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Floodplain evolution in the East Midlands, United Kingdom: the Lateglacial and Flandrian alluvial record from the Soar and Nene valleys

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Contents

This paper describes and interprets the floodplain stratigraphy of two low-energy rivers in the East Midlands, U.K. and goes on to present an appropriate model of floodplain evolution that may be applicable to other lowland rivers in temperate mid-latitudes. Work has been undertaken at the reach level to try and characterize the entire system. In general only laterally extensive exposures have been used, and detailed stratigraphic and microstratigraphic recording has been used to facilitate

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process interpretations. All the sites have been radiocarbon dated. The dataset consists of three sites in the Soar valley and five sites in the larger Nene valley. The most common sedimentary architecture is a basal gravel with occasional shallow channels, covered by a mixed loam (sand–silt–clay) of variable thickness with landsurfaces and evidence of bioturbation. This loam is frequently interrupted by palaeochannels, and finally all these sediments are buried by a massive silt–clay unit. Most of the palaeochannels were abandoned in the early Flandrian, or between *ca.* 3500 years BP and 2000 years BP. The superficial silty clay is diachronous largely because of the irregular topography of the floodplain onto which it was deposited; dates from non-palaeochannel sections range from *ca.* 3500 BP to *ca.* 2100 years BP. The floodplain stratigraphy has been profoundly influenced by soil development and sub-aerial processes, especially tree-throw, which has produced distinctive sedimentary structures. The loam unit is interpreted as a soil which developed in-between both silting palaeochannels and active channels. The landsurfaces are both earlier and contemporaneous with the later phase of channel abandonment. There is some evidence, *ca.* 5000–4000 BP, of a rise in floodplain watertables. An evolutionary model is proposed which can account for the stratigraphic evidence reported here. It is based upon the development of an anastomosing (stable multiple-channel) system from an initial braided-river topography and its eventual conversion to a predominantly single-channel system due to floodplain and channel siltation. The reduction of channels is compensated for by a change in channel types and capacities while the floodplain aggrades; this is the stable-bed aggrading-banks (SBAB) model, which necessitates no changes in discharges. It is suggested that the sub-meandering or straight to sinuous nature of many lowland U.K. channels may be due to their evolution from an anastomosing pattern where the least meandering channels survived typically with a box-S shape planform and at the edges of the floodplain. The sites also show that the Lateglacial fluvial history of the two catchments seems to have been very different, with incision and subsequent aggradation occurring during the Younger Dryas in the Nene but not in the Soar. Given the proximity and similarity of the two catchments this suggests that relatively minor local factors may have been able to push some catchments across fluvial thresholds. In contrast the Flandrian history of the two rivers has been broadly similar, although there is evidence of greater lateral instability and floodplain reworking in the Soar which may be due to hydrogeological factors or a different landuse history. This work strongly suggests that new process-based interpretations of floodplain stratigraphy, and new models of floodplain evolution may be required before alluvial history can be easily related to the changing Lateglacial and Flandrian climate of lowland U.K.

1. Introduction

The work in this paper aims to describe and interpret the floodplain stratigraphy of low-energy floodplains in the East Midlands, U.K. and formulate an appropriate model of floodplain evolution which would also be of general applicability. Although single sites may provide a wealth of information on local floodplain development, qualitative and quantitative models of floodplain formation must be based upon investigations at the reach to catchment scale to account for the effects of sediment routing through the system. For this reason, although several catchments were initially investigated, only two were selected for detailed study and dating on the grounds that they alone had a sufficiently large number of exposures. These reaches

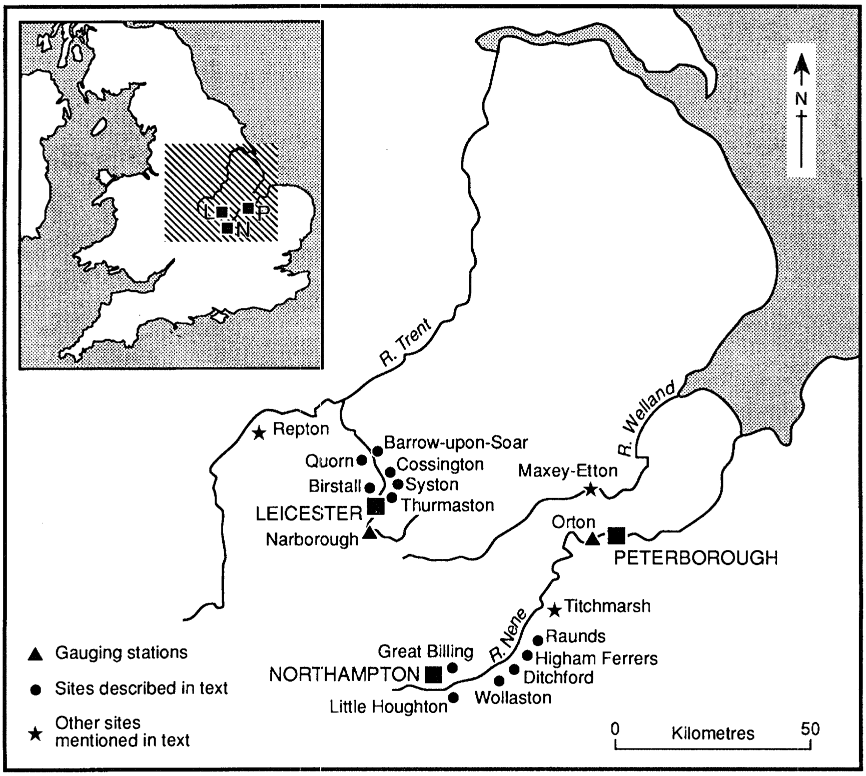

Figure 1. Location maps of the Nene and Soar catchments.

were the Soar from Leicester to the Trent and the Nene from Northampton to Peterborough (figure 1). It was essential to find sites that exposed the full depth of Lateglacial and Flandrian sediments as the techniques used included detailed logging with grids and frames, drawing of selected sedimentary structures and detailed excavation rather than standard logging of sedimentary architecture. This was done because the number of formal lithostratigraphic units encountered was small and it is the intra-unit and inter-unit sedimentary features that were of prime interest. For the same reason the primary use of auger-holes or boreholes was excluded and stratigraphic work limited to worked aggregate quarry faces that could be followed-back over a period of several years and also allowed the collection of as much organic material as possible for palaeoenvironmental reconstruction and radiocarbon dating. Particular attention was focused on the identification and investigation of palaeochannel scars, evidence of palaeolandsurfaces, and the sedimentary features of thin transitional sedimentary bodies. The sedimentary units described in this paper do not equate with beds in the strict lithostratigraphic sense, although they are all laterally extensive. Palaeochannel fills are considered to be part of the unit in which they occur or of the unit into which they grade upwards.

2. Models of lowland floodplain evolution

For relatively high to medium energy alluvial systems several generalized process models of floodplain evolution have been proposed. These range from equilibrium models based upon lateral channel migration (Wolman & Leopold 1957) and overbank and lateral deposition (Brackenridge 1984) to disequilibrium models based

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

upon episodic floodplain stripping (Nanson 1986). For low-energy systems several descriptive sedimentary models exist including those for single channel meandering systems (Allen 1965) and anastomosing systems (Smith & Smith 1980). Quantitative models have been formulated for floodplain evolution produced by avulsion and overbank deposition (Bridge & Leeder 1979) and overbank deposition alone (Pizzuto 1987). More recently there have been attempts to combine both lateral and vertical processes of sedimentation into a quantitative models of meandering river floodplain evolution (Howard 1992). Many of these models have either been inductive (e.g. Nanson's or Brackenridge's models), based on the process interpretation of observed alluvial sedimentary systems and their recent floodplain stratigraphy, or deductive (e.g. Pizzuto's model) in that they have started from fundamental principles and compared the resultant simulated stratigraphy to that observed in the field in order to test the models' 'goodness of fit' which will validate the assumptions made and approximations used. Although the former approach lacks quantitative rigour it does identify crucial processes and linkages that can be incorporated into quantitative process models.

An alternative approach has been to stress the importance of allogenic factors and view floodplains as indicators of changing climate, especially in uplands (Macklin & Lewin 1991; Macklin *et al.* 1992), or of catchment conditions (Trimble 1983). While this is undoubtedly valid, rivers and therefore floodplains do not respond simply or uniformly to these allogenic changes and this reinforces the need for process models of floodplain evolution at the reach and catchment scale that may identify which allogenic factors have been critical and which not. The ability, both to infer processes of floodplain formation and accurately test model simulations against reality, demands correct identification of the processes from sedimentary forms and structures. For this reason, and because we believe that none of the current models adequately explain the stratigraphy of the floodplains in the East Midlands, there is a need for a fresh examination of the evidence upon which a model of Lateglacial and Flandrian floodplain evolution can be based. Dating is extremely important in this work, not only because it provides a chronology, but because variations in floodplain stratigraphy are largely controlled by differential accumulation rates associated with different sedimentary processes.

3. The catchments and reaches

(a) *Relief and geology*

Both catchments lie on the eastern side of the Midland Plain of the U.K. (see figure 1), the Nene draining directly to the North Sea via the Wash, and the Soar draining to the North Sea via the river Trent and the Humber estuary. Both catchments are of low relief and have a relatively low and equable rainfall (see table 1). The solid geology of the Soar catchment is predominantly Mercia mudstone, although extensive areas in the east are also underlain by clays and limestones of Lower Jurassic age. Upstream of Peterborough the Nene catchment is underlain by Lias clay, Middle Jurassic Inferior and Great Oolite Series, Cornbrash and Oxford clay.

(b) *Geomorphological history*

Although neither catchment was glaciated during the Devensian, both retain substantial areas of glacial materials, including till and outwash, deposited during an earlier glaciation. However, these true glacigenic sediments are underlain in both

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Table 1. *Statistical comparison of the Nene and Soar catchments*

	area/ km^2	station height/ (m OD)	basin relief/ m	floodplain slope[c]/ ($m\ m^{-1}$)	areal rainfall/ mm	mean gauged flow/ ($m^3\ s^{-1}$)	% time flow less than/($m^3\ s^{-1}$) 95%	50%	10%
Nene[a]	1634	3.3	221	0.00049	622	9.01	34.4	7.0	2.8
Soar[b]	1292	32.0	189	0.00053	690	13.24	35.0	11.0	3.5

[a] Nene at Orton (TL 166972).
[b] Soar at Kegworth/Zouch (SK 492263).
[c] Slope over the study reach.

catchments by sands and gravels that appear to represent deposition by earlier trunk rivers; in the case of the Soar they have been termed the Baginton sand and gravel (Rice 1992) and, in the case of the Nene, the Milton sand (Horton 1970; Davey 1991). In both basins there is evidence that the pre- and post-glacial drainage patterns were significantly different, although the elevations of the bases of the Baginton sand and gravel and the Milton sand indicate that the regional levels of the trunk drainage lines were already only a few metres higher than those of the modern Soar and Nene. It can thus be inferred that deposition during the glaciation initiated a new drainage pattern, but the rivers subsequently excavated their valleys down to the old level. The stages in this downcutting are still poorly known, because such evidence as terraces is scanty and inadequately researched. In the nineteenth century at Barrow on Soar *Elephas antiquus* remains were recovered from an isolated patch of gravel (Plant 1859), but in general the only sites affording both stratigraphic continuity and the potentiality for dating lie beneath the modern floodplain. As shown below, these generally date from the mid-Devensian or later, so that there remains a major hiatus in our knowledge of fluvial history during pre- and early Devensian times.

(*c*) *Current hydrological conditions*

The lithological differences between the Soar and Nene catchments have relatively minor hydrological effects. Both rivers today have non-flashy régimes; the return period: discharge slope is marginally greater for the Soar than for the Nene (see table 1). As this table shows, the floodplain slope of the two reaches is comparable and the present channels are of low sinuosity; indeed the Nene often consists of two low sinuosity channels flowing at opposite edges of the floodplain. There have been major artificial channel works on both the Nene and Soar channel, cutting new straight sections and the creation or conversion of small channels for mill leets; there is no evidence that large secondary channels with meandering planforms were artificially created. The bi-channel pattern can therefore be regarded as a low slope, cohesive-bank, anastomosing system. Both rivers have been managed for over a century for navigation and flood control, and this has involved dredging, channel realignment and the construction of embankments and weirs.

4. Previous work

Relatively few sections in the deposits underlying the Soar floodplain have been recorded in the past half century, although numerous boreholes have proved the presence of one or two metres of silty clay overlying an equal or greater thickness of sand and gravel, which in turn rested on Mercia mudstone. The common assumption was that the sand and gravel represented a Devensian aggradation and the overlying finer material a Holocene overbank deposit; little refinement and elaboration of this basic chronology has so far been attempted. In 1972 Bell *et al.* described the sand and gravel at Syston and showed, from the composition of the clasts, that most of the gravel came from erosion either of earlier glacigenic sediments or of the Baginton sand and gravel. The age of the onset of aggradation was determined as 37420+1670−1390 years BP from radiocarbon assay of an organic lens near the base of the unit. The cold climatic régime characterizing the aggradation was inferred both from the biota preserved within the lens and from periglacial involutions that affected even the uppermost strata visible at this particular location. However, with the finer temporal resolution now being attempted, it is worth noting that there was no direct dating either of the involutions or of the main phase of the gravel except that it was post 37000 years BP. At Thurmaston, only a few hundred metres from the Syston site, Mathieson (1975) recovered organics from the finer sediments overlying the gravels. Of two radiocarbon dates obtained, one of 4220±130 years BP (BIRM-499) was from wood at *ca.* 1.5–2.0 m depth, and the other of 3720±130 years BP (BIRM-500) from peat at a depth of *ca.* 0.3 m. Until the present investigations began, these were the only radiocarbon assays providing information on the chronology of Devensian and Holocene aggradation along the Soar valley.

