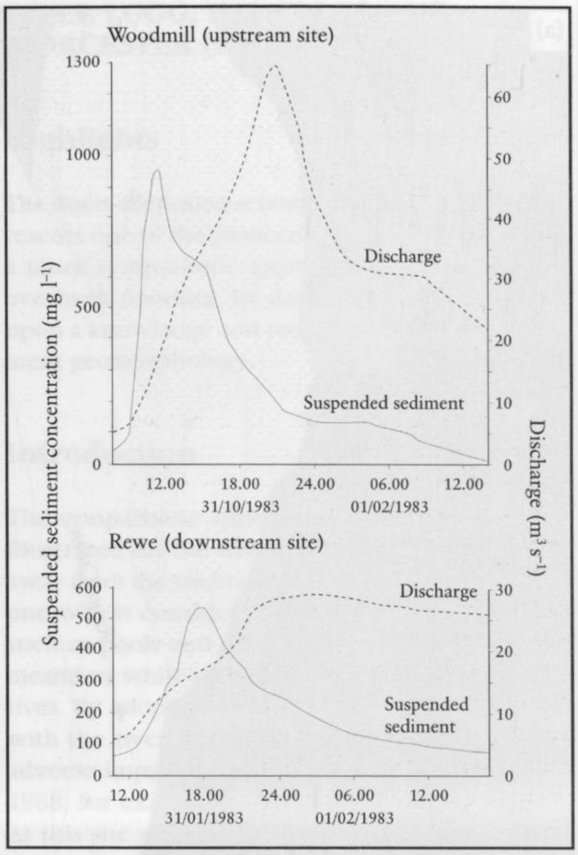

Figure 6.44 Conveyance loss and flood attenuation, River Culm. (After Lambert and Walling, 1987.)

Distributions of caesium-137 fallout from atmospheric testing in the late 1950s and early 1960s can be used to assess rates of deposition. By comparing floodplain samples with control samples, taken from above the flood limit, the level of Cs-137 'excess' can be used as a measure for both deposition and erosion. From 120 site samples throughout the reach, Walling and Bradley (1989) measured an increasing depth and rate of sediment deposition downstream in response to the increased width and decreased slope of the floodplain. For example, near Rewe and Stoke Canon rates up to 13 mm yr^{-1} were recorded. Micro-variation in deposition at Rewe, derived from 74 cores, is illustrated in Figure 6.45. Two small depressions close to the channel have the highest rates of sedimentation as they experience a backwater effect, while further from the channel a decline in the levels of caesium-137 'excess' represents reduced inundation and hence lower rates of deposition.