The sub-alluvial gravels and fine alluvial deposits of the Nene have received more attention than those of the Soar. Castleden (1976) showed that rockhead morphology beneath the floodplain is subplanar and in most locations does not have a large or deep channel buried beneath the gravels, although near Northampton Horton (1970) recognized a bedrock channel of glacial origin. From sedimentary structures that included involutions and ice-wedge casts, Castleden (1980) argued that sub-alluvial gravels were deposited on a fluvioperiglacial pediment occupying an older shallow valley. The gravels, generally planar and of medium grade, are composed chiefly of flint, chert, quartzite, Jurassic limestone and ironstone varying in shape from angular to rounded. He envisaged floodplain gravel aggradation (First Terrace) between 41000 and 14000 years BP, followed by vertical incision and formation of the First Terrace surface 11000 to 10000 years BP.

Holyoak & Seddon (1984) examined three sites containing fossiliferous deposits within the gravels and alluvial fines. At Little Houghton two palaeochannels with organic infills, lying at the interface of the bedrock and overlying gravels, were dated to the middle Devensian on the basis of their pollen spectra. Two palaeochannels at Titchmarsh revealed fossil biotas of contrasting ages. The lower channel within the gravels was estimated to be of late Devensian age, while the upper channel, cut into the gravels at the gravel-fine unit interface, was estimated to date from *ca.* 3000 years BP on the basis of its pollen spectra and mammalian fauna. At Orton Longueville a palaeochannel within the gravel unit also contained biota of probable late Devensian age. At this site it was also noted that tree roots penetrating the upper surface of the gravels were apparently truncated at the gravel fine-unit interface. Holyoak and Seddon questioned Castleden's (1980) alluvial chronology, pointing out the lack of supporting dates for the period between 25000 and 10000 years BP.

A radiocarbon date of 28230±330 years BP (BIRM-75) for gravel deposition was obtained by Morgan (1969) from organic clays within the lower layers of the suballuvial gravel at Great Billing. The oldest published date for cessation of gravel

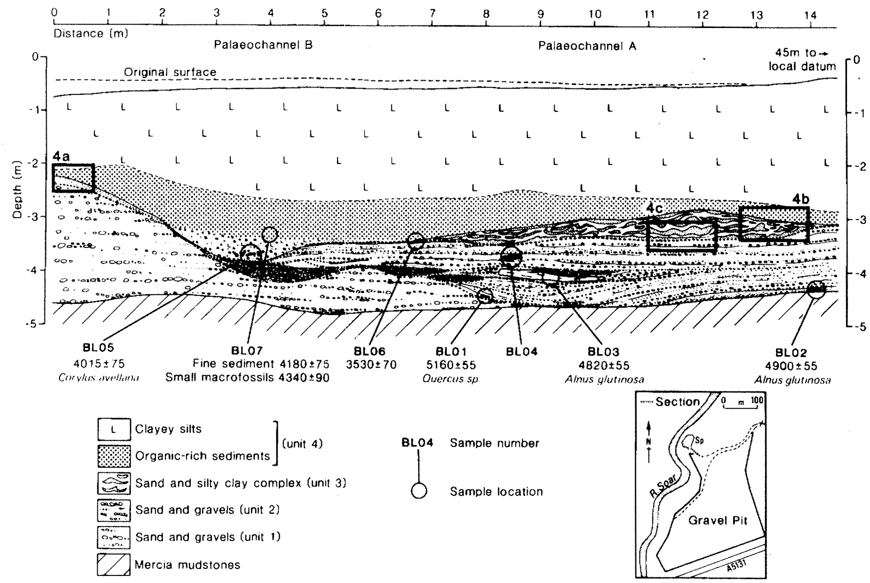

Figure 2. Stratigraphic section from Birstall. (Larger copies of the figures are available from the authors.)

deposition and accumulation of fine sediment is 9420 ± 70 years BP from wood at Wansford (Harkness & Wilson 1979). Felted plant material from near the base of the silty clay alluvium along a tributary of the Nene at Little Billing yielded a radiocarbon date of 3870 ± 55 years BP (Harkness & Wilson 1979).

The downstream sections of the Nene around Peterborough have been studied as parts of archaeological projects (Pryor 1986), and at Fengate such investigations have suggested rising water levels during the Middle Iron Age and flooding of the site in Late Iron Age. The contiguous Welland catchment in the Maxey–Etton area has yielded evidence of four generations of palaeochannels (French 1990). The first set consists of peat-filled meandering channels cut into the Devensian gravels, the second an anastomosing system of unknown date, the third a meandering channel system of probable Bronze–Iron Age date, and lastly a meandering river of medieval age.

5. The Soar sites

The stratigraphy of each site in the Soar valley will now be described, along with the dating evidence and palaeoenvironmental interpretation.

(*a*) *Birstall*

The site at Birstall (SK 598 084) is situated 5 km upstream of the Soar/Wreake confluence. The floodplain at this point is about 0.75 km wide and has a gradient of 0.0076. Gravel extraction over an extensive area of the floodplain exposed a large number of sections extending to the top 0.5 m of the bedrock. The main section (see figure 2) was extended laterally by using unpublished borehole (S.T.W.A) data and hand augering. There was between 4 m and 5 m of unconsolidated sediment overlying the Mercia mudstone with a sharp contact which is gently undulating or scalloped. The scallop-shaped features were between 2 m and 5 m wide and 0.2 m to 0.5 m deep and were presumably caused by fluvial erosion before, or during, the

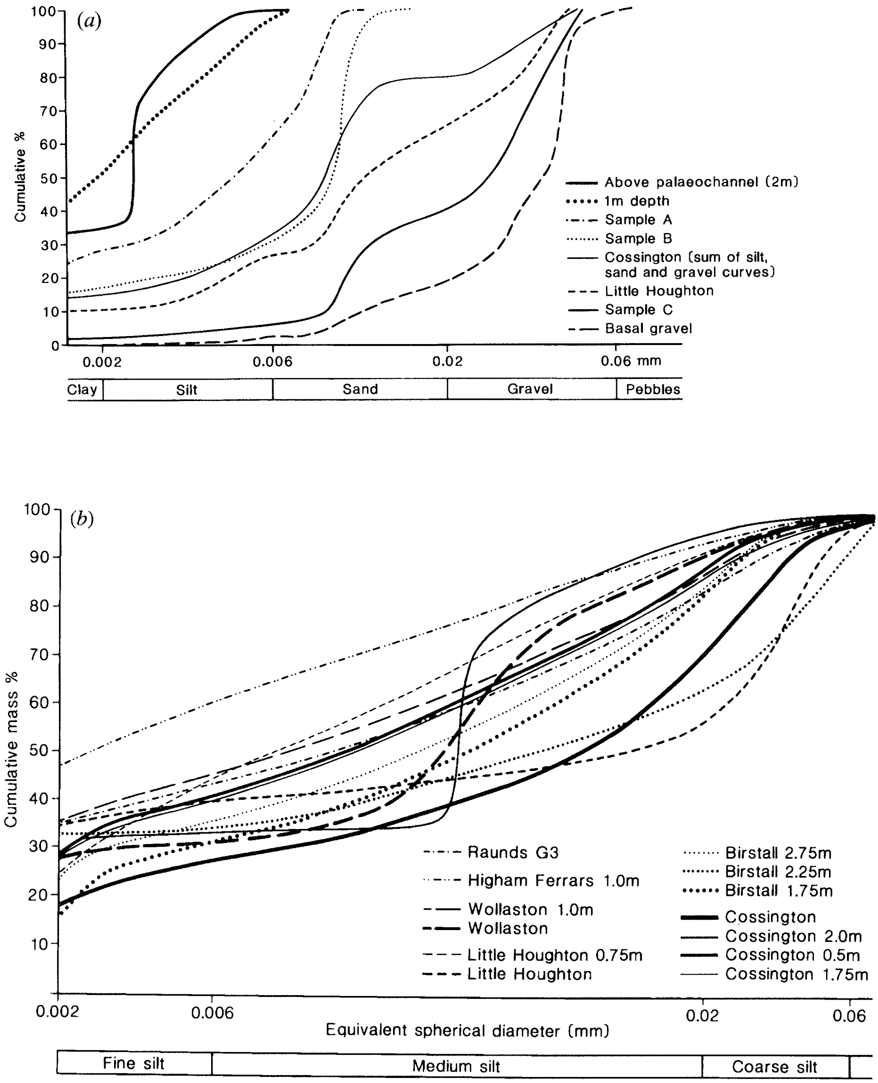

Figure 3. Grain size curves: (*a*) for coarse sediments (sieved), (*b*) for fine sediments (using a sedigraph). In (*a*) all curves are from Cossington except where stated otherwise.

deposition of the basal gravels. The unconsolidated sediments consist of five sedimentary units. The lowest unit was planar-bedded sand and gravel containing occasional clay drapes and lenses of cross-bedded sands. The unit was between 2.5 m and 2.8 m thick across most of the floodplain but thinned towards the margins and was occasionally absent. Each bed of gravel varied from 0.03 m to 0.25 m in depth and between 20 m and 40 m in lateral extent. The gravels were poorly sorted and ranged from cobbles to sands (see figure 3). The unit as a whole fined upwards and was generally capped by 0.20 m to 0.25 m of medium and coarse sand. Unit 2 contained a much higher proportion of finer grades and was more structured with woody macrofossils. However, it only extended over small areas of the floodplain where it was cut into unit 1. Within unit 2 were several thin beds ranging in thickness from 0.03 m to 0.10 m, each with a sharp boundary with the bed below and displaying a

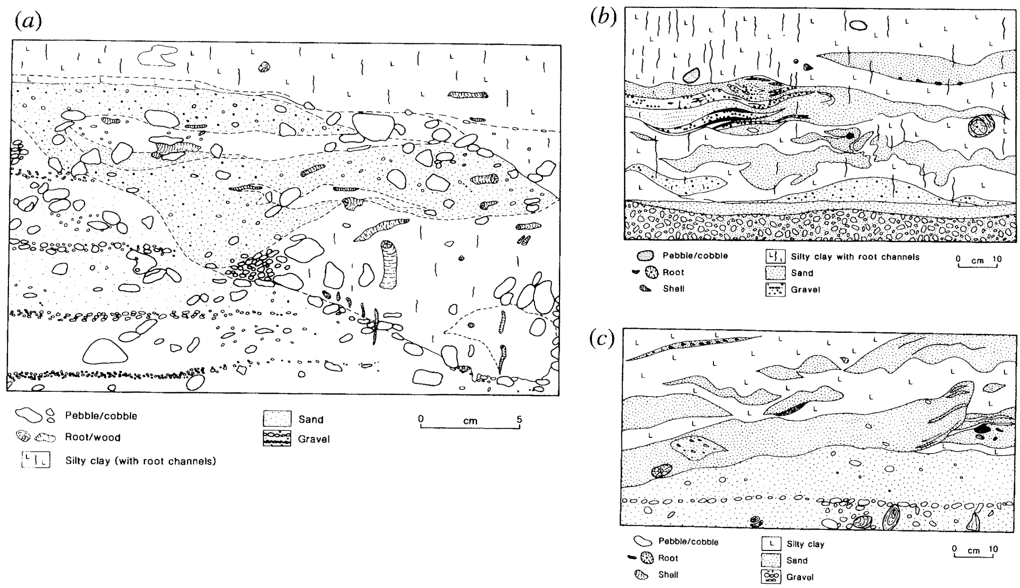

Figure 4. Detailed stratigraphy from Birstall: (*a*) the interface between units 1 and 2, (*b*) and (*c*) from unit 3 (see figure 2 for exact locations on the face).

distinct fining upward sequence from very well sorted fine gravels, through coarse, medium and fine sand laminations to a silty-clay drape at the top. The organic materials present were mainly small twigs, undegraded but too small to identify. They were always found in the fine sand fraction and were covered by the drapes. This suggests they were deposited as flow receded from a channel edge or point bar. Larger wood macrofossils were also found within unit 2 and could be identified and radiocarbon dated. These included pieces of *Quercus* spp., *Alnus glutinosa*, *Prunus* spp. and *Fagus sylvatica* and they ranged in apparent age from 5160 ± 55 years BP (SRR-3502) to 3530 ± 70 years BP (SRR-3507) with decreasing stratigraphic elevation. Also within unit 2 several lenses of framework gravels were located at a constant depth between 3.3 m and 3.7 m below the floodplain surface. These well sorted gravels had no internal structure and only one lens contained wood. This was a trunk of *Alnus glutinosa* over 2 m long and 0.24 m in diameter with no signs of abrasion and no evidence of decay. A radiocarbon assay from it gave a date of 4820 ± 55 years BP (SRR-3504). A third sedimentary feature that was found within unit 2 was a small channel infilled with silty clays, organic detritus and some sand. As well as the unidentified detrital macrofossils there were *in situ* roots of *Corylus avellana*. One root was dated at 4015 ± 75 years BP (SRR-3507) and showed evidence of decay. Figure 4*a* shows the interface between unit 1 and unit 4 and the disruption, chaotic fabric and abundant root wood associated with soil development and bioturbation probably by tree-throw. Unit 3 was a sandy and silty-clay complex composed of two different sediments, well sorted yellow sands and brown silty clay. These interdigitated and had highly disturbed wavy or involuted boundaries (see figure 4*b*, *c*). In some places small lenses of each sediment were found within the other; small woody macrofossils were found in both. Unit 4 was, at its base, an organic-rich clay blanketing the coarser deposits. The organic content decreased with increasing height in the unit (from 10% L.O.I. to 2% L.O.I.) as it graded into an

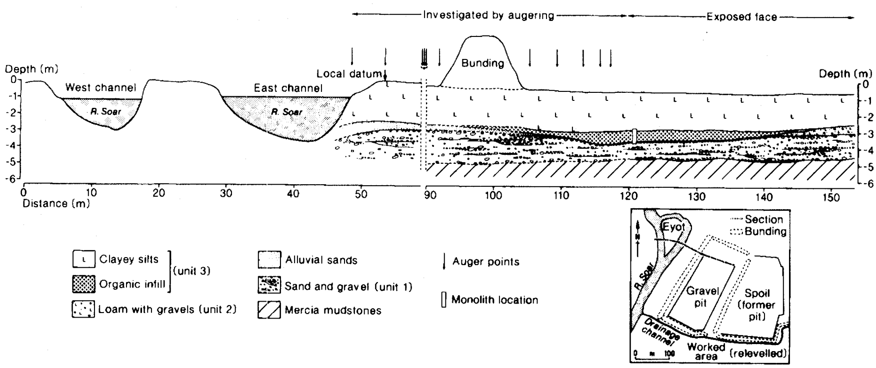

Figure 5. Stratigraphic section from Cossington Monolith location is also the location of the block-excavation illustrated in figure 7.

inorganic silty-clay. A radiocarbon date from the base of unit 4 gave a date of 4180 ± 75 years BP (SRR-3508) for the fine organic material and 4340 ± 90 years BP for the coarse plant macrofossils (SRR-3508). The superficial unit 4 is massive and structureless except for prismatic to blocky structure in the upper 1–1.5 m and crumb structure in the upper 0.2 m.

The section is interpreted as indicating (*a*) accumulation of Devensian gravels (unit 1) (*b*) erosion of these gravels sometime before the mid-Flandrian (*c*) aggradation (units 2 and 3) and channel migration away from the site and (*d*) fine overbank sedimentation (unit 4).

(*b*) *Cossington*

The site (SK 596 137) is located just downstream of the confluence of the Wreake with the Soar (figure 1), and the reach from which exposures were examined was 0.75 km in length and 0.8 km wide with a floodplain gradient of 0.00075. Several sections were logged as the extraction face was worked back. Radiocarbon samples were collected from one face which was typical of the stratigraphy of the site (see figure 5). The stratigraphy consisted of three sedimentary units. The lowest unit was the basal gravel which was a clast supported planar bedded sand and gravel which contained occasional clay drapes and lenses of cross-bedded sands. The unit was generally 1.5 m to 2.0 m thick but thinned to a metre or less in a few places. The constituent beds varied between 0.02 m and 0.20 m, were laterally continuous for 30 to 50 m, and composed of poorly sorted (see figure 3) sub-angular to well rounded clasts. A slight increase in sand was noticed up the unit profile. Above this, unit 2 comprised a gravelly loam of around 0.5 m thickness. The structure of this unit was complicated by lateral changes from a poorly sorted, structureless, clast-supported melange of sub-angular to sub-rounded medium gravel with numerous broken and fractured clasts to a finer matrix-supported gravelly loam with disrupted bedding which graded into thinly bedded silty sands with gravel stringers and clay drapes. In some areas this unit was composed entirely of horizontally bedded and cross-bedded medium to coarse sands. The overlying unit 3 had at its base an organic-rich clayey-silt and peat which reached a thickness of 0.75 m in some depressions that pinched-out the gravelly loam unit and cut into the underlying basal gravels. In these palaeochannel scars the boundary between the deposits was abrupt and marked by a gravel pavement (see figure 6). At the centre of one of the channel scars a block of

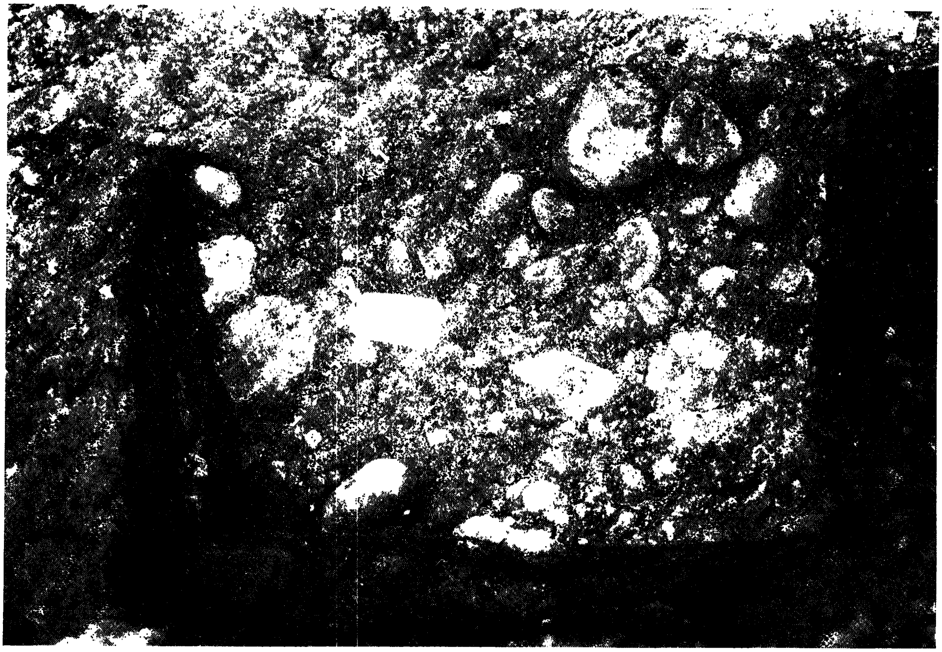

Figure 6. The excavated gravel pavement at Cossington.

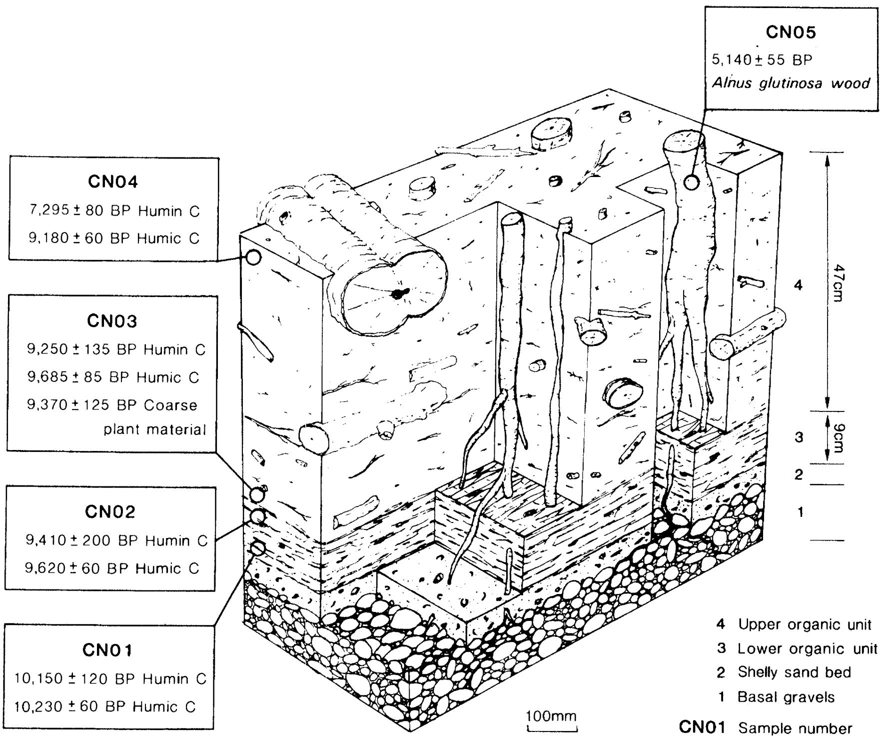

Figure 7. The 3D stratigraphy of the excavated block from Cossington (see figure 5 for location of the block).

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Figure 8. The excavated block at Cossington (see figure 7).

sediment (monolith in figure 5) was excavated in detail by hand-trowelling to establish the stratigraphic relations between the different organic components of the peat (see figures 7 and 8). This block can be sub-divided into three sub-units; at the base was a dark grey silty-sand (fine to medium) that contained abundant freshwater gastropod shells. The middle sub-unit was a mat-like deposit of stratified herbaceous plant macrofossils in a matrix of silty clay (L.O.I. 31%). The plant remains were identified as *Phragmites* spp. Above this reed peat was a sub-unit of 0.45 m of wood peat which was between 30% and 70% wood by volume and was generally sub-horizontally stratified. Most of the wood was sub-aerial tree parts which ranged from seeds and twigs to trunks over 3 m in length and 0.3 m in diameter. These remains were extremely well preserved and were identified as *Salix* spp. and *Alnus glutinosa*. There was also a vertically orientated wood component which penetrated the whole unit and 0.20 m into the underlying gravels. These macrofossils were also well preserved and all identified as *Alnus glutinosa* roots. The matrix of the top channel (wood peat) contained 66% organic matter the rest being silt and clay. Towards the top of the wood peat the inorganic content increased so the top 0.01 m had a L.O.I. of 33%.

Upwards, sedimentary unit 3 became a massive clayey silt that covered the floodplain to a depth of 2.0 m to 2.5 m. The roots in the sediment below abruptly terminated in the bottom 0.04 m of the clayey silt but showed no signs of disturbance. Five samples were dated from the matrix of organic-channel infill and from the penetrating roots. Fraction dating was used in an attempt to highlight any problems of contamination or mixed provenance (figure 7). The organic matrix, which was probably deposited by the infilling of a cutoff palaeochannel, seems to have started accumulating around 10200 years BP and continued until around 8000 years BP or a bit later. In its early stages it was covered by a reed bed which was

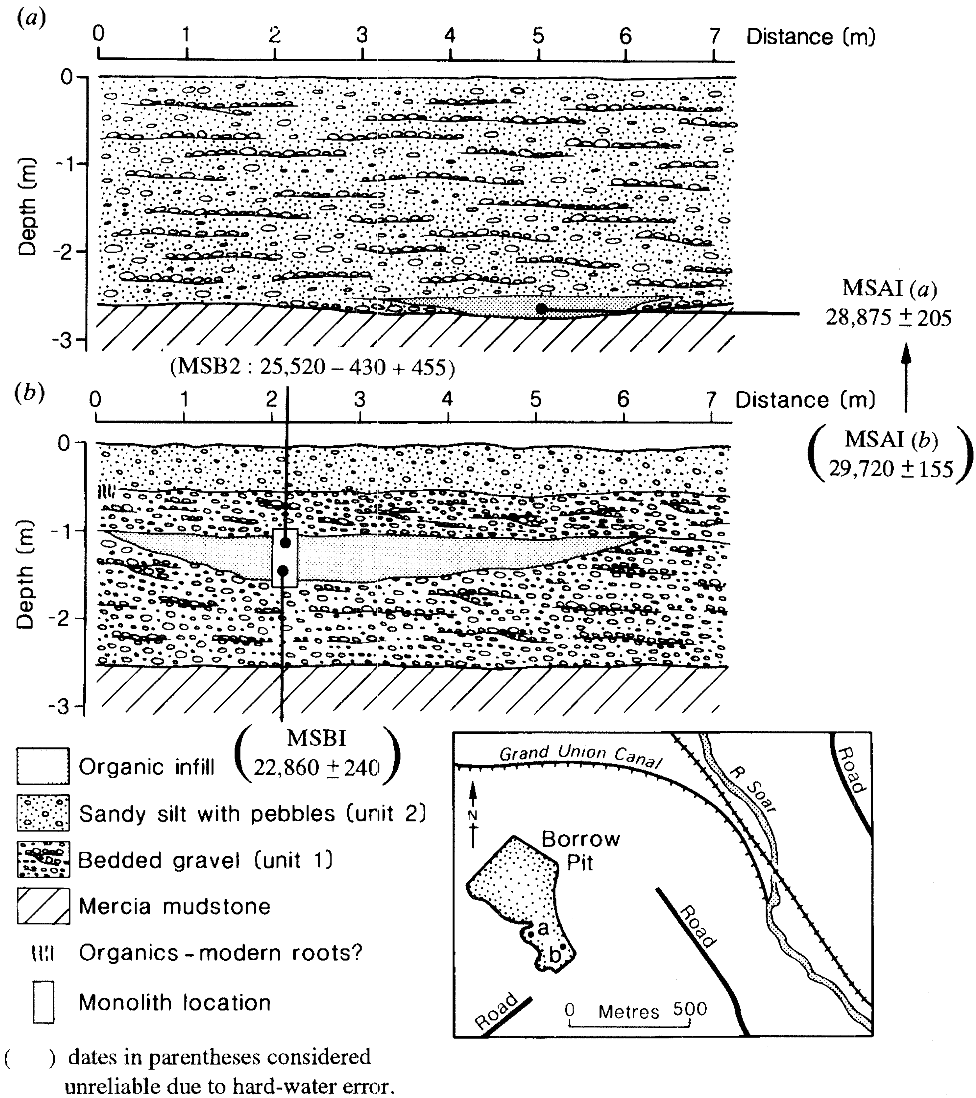

Figure 9. Stratigraphic section from Quorn, the borrow pit of the Mountsorrel Bypass (cross-section B).

replaced by alder woodland. This infill was subsequently penetrated by alder roots around 5200 years BP. The difference in the humic carbon and humin carbon fraction increases at the top of the block suggesting some contamination here. The lower dates are not statistically distinguishable at the one standard deviation level and the coarse plant material dated from CN03 sits in between the fraction dates. The site reveals a Lateglacial palaeochannel, which ran approximately parallel to the present river and silted up during the early Flandrian but was covered, as was the rest of the floodplain, by overbank clayey silts sometime between 8000 years BP and 5200 years BP. On this fine sediment alder woodland persisted sending roots down through the clayey silt into the underlying channel fill.