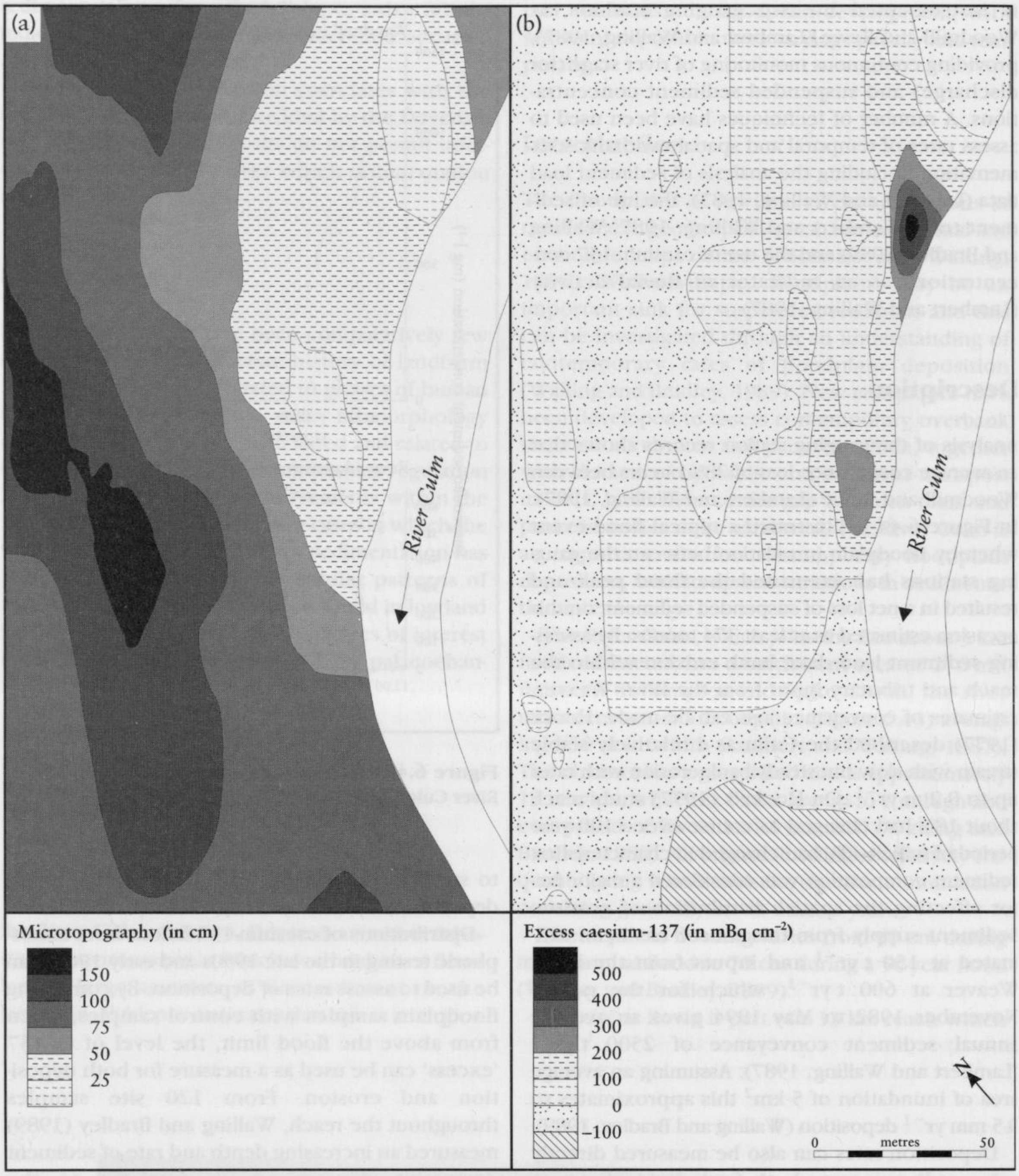

Figure 6.45 Microtopography and excess caesium-137, River Culm floodplain at Rewe. The contours on (a) refer to heights in centimetres above an arbitrary datum representing the lowest point on the floodplain surface within the sampled area. (b) The pattern of excess caesium-137. (After Walling and Bradley, 1989.)

Interpretation

Nicholas and Walling (1995) have now coupled a hydraulic model with a sediment transport component to model the dispersion of suspended sediment away from the channels during inundation. For their study reach at Rewe, a reasonable degree of correspondence was obtained between measured and predicted rates, with a maximum value of 1.0 – 1.5 mm yr^{-1} estimated for abandoned channel fill sites.

Significant spatial differences in sediment deposition occur at both the macro- and micro-scale. These can be related to variations in floodplain morphology and the frequency and level of inundation, while at the micro-scale small variations in surface topography control the velocity, duration and depth of inundation. Greatest deposition rates occur in closed depressions and small backwater embayments, while scour is also evident from caesium-137 measurements in secondary channels which convey flow on the floodplain during inundation. Analysis of the floodplain and suspended sediments showed that sand and coarse and medium silt fractions are preferentially deposited, although a significant proportion (>50%) of clay was also deposited, possibly reflecting the importance of aggregates in the natural transport process (Walling and Bradley, 1989).

Research along the River Culm illustrates the nature of contemporary overbank sedimentation and the magnitude of conveyance losses along a lowland floodplain river system. Fine alluvial deposits are widespread among the floodplains of many lowland rivers in the UK, and have been interpreted as evidence of long-term overbank deposition due to man-induced accelerated erosion in the Late Holocene (Brown, 1983; Burrin, 1985). Contemporary rates of deposition have been shown to exert considerable spatial variation with rates of up to 17 mm yr^{-1} reported (upstream of Rewe) against average rates of 1.5 mm yr^{-1} (Walling and Bradley, 1989). Through the use of hydraulic and sediment transport models, measured rates of deposition can now be predicted which will lead to a greater understanding of the processes of overbank sedimentation.

Overbank sedimentation is an important process on lowland streams. This is one of the few sites in Britain, and indeed internationally, where the processes have been studied in detail. New techniques have been used which have allowed measurements of contemporary rates and distribution of sedimentation. It is also a site at which longer-term studies of forms and processes provide a context for the interpretation.

Conclusion

This is a site that is becoming internationally important for innovative measurements of overbank depositions. Significant spatial variations in deposition are revealed and can be related to the floodplain topography and frequency of flooding.

RIVER LUGG, HEREFORD AND WORCESTER (SO 466612)

Highlights

The flood alleviation scheme on the River Lugg represents one of the pioneering approaches to adopt a more sympathetic approach to the control of overbank flooding, by designing a scheme based upon a knowledge and understanding of the catchment geomorphology.

Introduction

The sympathetic approach to river management illustrated on the River Lugg represents a move away from the traditional engineering approach to one which considers integrating natural features such as pools and riffles, berms and benches and meanders while still achieving flood control objectives. By adopting an approach that aims to work with the river rather than against it, many of the adverse impacts of channelization (see Brookes, 1988, for examples) will hopefully be minimized. At this site such effects have been recognized by integrating morphological and ecological aspects of management in the scheme through the following concepts (Lewis and Williams, 1984):

1. a stream and its floodplain are parts of the same open system;
2. a balanced view must be taken of economic, fluvial and ecological values;
3. incorporation of existing research knowledge should be undertaken.

The important aspects of fluvial geomorphology and hydrology to be included in the design include the existence of convergent–divergent flow; the complex relationships of sediment transport in streams; and the existence of geomorphic thresholds.