Table 2. *A list of the radiocarbon dates used in this study*

(Site codes: (BL) Birstall, (CN) Cossington, (MS) Quorn, (LH) Little Houghton, (WN) Wollaston, (DD) Ditchford, (HF) Higham Ferrers, (RS) Raunds.)

site no.	lab. no.	material	date (BP)	^{13}C value (‰)
Soar				
BL01	SRR-3502	wood (*Quercus*)	5160 ± 55	−27.2
BL02	SRR-3503	wood (*Alnus*)	4900 ± 55	−29.0
BL03	SRR-3504	wood (*Alnus*)	4820 ± 55	−29.6
BL04	SRR-3505	wood (*Prunus* sp.)	4285 ± 55	−27.3
BL05	SRR-3506	wood (*Corylus avellana*)	4015 ± 75	−28.2
BL06	SRR-3507	wood (*Fagus sylvatica*)	3530 ± 75	−27.6
BL07 (a)	SRR-3508	fine organic sediment	4180 ± 75	−28.7
BL07 (b)	SRR-3508	plant macrofossils	4340 ± 90	−29.3
CN01 (a)	SRR-3509	humic carbon	10230 ± 60	−30.0
CN01 (b)	SRR-3509	humin carbon	10150 ± 120	−30.3
CN02 (a)	SRR-3510	humic carbon	9620 ± 60	−29.4
CN02 (b)	SRR-3510	humin carbon	9410 ± 200	−30.0
CN03 (a)	SRR-3511	humic carbon	9685 ± 85	−29.8
CN03 (b)	SRR-3511	humin carbon	9250 ± 135	−30.0
CN03 (c)	SRR-3511	coarse plant debris	9370 ± 125	−30.6
CN04 (a)	SRR-3512	humic carbon	9685 ± 60	−28.8
CN04 (b)	SRR-3512	humin carbon	7295 ± 80	−28.0
CN05	SRR-3513	wood (*Alnus*)	5140 ± 55	−29.8
MSA1 (a)	SRR-4645	hand sorted twigs	28875 ± 205	−28.8
MSA1 (b)	SRR-4645	fine leaf frags./org. silt	29720 ± 155	−19.1
MSB1	SRR-4646	organic silt	22860 ± 240	−18.3
MSB2	SRR-4647	organic silt	25520 − 430 + 455	−16.0
Nene				
LH01	SRR-3611	wood (*Alnus glutinosa*)	3360 ± 65	−29.0
LH02 (a)	SRR-3612	fine particulate organic sed.	3440 ± 50	−28.5
LH02 (b)	SRR-3612	coarse organic detritus	3215 ± 50	−27.4
LH03	SRR-3613	herbaceous peat	4145 ± 60	−28.6
LH04	SRR-3565	wood (*Alnus*)	3565 ± 65	−27.2
WN01	SRR-3615	wood (*Quercus* sp.)	3400 ± 50	−25.1
WN02	SRR-3616	wood (*Quercus* sp.)	3300 ± 60	−26.3
WN03	SRR-3617	wood (*Quercus* sp.)	3275 ± 50	−27.5
WN04	SRR-3618	wood (*Corylus avellana*)	2270 ± 50	−28.5
WN05	SRR-3619	wood (*Prunus* sp.)	2410 ± 50	−28.6
WN06	SRR-3620	wood (*Prunus* sp.)	2395 ± 50	−26.4
WN07	SRR-3621	wood (*Prunus* sp.)	2420 ± 50	−28.4
WN08	SRR-3622	wood (*Prunus* sp.)	2540 ± 60	−29.4
WN09 (a)	SRR-3623	fine particulate organic sed.	9455 ± 55	−30.9
WN09 (b)	SRR-3623	coarse organic detritus	5540 ± 65	−27.6
WN10	SRR-3624	wood (*Prunus* sp.)	2485 ± 50	−27.6
WN11	SRR-3625	organic rich silt	9770 ± 50	−29.6
WN12 (a)	SRR-3626	fine particulate organic sed.	9250 ± 50	−29.8
WN12 (b)	SRR-3626	coarse organic detritus	8005 ± 75	−28.7
WN13	SRR-3627	wood (*Alnus*)	3605 ± 60	−28.5
WN14 (a)	SRR-3628	fine particulate organic sed.	2975 ± 65	−28.5
WN14 (b)	SRR-3628	coarse organic detritus	3070 ± 60	−28.9
WN15	SRR-3629	organic rich silt	2715 ± 65	−27.9
DD1	SRR-4644	silty peat	11220 ± 45	−28.5
DD2	SRR-4643	twigs and wood fragments	9485 ± 125	−28.1
DD3	SRR-4642	peat	10280 ± 45	−30.4

Table 2 (*cont.*)

HF01	SRR-3599	wood (*Fagus sylvatica*)	3270 ± 40	−26.8
HF02 (a)	SRR-3600	fine particulate organic sed.	3025 ± 50	−29.4
HF02 (b)	SRR-3600	coarse organic detritus	4130 ± 60	−27.9
HF03	SRR-3601	wood (*Sambucus nigra*)	2385 ± 50	−26.4
HF04	SRR-3602	wood (*Crateagus* sp.)	2520 ± 60	−27.7
HF05	SRR-3603	wood (*Salix* sp.)	3215 ± 45	−27.5
RS01	SRR-3604	organic rich silt	11395 ± 55	−28.6
RS02	SRR-3605	organic rich silt	9375 ± 40	−28.5
RS03	SRR-3606	wood (*Alnus*)	5195 ± 65	−28.9
RS04 (a)	SRR-3607	humic carbon	10870 ± 55	−29.3
RS04 (b)	SRR-3607	humin carbon	10970 ± 55	−29.7
RS05	SRR-3608	wood (*Alnus*)	3840 ± 50	−28.6
RS06	SRR-3609	monocotyledon fragments	2120 ± 50	−30.6
RS07	SRR-3610	organic rich silt	12420 ± 60	−29.7

(*c*) *Quorn* (*the Mountsorrel bypass*)

The borrow pit excavated for the Mountsorrel bypass (A6) near Quorn (figure 1) revealed an extensive section of the basal gravel (SK 556 181), which clearly showed the top of the Mercia mudstone was smooth and undisturbed, and bore no signs of former channelling by the river. The gravels themselves, normally about 2 m thick, were of a medium to fine grade with few clasts exceeding 0.08 m in length. Individual gravel beds often persisted over a distance of 20 m or more, and there were no unequivocal periglacial structures disturbing the relatively even bedding. There were occasional thin beds of cross-bedded sand and locally, towards the top, the sand appeared to become more persistent, although never to the extent that would justify its recognition as a separate unit. The overlying clayey silt was thin and impersistent. Over large areas the gravel passed up directly into the modern soil, and even where the silt was present it nowhere exceeded a thickness of some 0.40 m.

At two locations channel-like features filled with organic sediments were observed within the gravel (figure 9). The centre of the first of these lay at the contact with the Mercia mudstone, but its margins were underlain by the gravel. The second lay close to the top of the gravel and the organic materials appeared to have accumulated in a channel abandoned late in the main gravel aggradation. Pollen from the lower organic channel fill within the gravels was dominated by Gramineae, Cyperaceae, *Helianthemum*, Caryophyllaceae, *Artemisia*, and Umbelliferae. No tree pollen was recorded at all (standard TLP sum of 500 grains). This fact along with the very high *Helianthemum* percentage (18% TLP excluding Cyperaceae) clearly indicates a Glacial or early Lateglacial environment. Three samples were submitted for ^{14}C analysis. The date from the base was fraction-dated, as low C values indicated hard-water error in the fine leaf fraction. However, the hand sorted twigs gave a normal isotopic signature and a date of 28875 ± 205 years BP (SRR-4645). The upper channel organics unfortunately did not contain countable pollen and the ^{14}C dates were unreliable (table 2) due to their heavy isotopic signatures.

These dates, however, do reduce the window for the deposition of the basal sand and gravel in the Soar valley from between 37000 years BP and 7500 years BP to between 28000 years BP and 10000 years BP.

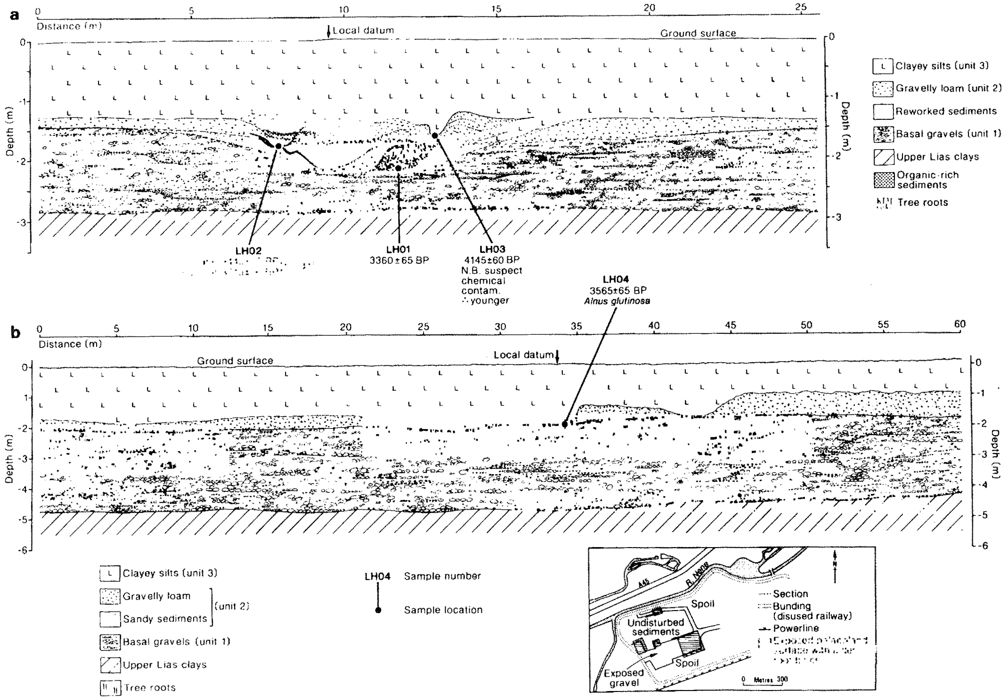

Figure 10. Stratigraphic sections from Little Houghton: (*a*) south, (*b*) north.

6. The Nene sites

(*a*) *Little Houghton*

The Little Houghton site (figure 1) is located 2 km downstream of Northampton (SP 782 598) where gravel extraction revealed a number of exposures, two sections of which are reproduced here and referred to as the south and north sections. The full south section (see figure 10*a*) which was 300 m long had, in the central 25 m, an accumulation of sandy sediments with a variety of sedimentary structures which also contained organic remains. The stratigraphy of the rest of the exposure consisted of three units. At the base was a 1.5 m thick accumulation of planar bedded sandy gravels, overlain by unit 2, a 0.15 m thick bed of gravelly loam which thickened towards the central part of the exposure. The loam is a mixture of gravel, sand, silt and clay as shown in figure 3*a*. The superficial unit 3 was a 1.25 m thick clayey-silt. It had no sedimentological structure but did contain a soil profile with strong ped development down to 0.40 m. The hollow in the basal gravels which was about 10 m wide and 0.75 m deep had organic remains preserved at three locations within it. One consisted of waterlogged *in situ* roots of *Alnus glutinosa* which gave a radiocarbon date of 3360 ± 65 years BP (SRR-3611). From about 7 m along the section a small erosional hollow contained small unidentified macrofossils in an organic matrix which was a poorly stratified and relatively unhumified peat (L.O.I. of 31 %). A bulk sample gave a date of 3440 ± 50 years BP (SRR-3612a) for the fine fraction and 3215 ± 50 years BP (SRR-3612b) for the coarse detritus: these dates are not statistically different at the single standard deviation level. The third organic sample was of small plant stems probably *Juncus* spp. which seemed to be *in situ* and to have been growing in a small pit or hollow. A sample of these gave a date of

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

4145 ± 60 years BP (SRR-3613). However, during the benzene preparation stage chemical contamination of the sample was noticed (D. D. Harkness, personal communication) and this along with its stratigraphic position suggests that the date may be some 800 years too old. The origin of this hollow is not known but it did not display the expected characteristics of a palaeochannel in terms of shape and an armoured layer or lag deposit, nor could it be traced in successive section faces. It did, however, show considerable evidence of having been disturbed by tree roots, with probable tree-throw producing the pits and the interdigitation of relatively coarse mound deposits with fine overbank silts. Although artifacts were not found the possibility of human activity cannot be entirely dismissed.

The north section (see figure 10*b*) was 60 m long and exposed 4.2 m to 4.8 m thickness of gravel, gravelly loam and silt similar to the south section. The middle unit of gravelly loam ranged in thickness from zero in the middle of the section to 0.9 m (see figure 3*a* for the grain size distribution). Over most of the section it was structureless but in several places it displayed an indistinct coarse horizontal bedding and occasionally lenses of sand with clear bedding and discontinuous silty drapes. The boundary between this unit and the basal gravels was abrupt but slightly undulating. The upper structureless clayey silt varied in thickness in sympathy with the uneven subsurface topography of the gravelly loam. *In situ* tree roots penetrated the upper 0.20 m to 0.25 m of the basal gravels. They were usually found in groups of two to five together and separated by 2 m to 5 m. All were identified as *Alnus glutinosa* and could be traced in plan as the overburden was removed. All the roots terminated 0.02 m to 0.04 m into the gravelly loam and inspection of the ends revealed they had decayed rather than been eroded or cut. A date was obtained from one of the roots of 3565 ± 65 years BP (SRR-3614).

Considering the data from all the sections, but excluding the anomalous date, there is a variation in the ages of the *Alnus glutinosa* roots of 205 years which, although the cause of death is not known, does not support a catastrophic explanation. The dating also shows that deposition of the gravelly-loam commenced before 3400 years BP and that it was penetrated by roots growing from a landsurface that must have existed at about the same level as its upper boundary. The most obvious interpretation of the sections is that before 3400 years BP a coarse floodplain soil developed on the gravels, along with some overbank deposition and above a groundwater table within the basal gravels. Preservation of the roots was probably the result of a rise in groundwater levels associated with the deposition of the superficial clayey silt.

(*b*) *Wollaston*

The site at Wollaston (figure 1), which lies 14 km downstream of Northampton (SP 891 641) revealed a similar stratigraphy to the sites at Little Houghton, with one section having three comparable sedimentary units (figure 11). At the base was 2.5 m to 3.0 m of planar bedded sub-rounded to well-rounded gravels, which thickened towards the middle of the floodplain. The middle unit consisted a poorly sorted gravelly loam which graded laterally into a bedded sand (unit 2, figure 9). The boundary between these two units was level but gradual. Above the gravelly loam/sand was the structureless clayey silt unit (3) which was variable in thickness depending upon the loam and gravel topography. The section (figure 11) exposed a complex stratigraphy. This complexity consisted of a 75 m wide depression in the basal gravels infilled with gravels, bedded sands and organic rich silts. Two

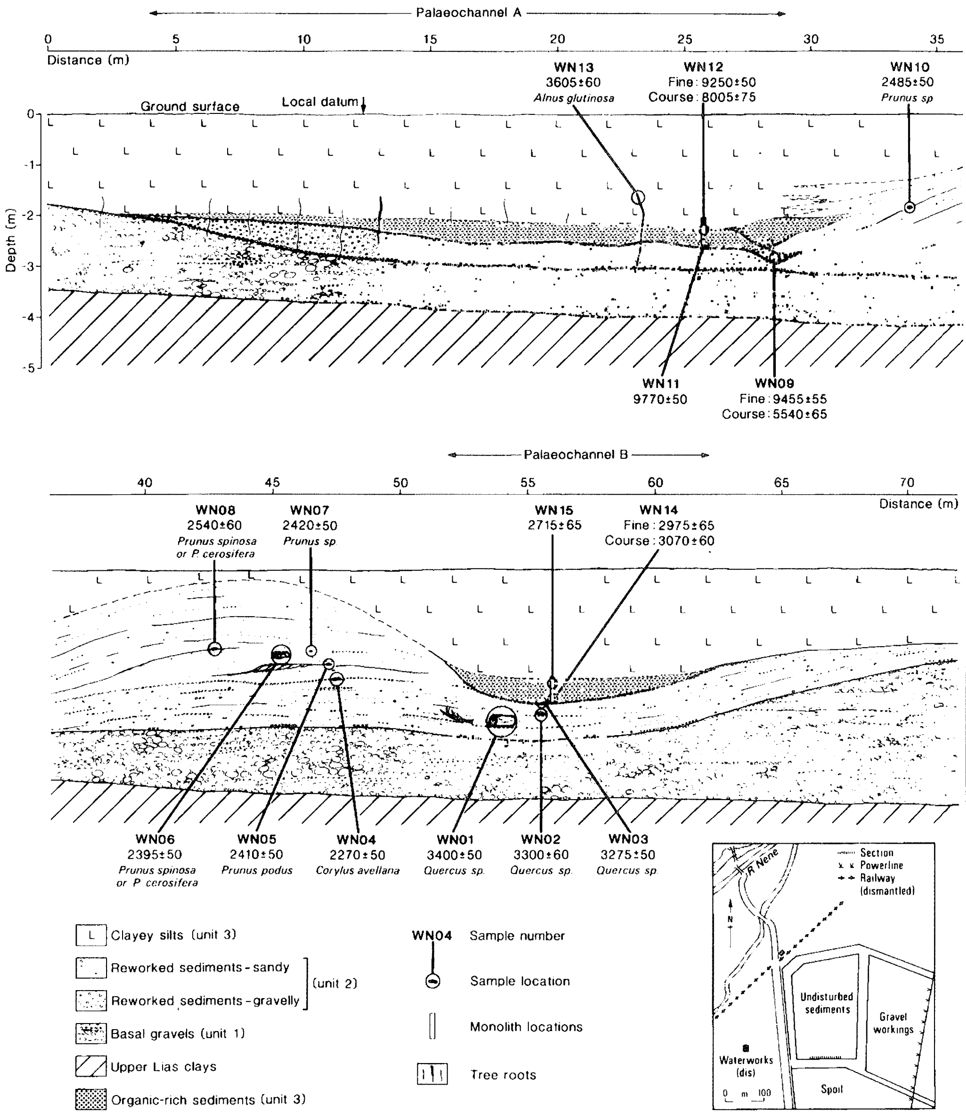

Figure 11. Stratigraphic sections from Wollaston (*a*) east (*b*) west.

palaeochannel scars rested upon the gravels and sands and were separated by a sand mound. Palaeochannel scar A rested on a sandy gravel with an armoured bed. It was infilled by 0.10 m to 0.35 m of organic silt and a few isolated gravel clasts. Both this infill and the underlying sandy gravel were penetrated by roots of *Alnus glutinosa*. In the middle of the section the palaeochannel was eroded and disturbed and it contained two small depressions. The lower depression seems, on the basis of the radiocarbon dating of organic material within it and also the continuation of the armoured layer into it, to be contemporaneous with the rest of the palaeochannel scar. It contained organics of mixed provenance probably resulting from monocotyledonous root penetration. The upper depression was gravel filled, cut into the palaeochannel fill and, along with the clayey silt above, interdigitated with the sandy mound. Palaeochannel scar B was excavated into the gravelly sand which had many gravel stringers, contained a large red deer (*Cervus elaphus*) antler with unbroken tines and a large trunk of *Quercus* spp. The antler had a cut-mark around its base, presumably made in order to snap-off a section, very much like the Iron Age worked antler recorded from Collfryn hillslope enclosure, Powys (Britnell *et al*. 1989). The mound (unit 2) consisted of cross-bedded and planar-bedded sands inter-bedded with discontinuous silty clay drapes and occasional gravel-stringers. This unit graded up into the sand mound that separated the two palaeochannel scars. However, there were several important vertical changes, and an almost continuous gravel stringer separated the lower sands which contained gravel and wood of *Quercus* spp. only, from a convexly-bedded upper sand which showed some micro-epsilon type bedding and contained wood of dry-land trees only (*Prunus* spp.). The interdigitation of these sands with the clayey silt, their convex bedding, height, and macrofossil content all suggest they accumulated, at least in part, either subaerially as a levee, or as a flood-mound rather than a mid-channel bar. Wind action is thought unlikely due to the bedding and unfavourable vegetation. The palaeochannel to the east (palaeochannel B) also had an armoured bed, was infilled with organic silty clay and was covered by the superficial massive clayey-silt (unit 3).

The site history is obviously complicated, but the following scenario can be suggested. A broad scar was cut into the basal gravel unit before 9800 years BP. Fluvial gravels were deposited in the scar and a second channel bed formed. This was abandoned by 9770 years BP. This channel started to infill but the eastern end was eroded by a channel which deposited the gravelly-sand around 3300 years BP. A new channel bed seems to have existed between 3275 years BP and 3070 years BP when the channel started to infill. It is quite possible this was just a flood-channel. Bank deposition of sand seems to have continued from 2540 years BP to 2270 years BP during which time the basal metre of the clayey-silt superficial unit was deposited. It seems that the deposition of the mound and the fine floodplain surface kept pace until after this date when overbank fine deposition overtopped the mound.

(*c*) *Ditchford*

The site at Ditchford (SP 929 680, figure 1) lies 6 km downstream of Wollaston and displayed a less complex but typical stratigraphy. The sub-gravel surface of the Upper Lias clay was highly undulating and cut into by a small (7 m wide by 0.7 m deep) channel with an organic infill (see figure 12). This channel post-dated some gravel deposition as it was cut down through 0.30 m of the basal gravels. Another extremely small palaeochannel remnant was found in the middle of the gravel unit (1) and at the top a large palaeochannel, 40 m wide was cut 1.3 m into the top of the gravels. It was filled with organic silty clay, contained abundant wood (*Quercus* spp.) and was penetrated by *in situ* roots of *Alnus glutinosa*. Away from the edge of the palaeochannel a sandy gravel (unit 2) thickened to a maximum of 0.3 m. It showed no sedimentary bedding or structures. Covering the whole exposure was 0.9 m to 1.75 m of brown clayey-silt (unit 3). Pollen from the basal channel is dominated by Cyperaceae, Gramineae, *Pinus* and *Betula*. Excluding Cyperaceae, which is over-represented (80 % TLP including Cyperaceae), the arboreal pollen is 41 %, suggesting cool-temperate forest conditions. A sample of silty peat from the base of the channel gave a ^{14}C date of 11 220 ± 45 BP (SRR-4644) confirming a Lateglacial interstadial date from the channel which was cut through a thin layer of unit 1 into bedrock. The erosion of the channel and deposition of the bulk of the gravels in the section

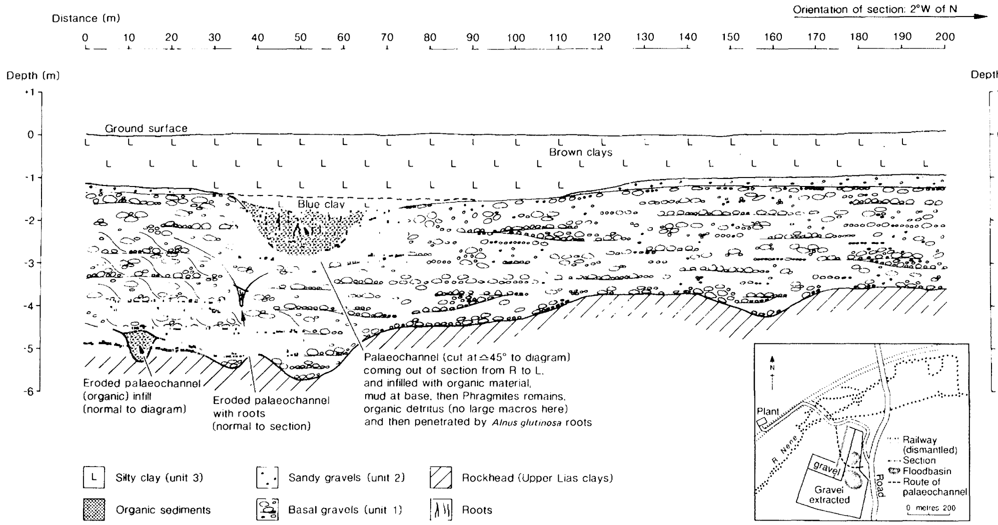

Figure 12. Stratigraphic section from Ditchford.

occurred during the Lateglacial stadial or the Flandrian. The pollen from the base of the upper channel was dominated by Cyperaceae, Gramineae, *Rubiaceae* and *Filipendula*. The only trees are *Betula* and *Pinus*, and the arboreal pollen is low at 10% TLP (excluding Cyperaceae) suggesting a possible Lateglacial date which is confirmed by a ^{14}C date of $10\,280 \pm 45$ BP (SRR-4642). From macrofossils it is clear that the upper part of this infill was penetrated by roots of *Alnus glutinosa* making the site very similar to the channel infill at Cossington and Wollaston, i.e. a very early Flandrian accumulation of herbaceous peat with superimposed mid-Flandrian woody macrofossils. Twigs from a very small channel in the middle of the gravel body seem to have given an anomalous date by about 1000 years, the reasons for this are unknown, but it is unlikely that the upper date is significantly too early. The dates indicate that erosion of some older gravels and deposition of nearly 4 m of gravel took place in under 1000 years during the Lateglacial stadial.