Description

The River Lugg floodplain was often inundated by prolonged flooding, reducing the productivity of the floodplain (RSPB *et al.*, 1994). One reach was characterized by an active meandering system which increased downstream sediment load and led to sedimentation downstream at Leominster,

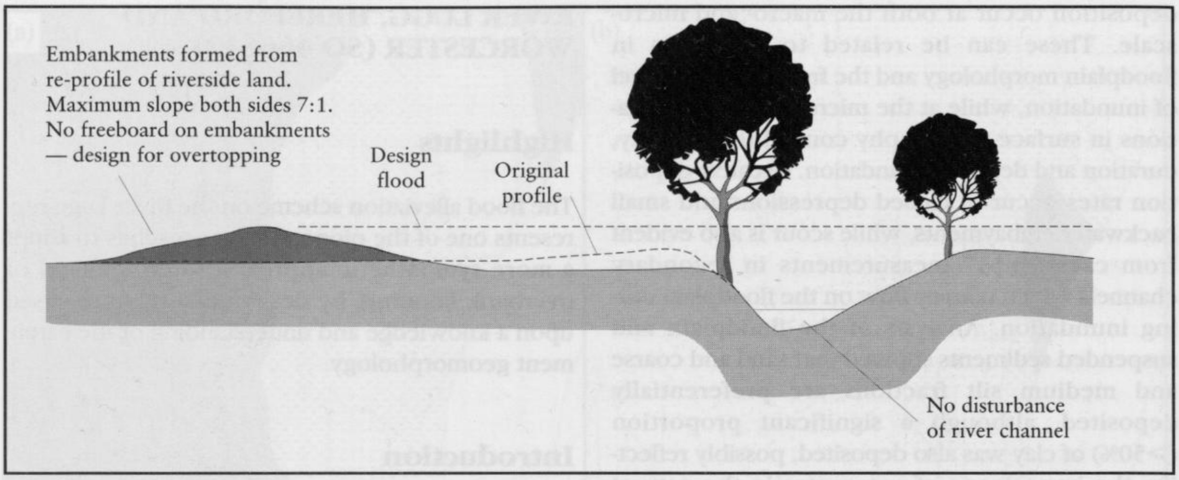

Figure 6.46 The re-profiled floodplain section, River Lugg. (After RSPB *et al.*, 1994.)

associated with a subsequent reduction in gradient. In order to address these problems, a number of features were included in the scheme, which was constructed in 1980 – 81:

1. Low, flat, roll-over embankments set well back from the river channel to allow meander migration using spoil from the reprofiling of the land intervening between the riverbank and embankments (Figure 6.46). This provides a sufficient cross-sectional area to convey the 1 in 5 year flood.
2. Meander re-profiling, where no meanders were removed, but the inside of each was reprofiled to a natural shoal (Figure 6.48) and on active bends willows were planted to stabilize exposed banks and assist in habitat creation (Figure 6.47).
3. Low stone weirs (Figures 6.49 and 6.50): four of these were constructed entirely of rock with relatively flat aprons. They were designed to reduce flood velocities by creating still-water reaches upstream to dissipate energy in the river. Secondly, they are attractive landscape features, and they provide a new stable rocky habitat in a stretch which otherwise is composed of unstable sediments. Each weir was separately designed, varying according to both the location and the size of available materials. A typical plan and section is shown in Figure 6.49, illustrating the apron of blocks, shallow crest and long downstream glacis, with an end sill and several stones raised above the general level as breakwaters, the whole contained within blockstone wing-walls. The weirs are uncemented, and consist of armouring stones and gravel over either permeable or impermeable membranes.

The approach taken along the Lugg was not universal, with areas of high ecological interest being excluded from the scheme while others received higher standards of protection for flood control (RSPB *et al.*, 1994). For example, upstream of the reach in Figure 6.47, traditional engineering approaches were used to construct concrete weirs with fish passes, and channel straightening was undertaken to create long sinuous curves of trapezoidal cross-sections.

Interpretation

An appraisal of the scheme was undertaken in 1993, confirming that the primary aim of reducing flooding incidence had been achieved (RSPB *et al.*, 1994). The embankments downstream of Lugg Bridge (Moreton-on-Lugg) had been overtopped occasionally but without undermining stability; however, floodgates have been added to allow the water to return to the river as the discharge recedes. The hydraulic performance of the schemes has also been studied on behalf of MAFF (MAFF/UEA 1991; MAFF/HR Ltd FR312 Nov. 1992 — cited in RSPB *et al.*, 1994).