(*d*) *Higham Ferrers*

The site is located 2 km downstream of Ditchford and 4.3 km downstream of the confluence of the river Ise with the Nene (figure 1). The floodplain at this location is about 0.85 km wide and has a gradient of 0.000 66. Gravel extraction exposed large sections on the south side of the river. The stratigraphy typically consisted of a basal sandy gravel unit, a gravelly loam and a superficial clayey silt unit (figure 13). The gravel unit (1) was 3.0 m to 3.3 m thick, planar-bedded with well-rounded clasts. The gravelly loam (unit 2) was 0.25 m to 0.5 m thick with a poorly defined lower boundary grading into the basal gravels. In the main section the gravelly loam was thicker and extremely complex with an abrupt but uneven boundary with the overlying clayey silt that decreased in height as the loam thinned to the east where there was a marked erosional surface, with an armoured layer indicating a palaeochannel scar. This unit (2) consisted of lenticular sub-units of gravels, gravelly loam, sands, silt and clay. Cross-bedding is evident as is the interdigitation of fine sediment with sand lenses. Some of the sub-units display a fining upward. The infilling of the palaeochannel scar (at 0–10 m) seems to pre-date, or be con-

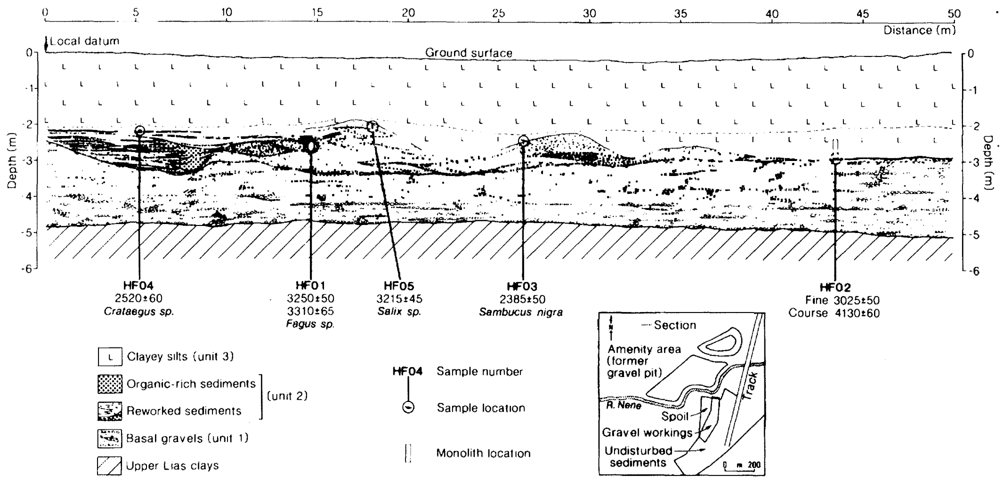

Figure 13. Stratigraphic cross-section from Higham Ferrers.

temporaneous with, the deposition of at least the upper half of unit 2 commencing somewhere between 4000 years BP and 3000 years BP. It contained a large trunk of *Fagus sylvatica*, which still had bark and showed no signs of sub-aerial decay. It gave a date of 3270 ± 40 years BP (SRR-3599). Above this a wooden stake (probably *Salix* spp.), sharpened and with the lower bark missing was stuck into the gravels and dated to 3215 ± 45 years BP (SRR-3603). This may have been a bank-side stake related to a fish weir or perhaps part of a hurdle fence. Two other samples were recovered from the boundary between the unit 2 and the overlying clayey silt (unit 3). The first was a rather worn and barkless fragment of *Crataegus* spp. which produced a date of 2520 ± 60 years BP (SRR-3602) and the second was a worn fragment of *Sambucus nigra* which was dated at 2385 ± 50 years BP (SRR-3601).

The stratigraphy suggests a Flandrian erosional surface cut across early sand and gravel. There was both within-channel and marginal channel sedimentation in a palaeochannel around 3300 years BP with burial of both the sediment and the palaeochannel infill by overbank clay and silt around 2400 years BP.

(*e*) *Raunds*

The sites in the Raunds (figure 1) area lie some 24 km downstream of Northampton in a reach, which is about 4 km long and rests upon Upper Lias clay at a depth of 4.0 m to 4.5 m. Overlying the bedrock were the basal gravels which were planar or massively bedded, and varied between 1.75 m to 2.5 m thinning out under palaeochannels. Above the gravels unit 2 varies laterally from matrix supported gravels to interbedded sand and silt with clay lamina. It includes poorly sorted loamy sub-units of variable thickness ranging from 0 m to 0.45 m. A clayey silt covers the floodplain to a depth of between 1.5 m and 1.75 m.

The upstream site (SP 965 720) revealed a long section across the floodplain (see figure 14). This showed a gently domed gravel surface with a palaeochannel containing an organic fill at the south end and a less clear palaeochannel scar with an inorganic infill at the north end. A sample from the base of the organic silt, above the armoured layer, gave a date of $11\,395 \pm 55$ years BP (SRR-3604) and a date from the uppermost preserved organics in the fill gave a date of 9375 ± 40 years BP (SRR-3605). The gravel surface contained numerous root mats of *Alnus glutinosa* all *in situ*

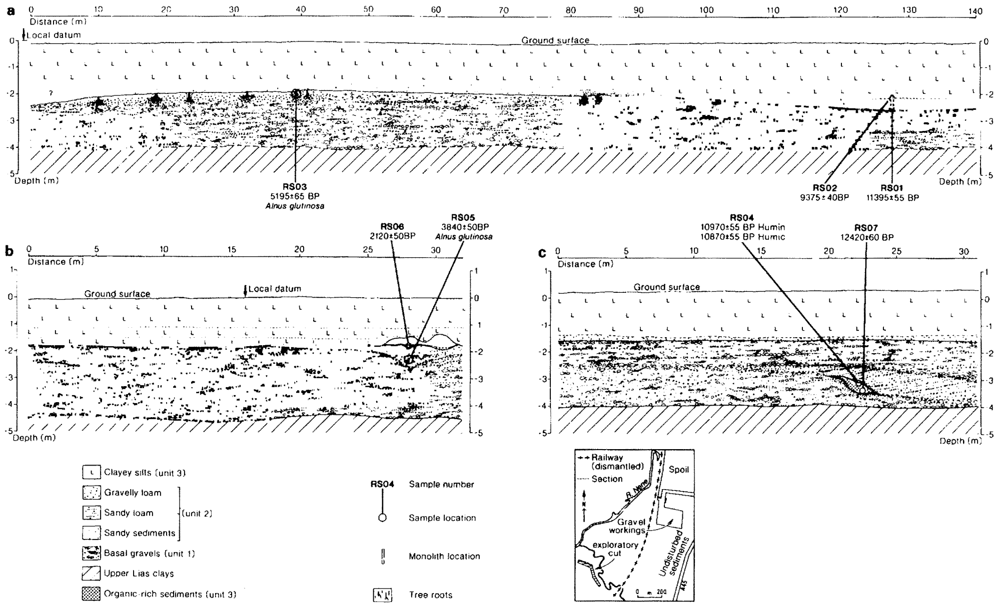

Figure 14. Stratigraphic sections from Raunds: (*a*) upstream site, (*b*) and (*c*) downstream sites.

and generally surrounded by a finer, reworked, saucer-shaped zone of gravels. A sample from one of the roots gave a date of 5195 ± 65 years BP (SRR-3606). These roots also extended a few centimetres up into a poorly stratified sandy loam which covered the gravel surface in between the palaeochannel scars and was thickest towards the centre. Because this lay within 2 m of the present floodplain surface and was disturbed by tree roots it was interpreted as a palaeolandsurface dating from at least 5000 BP.

At the downstream site (SP 972 725) two exposures were logged in detail (see figure 14*b*, *c*). One showed a peat lens within the gravel unit (figure 14*c*). The lens was interpreted as a small channel fill that had been partly destroyed by erosion. This was inferred from the armouring of the lower peat gravel contact, the occurrence of some horizontal stratification within the peat, and an upper erosional boundary. The gravels deposited above the upper peat contact, were slightly finer and clearly cross-bedded unlike the basal gravels or those stratigraphically below the peat which showed wavy-planar bedding. A sample from the base of the peat gave a date of 12420 ± 60 years BP (SRR-3610) and two dates from the top gave statistically indistinguishable dates of 10870 ± 55 years BP (SRR-3607a) and 10970 ± 55 years BP (SRR-3607b) from the humic and humin fractions respectively. The upper metre of gravel had less wavy bedding and was better sorted than the rest. Above this was a thin gravelly loam (unit 2) covered by just over a metre of massive clayey silt (unit 3). One face (see figure 14*b*) showed some disturbance of the gravel–fines contact. This was a slightly conical pit filled with silty sand with some fine gravel and with *in situ* roots of *Alnus glutinosa*. Above this was a very thin silty layer which contained remains of *Juncus* spp. dated at 2120 ± 50 years BP (SRR-3609). Sitting on top of this was a curious asymmetrical sandy mound, covered, as is the rest of the exposure, by the superficial clayey silt. Although the exact sedimentary sequence is difficult to ascertain, the most likely explanation of the feature is a tree-throw pit and mound developed on an area previously disturbed by tree growth. The mound closely

Table 3. *Dates of other palaeochannel segments at Raunds*

site code	max. depth below ground level/m	width/depth	*post-terminum* date of channel abandonment (years BP)	lab. no.
WCT	3.33	17	5500–4000[a]	—
E1	2.19	19	9370 ± 170	Har-9243
C1	1.87	—	4300 ± 100	Har-9241
D1	1.80	—	1970 ± 80	Har-9242
RS01[b]	2.35	—	11395 ± 55	SRR-3604
RS07	3.45	—	12420 ± 60	SRR-3610

[a] Estimated on the basis of pollen, macrofossil analysis and archaeology.
[b] From this study, included for comparison.

resembles contemporary tree throw mounds and their stratigraphy (Schaetzl *et al.* 1989; Johnson 1990; Schaetzl 1990). The sandy mound may also be, at least in part, due to increased overbank deposition of the flood saltation load probably caused by localized floodplain roughness elements (Brown 1983).

Archaeological excavations in the same reach have discovered several other palaeochannels segments, the dates of which are given in table 3 (Brown 1994). The existence of several Prehistoric archaeological sites on and at the edge of the floodplain shows that the higher sub-alluvial gravel islands have not been eroded by channel flow since the Neolithic Age (*ca.* 5000 BP) or even earlier. At one site a Neolithic landsurface revealed 35 tree-throw pits (McPhail & Goldberg 1990) and its position close to Raunds Upstream site indicates that little or no channel migration has occurred since that time between the present east and west channels of the Nene.

7. The biological evidence and floodplain palaeoecology

All of the sites investigated contained wood which was identified using thin sectioning and standard identification manuals for sub-aerial wood and roots (Schweingruber 1978; Cutler *et al.* 1987). Particular attention was paid to the condition of the wood. The most common species were, in order: *Alnus glutinosa* which was found at all sites except one, *Quercus* spp.; *Corylus avellana*; *Salix* spp.; *Prunus spinosa*; *Crataegus monogyna*; and *Fagus sylvatica*. Acorns and hazel nuts were recorded from several sites often showing evidence of rodent attack. In addition to the woody macrofossils remains of *Phragmites* and *Juncus* spp. were recovered. The only species which were definitely recorded *in situ* were *Alnus glutinosa* and *Phragmites* spp. The general distribution of species, in terms of frequency at each site, is closely approximated by the dated wood which allows the dated wood to be used as a proxy of changes in species composition over time (figure 15). Using all the available radiocarbon dates it can be seen that there are two clusters of dates, one in the Lateglacial and early Postglacial and a second peak between 6000 years BP and 2000 years BP. Also noticeable is the almost total dominance of the early peak by *in-situ* mixed organics (i.e. *in situ* but of several species or organic sediment types) and the greater importance of allochthonous organics in the later peak. If we look at the distribution on a species basis it is clear that the late peak indicates a change from *Alnus glutinosa* and *Quercus* spp. to a more mixed assemblage in the third millennium with complete replacement by the dry-land component (*Prunus*, *Sambucus* and *Corylus*) by the second millennium BP. This indicates the replacement of the natural

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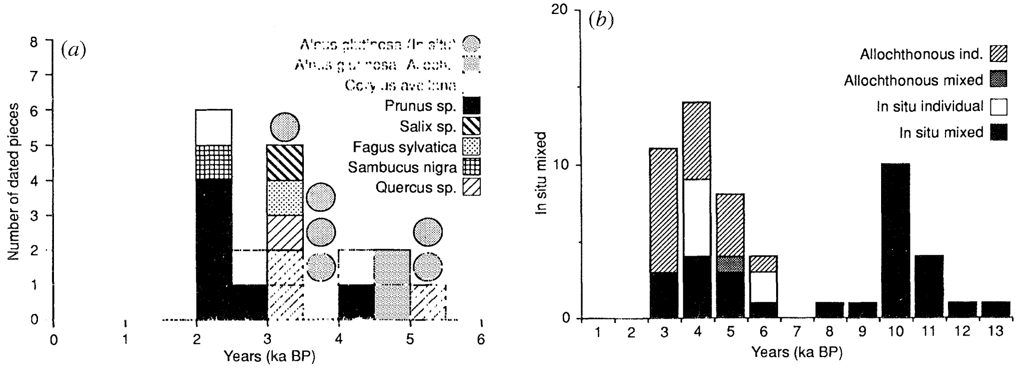

Figure 15. (*a*) Combined dated wood histogram indicating species, and (*b*) radiocarbon sample type by sample age.

wet woodland by dry-land trees typical of managed floodplains with hedges and hedge-like communities. Where the *Alnus* is *in situ* as roots it is likely that it is the last generation that was preserved as the subsoil became waterlogged. No subsequent disturbance by *Alnus* roots even though most of the sites are within rooting depth, implies that the preserved roots were from the last generation of *Alnus* trees on the sites. This gives the latest dates for deforestation which vary from 5140 ± 55 years BP (SRR-3513) at Cossington and 5195 ± 65 years BP (SRR-3608) at the Upstream site at Raunds, to 3360 ± 65 years BP (SRR-3611) at Little Houghton. This agrees well with pollen evidence that the deforestation of the Nene floodplain in the Raunds reach as a whole was relatively early (Brown unpublished work), but also, as we might expect, there was considerable variation even within a floodplain reach. The managed floodplain persisted to the present, and Robinson (1992) has shown using non-woody macrofossils from palaeochannels that the floodplain was under hay meadow management before becoming alluvial flood meadow. *Fagus sylvatica* wood from Higham Ferrers was double dated at 3250 ± 50 years BP (SRR-3599a) and 3310 ± 65 years BP (SRR-3599b) the dates being indistinguishable at the one standard deviation level. This relatively early date confirms the presence of beech on either free-draining parts of the floodplain, such as gravel islands, or at the floodplain edge, during the Middle Bronze Age. *Fagus* wood has been recovered from the Somerset levels trackways as early as 4500 years BP, but its expansion to form a significant component of woodlands in late Prehistory has been linked to anthropogenic disturbance (Godwin 1975). Interestingly it is hardly represented at all in pollen diagrams from this area. In general there is relatively little evidence of the vegetation on the slopes and terrace remnants, and in particular there is no evidence of *Tilia* wood (although fruit have been found) and little pollen, which contrasts to many other sites in the Midlands (Greig 1982; Brown 1988); these include pollen from peat exposed at the base of the floodplain silt clay at the sewage treatment plant near Cossington in 1960 (F. Oldfield, personal communication).

8. Floodplain evolution

(*a*) *The Devensian-Lateglacial*

The dating of the lower gravels reveals that the basal gravels in the Soar valley started to be deposited during the Glacial Maximum/Pleniglacial and may well have ceased deposition by the Lateglacial, although further research is required to prove this. By contrast the basal gravels of the Nene were, at least at Ditchford, deposited entirely during the Younger Dryas. The reasons for this difference are not clear, but it does reinforce the idea that Lateglacial fluvial response to climate change was rather variable depending perhaps upon local factors such as basin size, altitude, aspect and geology. These factors may have tipped basins over fluvial thresholds, so some basins, or even reaches, may have removed all Pleniglacial deposits and aggraded (e.g. Nene) whereas others may have seen far less Lateglacial erosion and deposition, preserving the older record (e.g. Soar).

Three sites provided evidence of Lateglacial conditions, these being, Raunds, Ditchford and Mountsorrel. The peat channel infill at Raunds downstream site was deposited during the Windermere Interstadial judging from the fact that the lower gravels pre-date 12400 years BP and deposition of the upper epsilon cross-bedded gravels post-date 10900 years BP. Ditchford suggests a similar sequence of events. The Raunds and Ditchford sites suggest abandonment of a Devensian pollen stage I channel associated with braided river gravel deposits (observed at Raunds), its infilling during pollen stage II (Windermere Interstadial) with one phase of partial erosion during pollen stage III (Loch Lomond re-advance). Because some palaeo-channels, such as the one at Raunds, have not been destroyed by erosion despite their relative shallowness, it seems the renewed fluvial activity during the Loch Lomond Stadial only effected parts of the floodplain. This sequence is remarkably similar to that described by Rose *et al.* (1980) at Sproughton on the river Gipping and it adds weight to the common assumption that in the lowlands, braided channels were replaced by meandering channels and a lack of gravel deposition during the Lateglacial Interstadial. The Loch Lomond re-advance may have marked a return to braided river pattern in some places.

At Mountsorrel data from the lower channel suggests gravel deposition began earlier and was not as eroded and reworked during the Lateglacial as was the case in the Nene valley.

These three sites show that while the basal gravel unit in both the Nene and Soar valleys was deposited during the Glacial Maxima, significant differences in erosion/deposition occurred during the Lateglacial. In the Nene channel abandonment and organic infilling occurred during the Lateglacial Interstadial. During the Loch Lomond Stadial there was renewed erosion and braided-river gravel deposition. However, in both valleys there was widespread abandonment of channels with relatively little gravel reworking by meander migration during the early Holocene. The model of floodplain evolution put forward below seeks to use these Lateglacial conditions as the starting point for Holocene channel and floodplain development. The Lateglacial sequence described partly corroborates Castleden's (1980) Nene model as there does seem to have been significant incision between 11000 years BP and 10000 years BP, but with minor incision before 12400 years BP and lateral erosion with gravel deposition during the Lateglacial Stadial (*ca.* 11000–10000 years BP). However, this is not the case with the Soar. There is no obvious explanation of the apparently different behaviour of the two catchments during the Lateglacial. However, it may be significant that the alignment of the Nene valley is generally west–east and so it may have had a greater snow-melt runoff from south facing slopes than the south–north trending Soar valley.

(b) The Flandrian

From the end of the Lateglacial Stadial until *ca.* 9500 years BP a series of palaeochannels were abandoned and started to fill with organic sediments. This occurred in both valleys (at Raunds, Wollaston and Cossington). There is, however, very little evidence, such as point-bar sediments, of lateral channel deposition during this period. The channels had high W/D ratios but varied in size, suggesting they were not caused by channel migration as successive chute or neck-cutoffs but represent the abandonment of secondary channels in a multi-channel system. There is no evidence during this period of recognizable landsurfaces, due to a lack of soil development, subsequent superimposition of later pedogenic disturbance and possibly some erosion. There is a marked gap in the occurrence of dateable organics, as at no site was there a channel fill which began accumulating between 9400 years BP and 4200 years BP. There is also no evidence of extensive lateral reworking of sediments during this period. There is, however, evidence of stable landsurfaces supporting trees (dense alder wood and alder/oak wood on dryer areas). It is unlikely that this is an artefact of preservation and more likely that it is the result of channel stability and a lack of channel abandonment. The evidence from Cossington and Raunds indicates that dense alder wood covered even the Late Devensian palaeochannels by 5200 years BP. The preservation of the last generation of alder roots at these sites may also indicate a rise in groundwater levels before alluviation. This is in agreement with Robinson's (1992) observations at Raunds and other sites in the southern Midlands.

After 5200 years BP major channel change takes place with seven of the eight sites showing either channel abandonment and/or lateral deposition. However, this channel migration was confined to small areas of the floodplain, and the majority of it was not reworked. In between channels there existed landsurfaces which show pedogenesis and biological mixing including the mixing of the underlying sand and gravel with silt and clay. This is the process which created the rather variable clay–silt–sand–gravel (gravelly loam) unit sandwiched between the gravels and the overbank silt–clay unit. It is therefore primarily a pedological rather than sedimentological unit. The incorporation of the fines into this soil, and the lack of organics in the lower silt-clay, marks the onset of the overbank deposition, which covered the whole floodplain including all the palaeochannels, but is difficult to date. The dates for roots which penetrate the boundary clearly indicate it is diachronous and ranges from as early as 3600 years BP to 2700 years BP even within a distance of 30 m (Wollaston). Although not adequately dated there is evidence that the majority of the unit is Medieval and Saxon (as it covers archaeological sites of that age (Foard & Pearson 1985)) and may be associated with the formation of the alluvial flood meadows. In both floodplains since *ca.* 2000 years BP there has been relatively little channel change and no major channels have been abandoned. What has happened is that the floodplain and channel banks have aggraded while channel beds have remained at approximately the same elevation although they have been covered by a blanket of fine sediment. This development, termed stable-bed aggrading-banks (SBAB), may have led to a natural decrease in the frequency of flooding since the Medieval period as would be predicted by the Leopold & Wolman (1957) overbank model.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

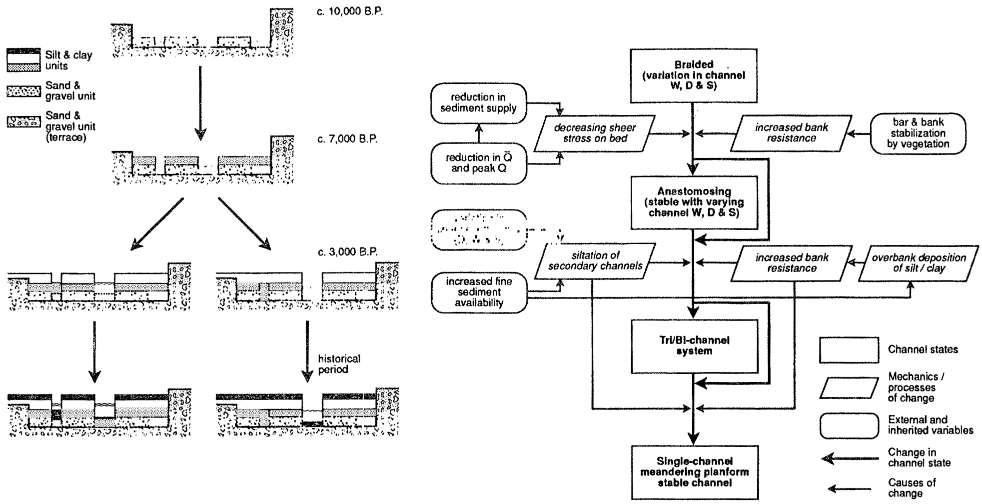

Figure 16. Anastomosing stable-bed and non-stable-bed aggrading-banks evolutionary model of the Flandrian floodplains of the Nene and Soar valleys. Upon this development is superimposed both lateral channel migration and any effects of climatic or catchment hydrological change. This development has included a general decrease in W/D ratio, an increase in the suspended sediment load relative to bedload and an aggradation of the floodplain surface. Not included are the effects of any artificial channel works which in recent times have tended to reduce channel numbers and decrease sinuosity.

(c) *Stratigraphic criteria for the model*

From the stratigraphic evidence at the eight sites, the following sedimentary events, characteristics and relations must be incorporated into any explanatory model of the floodplain evolution of the two valleys.

1. The occurrence and preservation of palaeochannels abandoned suddenly and their subsequent infilling.
2. The occurrence of palaeochannels of very different ages with very similar bed elevations.
3. The non-disturbance of large areas of the gravel surface underlying the clay-silt unit.
4. The occurrence of stable landsurfaces across much of the floodplain during the middle and later Flandrian.
5. The lack of evidence of any past floodplain surface higher than the present.
6. The initial conditions of a multiple channel pattern.

(d) *The floodplain evolution model*

To accommodate these criteria, an evolutionary model is proposed (figure 16) which is based upon a changing multi-channel system, channel siltation and floodplain aggradation. As depicted in figure 16 there is more than one pathway and different reaches or catchments may have reached different stages at different times. In the model Flandrian floodplain evolution commenced with stabilization of the bars of a braided channel system, a reduction in flow and stream power and a concentration of flow into a smaller number of channels. While the deeper thalwegs of the braided system retained flow, the higher channels may have been isolated and left with standing water, which provided ideal conditions for hydroseral succession. This explains the first generation of abandoned channels. Considering the extremely rapid rate of climatic change at the end of the Lateglacial (*ca*. 50 years, Dansgaard *et al*. 1989) it is not surprising that floodplain channels and stratigraphy could not adjust, but instead left much of the floodplain in a relict state. There was very little aggradation or channel erosion, probably due to both a reduction in mean annual and peak discharges, and there was little silt and clay sedimentation largely because of low suspended sediment loads. It is also likely that during the later Pre-Boreal and Boreal river discharges were reduced by effluent losses to the underlying gravels. Conversely there is evidence from the preservation of organics around 5000 years BP of a rise in floodplain groundwater levels before the abandonment of the late Flandrian palaeochannels and accelerated overbank deposition. Between 4200 years BP and 2000 years BP all the remaining channel abandonments occurred. This is accommodated in the model by assuming more than one channel persisted into the middle/late Flandrian (i.e. an anastomosing pattern *sensu* Smith & Smith (1980)) when the second phase of channel siltation occurred owing to increased suspended sediment loads. This is preferred over the alternative explanation that this represents a period of frequent avulsion, for two reasons. First, there is no evidence of crevasse-splay deposits that might be associated with a period of frequent failure of levees and avulsion, and second, the historical pattern of both valleys is bi-channel in several reaches and there is no evidence that the channels in these locations were created artificially, although considerable channel re-alignment and the cutting of small leets and bypass channels has occurred.

This evolutionary model relies on two driving processes, first the progressive abandonment of secondary channels and second the non-constant increase in floodplain and channel bank height caused by overbank deposition. Because palaeochannel bed height is relatively constant these two processes must have been related as there was a deepening of channels (and probably a narrowing) during the Flandrian as revealed by a decrease in palaeochannel W/D ratios (Brown & Keough 1991). This is related to the growth of silt/clay banks, but unlike the Schumm (1960) hypothesis the relation between W/D ratio and bank silt/clay percentage is fundamentally the result of fixed channel location, increased concentration of flow and accelerated overbank deposition. The stability of channels in an anastomosing system is thought to be largely the result of the development of bank and island vegetation (Smith & Smith 1980). The concentration of discharge into a smaller number of channels is compensated by the increased channel capacity produced by the increased bank height. This maintains continuity and the approximate constancy of the overbank return period. This is the basis of the SBAB model of constant bed elevation with an aggrading floodplain. This tendency towards little change in bed elevation but floodplain surface aggradation has been noted in other studies of Holocene stratigraphy such as the Duck river in Tennessee (Brackenridge 1984) and the Gipping in East Anglia (Rose *et al*. 1980).

Only one site, Wollaston, has convincing evidence that a channel was recently migrating, in this case between 3275 ± 50 years BP and *ca*. 3000 years BP. Migration seems to have been minimized by the progressive increase during the Flandrian of bank resistance caused by the deposition of silt and clay and the development of dense floodplain woodland. Between 4000 and 2000 years BP siltation may have been rather localized in old channels and floodplain depressions, which exist at all the sites owing to the greater relative relief of the floodplain before burial by the bulk of the superficial silt and clay unit. This differential siltation may have been a major factor in the transition from an anastomosing system to a single channel system with meandering planform channels, as currently exists in much of the Nene and nearly all of the Soar.