Upstream of Leominster, above the reach illustrated in Figure 6.47, the channel has remained featureless, with the fish passes prone to blockage by debris. In the lower section (Figure 6.47) major pools have developed below each weir with

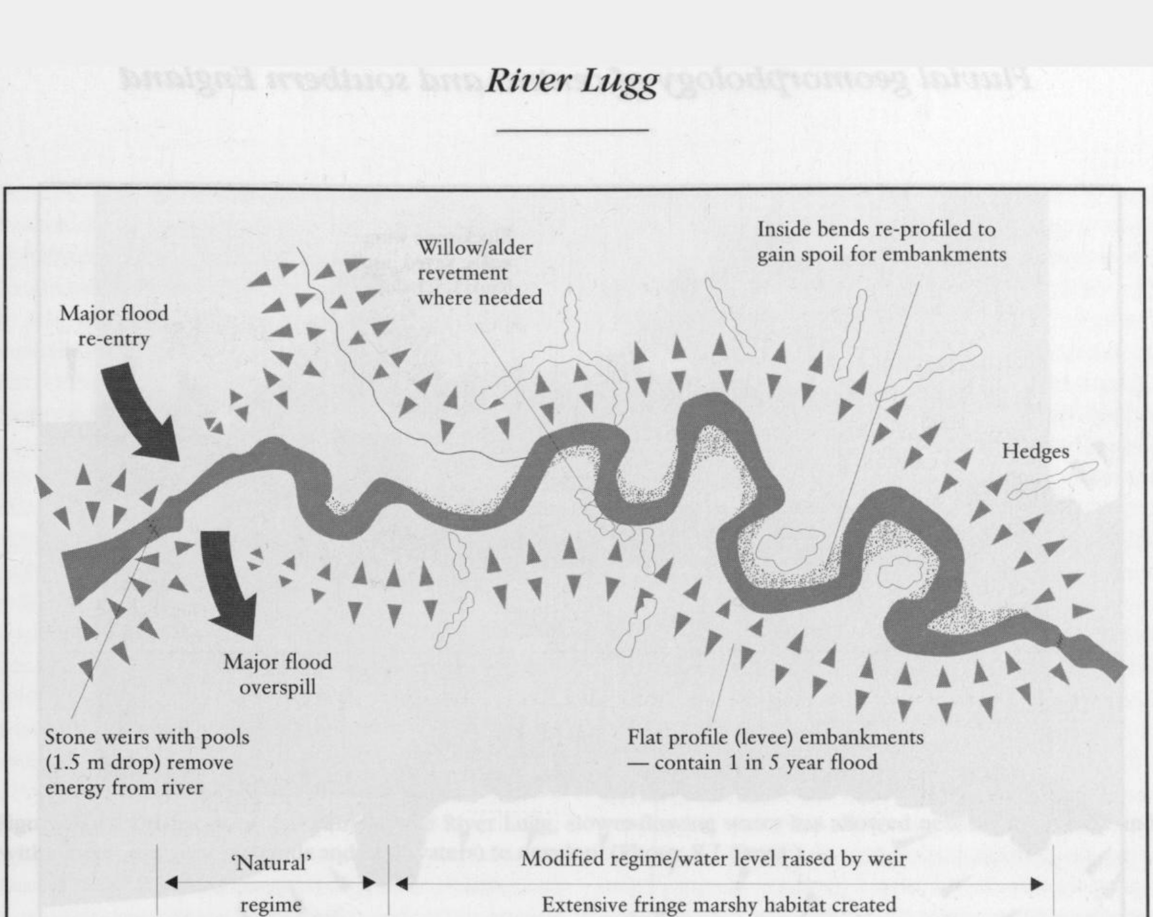

Figure 6.47 Flood alleviation work combined with channel stabilization activities, River Lugg. (After RSPB *et al*., 1994.)

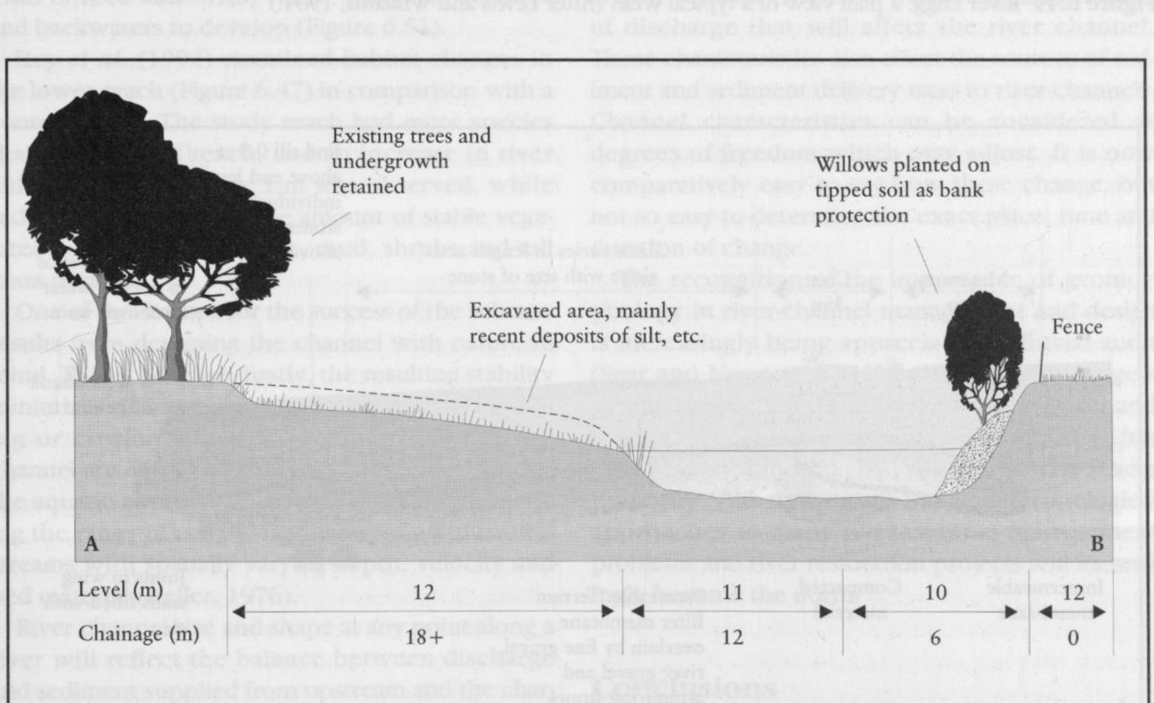

Figure 6.48 River Lugg: a typical cross-section at a re-profiled meander. (After Lewis and Williams, 1984.)

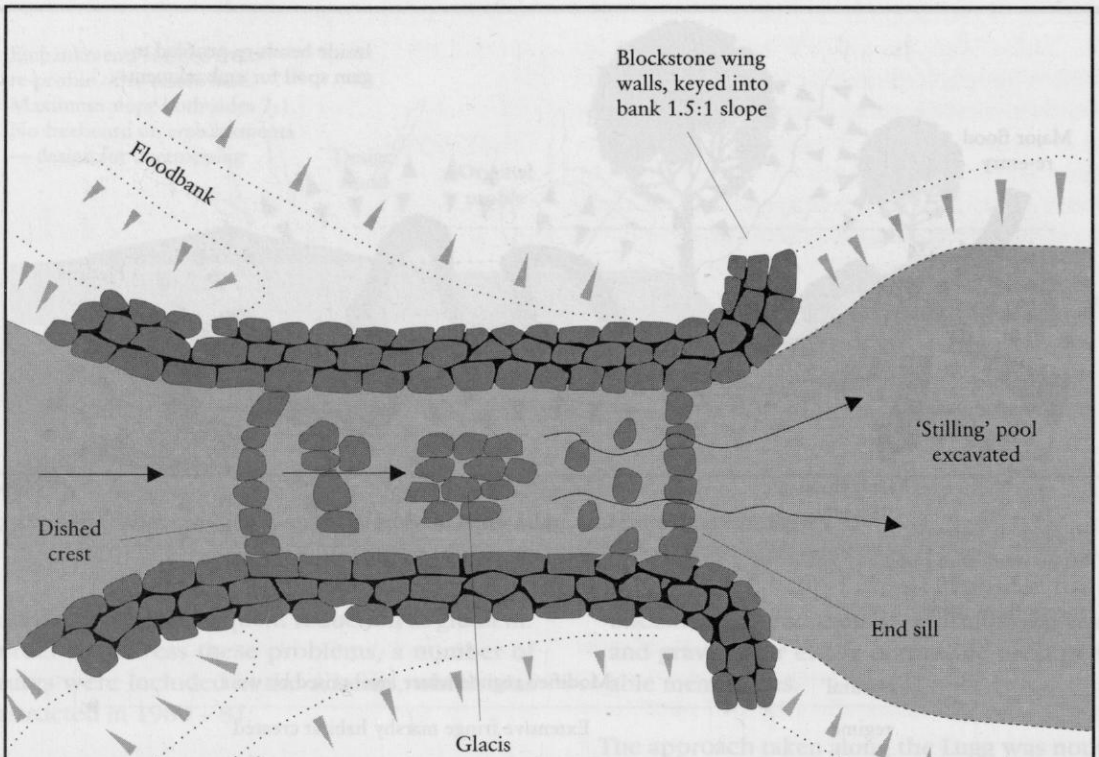

Figure 6.49 River Lugg: a plan view of a typical weir. (After Lewis and Williams, 1984.)