Important mechanisms include the selective siltation of some channels in a multiple channel system and adjustment of the non-silting channels. The siltation of side-channels is well known from historical examples, with the pattern being siltation at the upstream end producing infills which vary from entirely inorganic near the channel division to organic sediments in pools (Erskine *et al.* 1982). An early form of river engineering was the construction of groynes to deflect flow into one channel effectively increasing the siltation of secondary channels. The purpose was generally to maintain depths for navigation. One important factor in the siltation of secondary channels is greater sinuosity which lowers slope and shear stresses especially at moderate flood flows. The greater sinuosity is often a function of the channel size, which is probably one of the reasons why bi-channel systems with similar channel sizes seem to be both relatively common and stable. The lower slope of these more sinuous channels will cause velocities to fall below those critical to maintain coarse suspended sediment in suspension more frequently than in the main channel and reach critical shear stresses required for bed scour less frequently. Many other factors may be involved, an important one being the fact that the smaller channels are more prone to damming by organic debris and more likely to be dammed by beavers. Both situations are applicable to middle Flandrian Britain (Coles 1992). In addition the development of backwaters associated with secondary channels can cause chemical variations in water quality inducing flocculation of clays at channel divisions. While many U.K. systems metamorphosed to single-channel systems, it is noticeable that those with resistant banks, high dissolved and low suspended loads such as many of the Wiltshire valleys, did not and remain anastomosing today.

9. Discussion

This discussion concentrates on two areas. First, the role of climatic and anthropogenic change in relation to the evolutionary model proposed and second the need for multi-channel models of floodplain evolution in lowland temperate floodplains. The model proposed in this paper highlights two periods of channel and floodplain transition: first, during the very Lateglacial and earliest Flandrian, from braided and actively meandering to stable-anastomosing, and second, in the late Flandrian, from stable anastomosing to stable single-channel, sinuous to straight channels. The first transition was undoubtedly climatically induced, but the second is more complicated. In between these transition periods the floodplains were relatively insensitive to small to medium régime changes owing to a lack of available sediment and high bank resistance. The late Flandrian transition displays both hydrological change and alluviation. Although almost universal this alluviation is diachronistic reach to reach and valley to valley. Owing to the temporal imprecision of the alluvial record in the lowlands (especially of floods) and the status and relative magnitude of climatic discontinuities (Wendland & Bryson 1974) it is unwise to try and correlate climatic events with the existing stratigraphic record in the lowlands. The integrated effects of major land-use changes in large catchments during the Bronze and Iron Ages are difficult to quantify. However, the superficial unit did not come from bank erosion, but was derived from topsoil, as indicated by grain size and magnetic enhancement (Brown 1992). It therefore represents an increase in sediment availability and the most likely cause of this in temperate environments is a change

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

in soil erodibility caused by land-use changes, especially agricultural practices such as autumn or winter ploughing (Shotton 1978). The relative insensitivity of these lowland valleys to climatic change but sensitivity to changes in fine sediment loads stands in contrast to upland valleys (Macklin & Lewin 1989). The middle and late Flandrian floodplain evolution of the Nene and Soar valley channel systems is fundamentally seen as being driven by the inherited conditions of the system and catchment changes but with variations in floodplain aggradation rates probably reflecting periods of increased and decreased flooding.

Most descriptive models of low-energy floodplain evolution applicable to North West Europe have failed to consider the role of multiple channel or anastomosing systems. This is because although some exist today, they are rare partly because of deliberate channel training. Because this and other studies (Cheetham 1980; Rose *et al.* 1980; Dawson & Gardiner 1987) suggest that active multiple channel systems existed in the Lateglacial and we now know that the pace of climatic change was greater than the relaxation times of fluvial systems it is apparent that Flandrian floodplains must have inherited multiple channels in most areas. The multiple-channel hypothesis does not suffer from the apparent contradiction between evidence of channel stability with low-stream power and bursts of very frequent avulsion needed to explain the palaeochannel time-clusters. Any high frequency of avulsion is difficult to explain in the very early Flandrian when flow variability must have decreased after the waning of periglacial conditions during the Lateglacial stadial. This alternative approach based on the continuation of anastomosing systems, emphasizes the role of floodplain and channel sedimentation, and the evolution of floodplain vegetation, both of which were the product of climatic régime and human actions through alterations in sediment availability.

10. Conclusions

1. The Flandrian floodplain evolutions of the Nene and Soar valleys have been fundamentally very similar, however, there is some evidence of greater lateral instability and floodplain reworking in the Soar. This is probably due to its geologically influenced hydrological régime. Catchment response to climatic and anthropogenic environmental change during the Flandrian will differ because of hydrogeological influences on catchment régime.

2. In the Soar the sub-alluvial gravels were deposited between *ca.* 28000 years BP and 10200 years BP. The bulk of the sub-alluvial gravels (First Terrace) of the Nene were deposited between 11200 years BP and 10200 years BP. In the Soar the Lateglacial Stadial only caused minor reworking of the floodplain but in the Nene it caused a major incision and aggradation cycle. Studies involving detailed stratigraphic analysis of a number of reaches or sites is essential in order to avoid the erroneous generalization of erosion and deposition from one or two atypical but rather visible sites.

3. The principal processes (contained in the proposed model of floodplain evolution) which have influenced the evolution of both systems have been the development of an anastomosing system from an initial braided-river topography and its conversion to a predominantly single-channel system due to floodplain and channel siltation. This has been accompanied by a change in channel types and capacities, illustrated by the SBAB model. The sub-meandering or straight to sinuous nature of many lowland UK channels may be due to their evolution from an

anastomosing pattern where the least meandering survived along with the box-S shaped pattern. To test this more research is needed on the transition from braiding to anastomosing systems.

4. The floodplain stratigraphy of both valleys has been significantly influenced by soil development and sub-aerial processes, especially during the middle Flandrian when fluvial activity was at its minimum. Adequate modelling of Quaternary alluvial sequences should include a wider variety of processes and possible conditions than has currently been recognized. This would include the effect of different initial conditions, variations in floodplain groundwater tables, the differential siltation of secondary channels and sub-aerial floodplain processes, especially bioturbation. All of these have been important in creating the alluvial stratigraphy of the Soar and Nene valleys as described in this paper.

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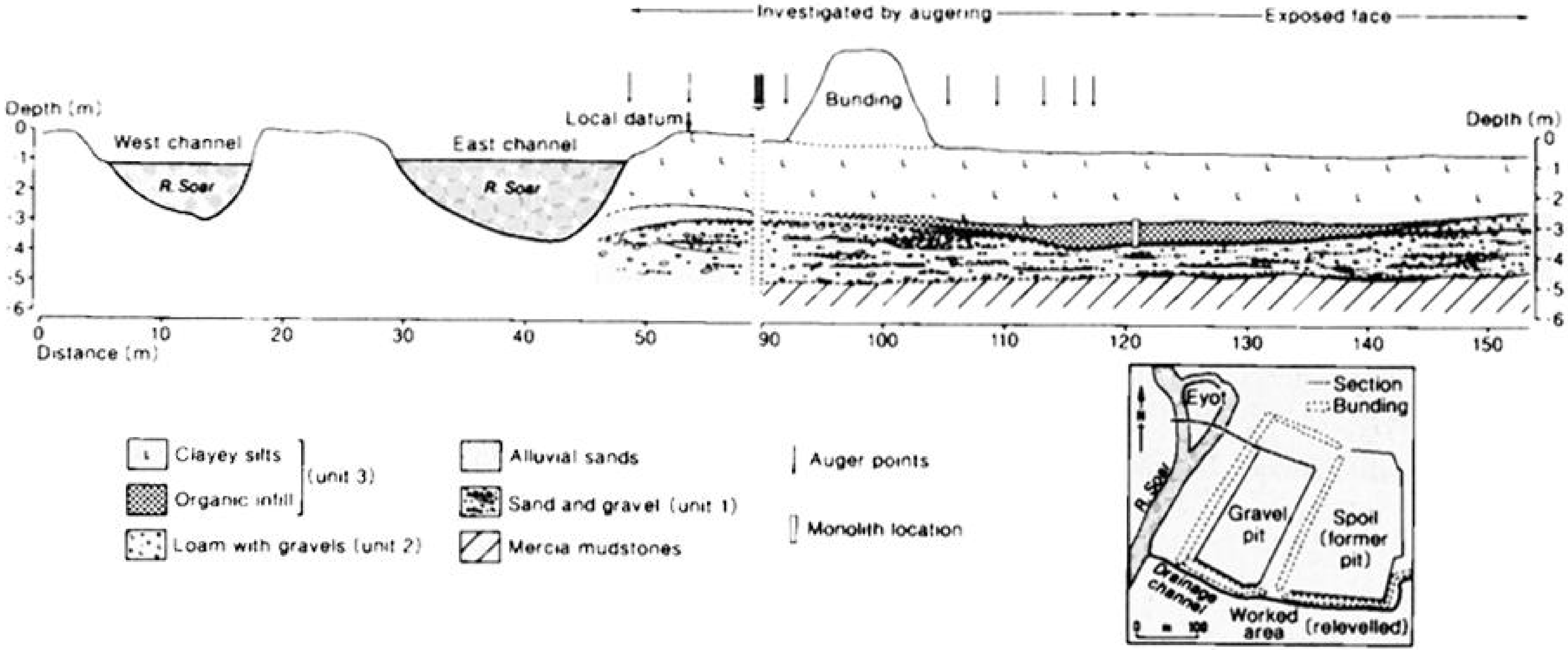

gure 5. Stratigraphic section from Cossington Monolith location is also the location of the block-excavation illustrated in figure 7.

Figure 6. The excavated gravel pavement at Cossington.

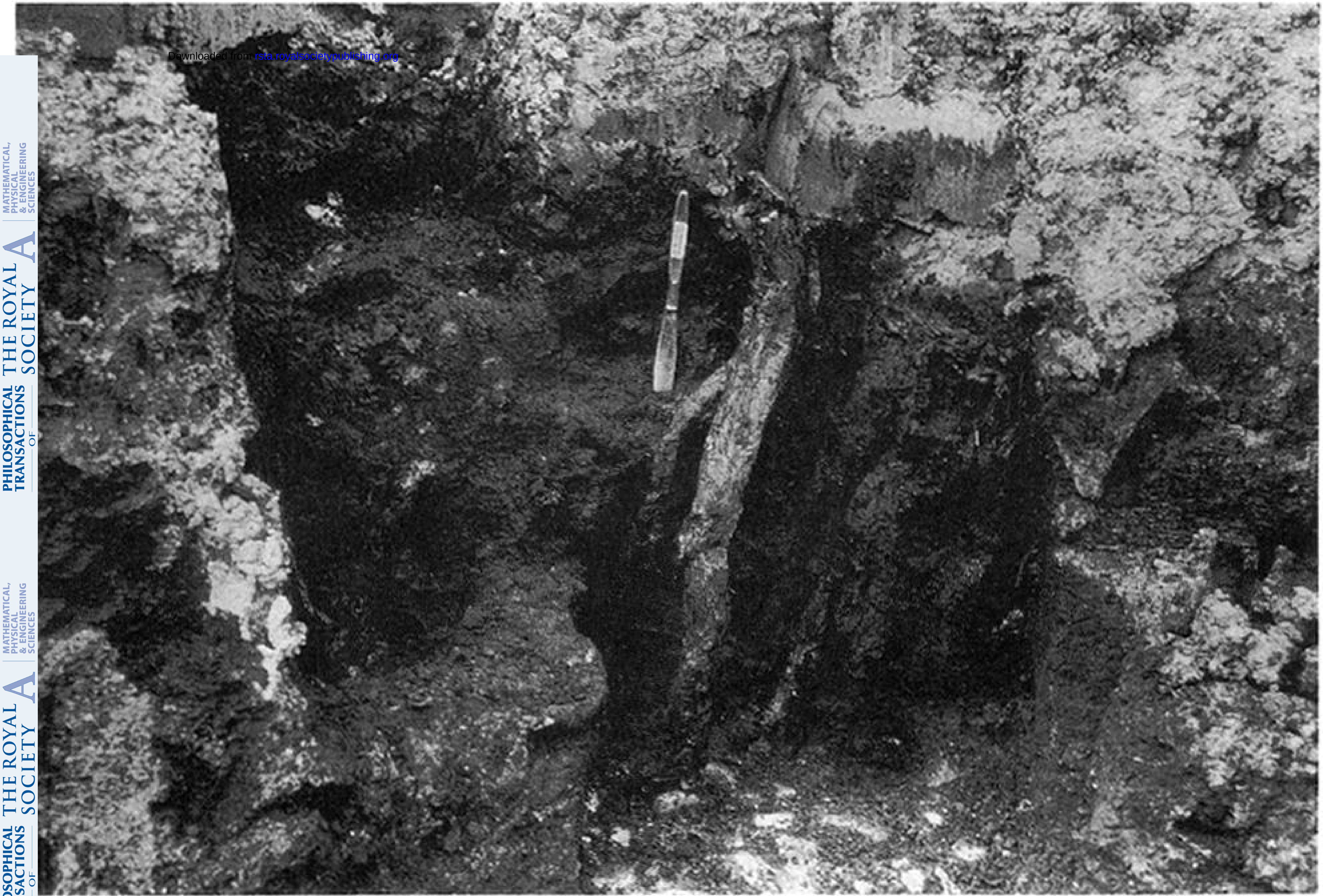

Figure 8. The excavated block at Cossington (see figure 7).