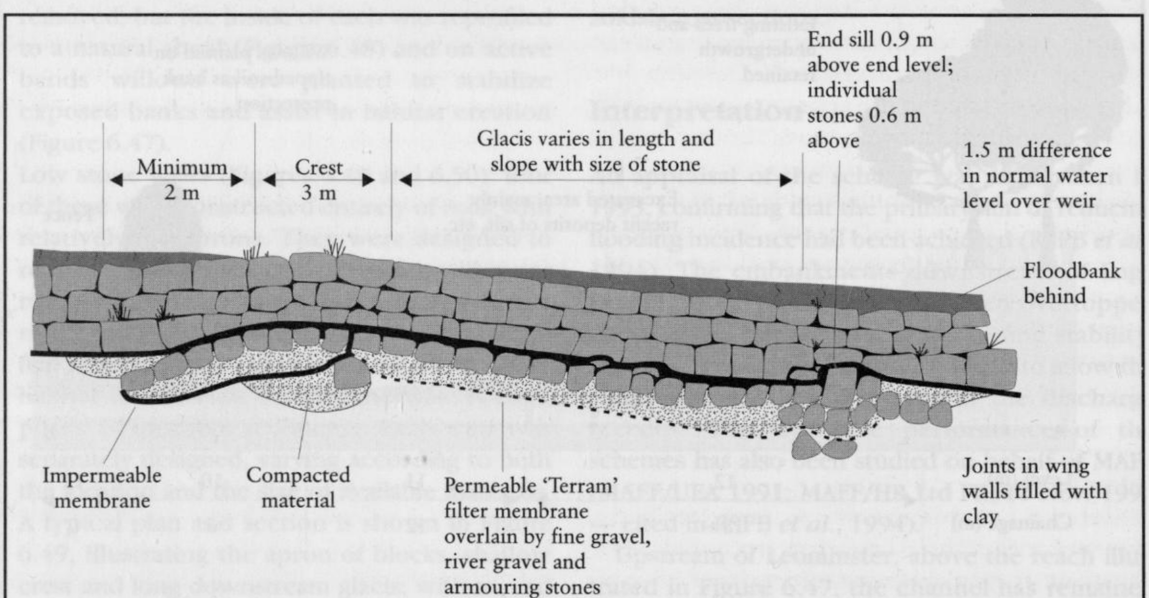

Figure 6.50 River Lugg: a long sectional view of a typical weir. (After Lewis and Williams, 1984.)

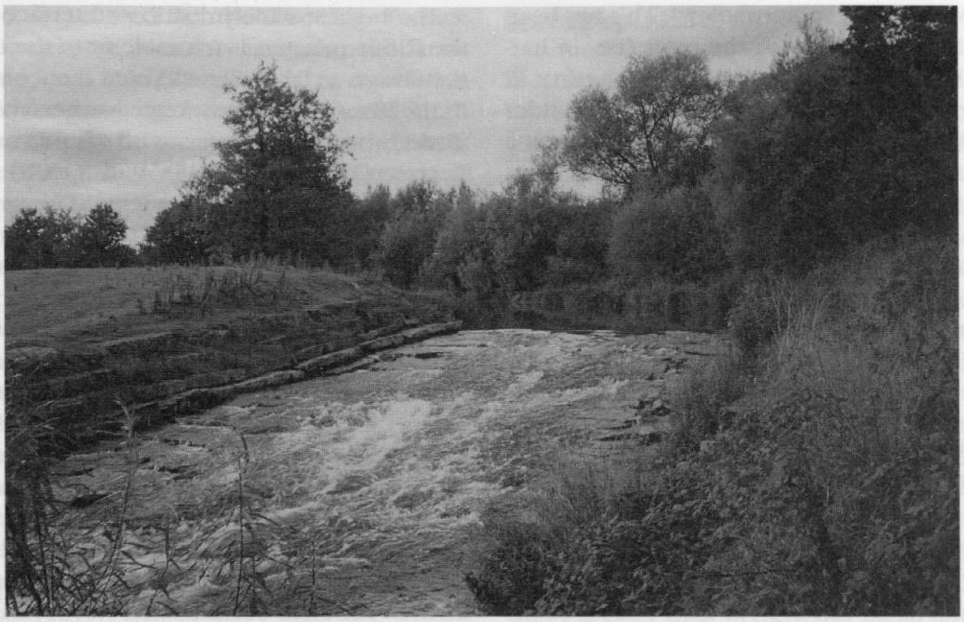

Figure 6.51 Upstream of the weirs on the River Lugg, slower-flowing water has allowed new habitats (reed and 'withy' beds, marginal wetlands and backwaters) to develop. (Photo: R.J. Davis.)

associated downstream riffle formation; in some cases it has been necessary to remove shoals which have been created. Upstream of the weirs, the slower-flowing water has allowed habitats in the form of reed and 'withy' beds, marginal wetlands and backwaters to develop (Figure 6.51).

Hey *et al.* (1994) examined habitat changes in the lower reach (Figure 6.47) in comparison with a control reach. The study reach had more species than the control reach, and an increase in river cliffs and berms above 1 m was observed, while reductions occurred in the amount of stable vegetated bank, shingle margin, mud, shrubs and tall grass present.

One of the reasons for the success of the scheme results from designing the channel with nature in mind. This is because, firstly, the resulting stability minimizes maintenance costs necessitated by silting or erosion. Secondly, the aesthetics of the channel are enhanced (Leopold, 1969) and, thirdly, the aquatic ecology can be preserved by maintaining the range of ecological niches found in natural streams with spatially varying depth, velocity and bed material (Keller, 1976).

River channel size and shape at any point along a river will reflect the balance between discharge and sediment supplied from upstream and the characteristics of bed and bank material, vegetation and slope at that point. Discharge will depend upon the size of the drainage area, the frequency and character of precipitation as rain- and snow-fall and the drainage basin characteristics of rock, soil, land use, vegetation and topography. The interaction of such drainage basin characteristics determines how rapidly climatic effects are translated into a pattern of discharge that will affect the river channel. These characteristics also affect the sources of sediment and sediment delivery rates to river channels. Channel characteristics can be considered as degrees of freedom which may adjust. It is now comparatively easy to say how these change, but not so easy to determine the exact place, time and duration of change.

The recognition of the importance of geomorphology in river-channel management and design is increasingly being appreciated. A fluvial audit (Sear and Newson, 1991), which provides a geomorphological methodology for identifying and solving river-channel management problems, has now been adopted by the National Rivers Authority. This now means that geomorphological approaches to many river-channel management problems and river restoration projects will increasingly become the norm.

Conclusions

The River Lugg provides an important example where river management has been based on principles of fluvial geomorphology. This has been described as working with the river (i.e. in harmony with fluvial processes) rather than against it. The design of the scheme has included meander easing and construction of floodbanks in such a manner as to minimize subsequent levels of management. Post-project appraisal of the site has indicated that the scheme has achieved its objectives. This demonstrates that flood alleviation schemes can benefit from a geomorphological input which will provide the morphology of enhanced habitat formation and colonization.

WILDEN, HEREFORD AND WORCESTER (SO 823730)

Highlights

This is a key site with major remnants of the Severn Terraces, which have been the basis for reconstruction of depositional conditions and the sequence of changes of the river at a significant stage of its development after deglaciation.

Introduction

The site is of particular importance because it relates to a critical time in the history of the Severn valley at about 15 000 BP. At this locality, examples of each of the major terraces of the Severn exist. The terrace deposits are preserved most extensively south of Ironbridge Gorge where four major-, and up to three higher-level, terrace remnants have been identified (Wills, 1938; Beckinsale and Richardson, 1964). The lower terrace levels are traceable over considerable distances before plunging beneath the modern floodplain south of Gloucester. The sequence covers the period from the warming of the early Windermere Interstadial to the cold conditions of the Loch Lomond readvance.

Description

Wilden lies on the Worcestershire Stour, near its junction with the Severn at Stourport, and in this region all four terraces of the Middle and Lower Severn (in descending order, Kidderminster, Main, Worcester and Power House) are well developed.

The highest widely distributed terrace deposit, the Kidderminster, is traceable from the mouth of the Severn as far as Bewdley and thence upstream to the River Stour. The terrace has been correlated directly at Tewkesbury with the Avon No. 4 Terrace (Tomlinson, 1925). Wills (1938) deduced from the distribution of the terrace fragments and the erratic content that the Kidderminster Terrace pre-dated the Devensian Irish Sea glaciation, arguing that the absence of terrace fragments from the Severn Valley upstream of the Stour confluence was evidence for a Devensian glacial origin for the diversion of the Severn through the Ironbridge Gorge. It is important to note that the Kidderminster Terrace is strikingly developed along the Stour, but not along the Severn upstream from Stourport. This shows that the reach of the Severn between Stourport and Ironbridge must have been insignificant until the time of development of the Main Terrace.

The most extensive deposit is the Main Terrace which is traceable from Apley Park near Bridgnorth, where it is approximately 30 m above the present floodplain, to Gloucester, where it descends beneath the modern alluvium. This deposit has a variable width and thickness, but in places has a cross-valley extent in excess of 1 km and a thickness of up to 10 m. Wills divided the deposit into two units; a lower Main Terrace running into Ironbridge Gorge (related to Avon No. 2); and an upper terrace unit which he associated with No. 3 Avon Terrace. However, there is little evidence for the presence of multiple levels in other areas and Shotton (1953) has shown that the correlation of an upper surface with the Avon No. 3 Terrace is unlikely. The terrace spread south to the south of the Ironbridge gorge and the typical erratics, from the Irish Sea, SW Scotland and the Lake District that were transported by the Late Devensian (Irish Sea) glacier, are represented. Radiocarbon dates from No. 2 Terrace of the River Avon, which grades into the Main Terrace at Tewkesbury, suggest that this event was not earlier than 25 000 BP and possibly later, although no single absolute date has yet been determined on any of the terrace deposits along the main river.

The surface of the Worcester Terrace lies approximately 8 m below that of the Main Terrace, and is traceable from Tewkesbury to Bewdley, and also in the Bridgenorth and Bewdley area. Wills (1938) correlated the terraces with the Avon No. 1 Terrace and with the Uffington Terrace north of the Ironbridge Gorge. However, Coope and

Shotton (1981) argued that the Worcester Terrace has no correlative in the Avon basin, and Shaw (1969) and Jones (1982) suggested that correlation of the terrace fragments between north and south of the gorge may be unsound, due to the likely presence of stagnant ice in the area downstream of the Uffington Terrace at the time of its formation. Shotton (1977) and Shotton and Coope (1983), have proposed that the Worcester Terrace is an outwash deposit of the Late Devensian glaciation, although Wills (1938) and Poole and Whiteman (1961) attempted to relate the terrace to a level of Lake Lapworth. On the basis of its erratic content, it has been proposed that the Worcester Terrace could have been associated with the Ellesmere readvance (Beckinsale and Richardson, 1964; Shotton and Coope, 1983).

Altitudinally below the Worcester Terrace, there are a number of discontinuous low terraces. Wills (1938) collectively named these the Power House Terrace, and proposed that they represented the upper part of an infilling of a channel deeper than the bed of the present river. Williams (1968) recognized a sand and gravel unit up to 12 m thick under most of the lower Severn that Brown (1982) considered to be the first depositional unit of the Lateglacial or postglacial valley fill. Below Worcester, these deposits underlie the Holocene alluvium, but in the vicinity of Bridgnorth the low terrace deposits lie approximately 8 m above the current floodplain surface and at Stourport the Power House Terrace is evident up to 2.5 m above the floodplain. A Holocene floodplain sequence is also recorded at Wilden (Wilden Marsh, 50 m north of the Power House Terrace exposure) with 3 m of basal gravels, wood peat and fen peat capped by silt and clay. The sediments have provided an unusually complete Holocene pollen record which includes a flooding episode, hiatus *c.* 3000 BP and increased overbank deposition of silt and clay over the last two millennia (Brown, 1988).

The Power House Terrace cannot be regarded as a unitary sedimentary body and dates are not necessarily applicable elsewhere. However, at Stourport the culmination of gravel deposition has been radiocarbon dated at 12750 ± 220 BP (Shotton and Coope, 1983). The Power House Terrace is also visible at Hartlebury Common Nature Reserve (NGR 50820750) as a distinct surface. At Harlebury Common there is also a palaeochannel (Rush Pool) in the terrace surface, infilled with peat deposited through most of the Holocene. There is, however, probably a mid-Holocene hiatus in the sequence (Brown, 1984).

Interpretation

In a study of the terraces of the lower Severn, including the Main and Worcester terraces (Dawson and Gardiner, 1987), the sedimentary characteristics were needed to indicate the character of a deposition of the sediments. Thus at the time of deposition of the Main Terrace, flow is interpreted to have been in a low sinuosity that, at least locally, was dominated by large channels containing sandy bedforms and bars (Dawson, 1985).

Several approaches were employed to estimate palaeodischarges. The calculations (Dawson and Gardiner, 1987) indicated that mean annual flood discharges during an initial stage in the deposition of the Worcester Terrace were four to seven times greater than similar discharges during the formation of the Main Terrace. It seems that Worcester Terrace mean annual flood discharges were 3.8 times greater than present-day mean annual flood values at Bewdley. It was concluded (Dawson and Gardiner, 1987) that palaeodischarges were greatest during the deglacial period, increasing after the formulation of the Main Terrace to a maximum prior to the aggradation of the Worcester Terrace, when there was a period of high sediment availability.

On an eroded surface of the remnant of the Power House Terrace, there are organic deposits that have a maximum date of about 13 000 BP and which contain a typically Lateglacial insect fauna. A similar Lateglacial deposit occurs upon Avon No. 1 Terrace gravels at Fladbury, giving a maximum possible date for the Power House/Avon No. 1 Terrace.

Organic deposits of this age occur in cutoff channel deposits in the valley bottom here (Shotton and Coope, 1983), suggesting that this time the rivers adopted a meandering mode within lateral cut and fill. The organic deposits lie upon an uneven surface of the gravels, filling some sort of hollow or channel. If, however, the river is followed below the point where the new course joins the original channel, it is seen to cut into a gravel bank at least 2 m high, bounding the conspicuous terrace remnant that lies to the south. This exposed material of the Power House Terrace is made entirely of gravel with no trace of organic deposits. This is further evidence of the irregular nature of the top of the gravels and strengthens the case that there is an appreciable lapse in time between their deposition and any overlying peats and silts of Lateglacial age.

Regularization of the River Stour entailed cutting a section through the Power House Terrace, which stands only 2.6 m above the alluvial level. In places, coarse gravel reaches the terrace surface, as can be seen where the river cuts naturally into the terrace. However, the diversionary cut made by the then Water Authority revealed a broad channel filled with sand, silts and peats, going in places below the present river level, and lying upon the terrace gravel. Seven radiocarbon dates led to the conclusion that the channel infilling commenced about 13 000 years ago and gave way to inorganic sand close to 10 000 years ago. Evidence from insect — mainly beetle — remains strongly supports the dates, and the fauna indicate climatic conditions at their warmest at the older date, not becoming really cold until the latest samples.

Conclusion

Starting with the time when the final surface of the Main Terrace formed the floodplain, not earlier than 25 000 years ago, and finishing with the deposition of the Wilden gravels prior to being overlain by channel deposits from 13 000 BP, there was an astonishingly rapid sequence of events in not more than 12 000 years, and conceivably in as little as 5000 years (levels refer to Stourport):

1. erosion into the Main Terrace at 43.5 m OD down to the rock base of the Worcester Terrace at 22.9 m OD (−20.6 m);
2. aggradation of the Worcester Terrace gravels to 31.1 m OD (+ 8.2 m);
3. erosion into the Worcester deposits down to an estimated base at 13.4 m OD (−17.7 m).