The Late Quaternary palaeohydrology of the middle Trent: a sedimentological study

The Late Quaternary Palaeohydrology of the Middle Trent — A sedimentological study

by

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Abstract

A systematic palaeohydrological investigation has been carried out on the floodplain gravels of the Middle Trent Valley at Barrow-upon-Trent. Consideration has been given to the relationship between floodplain stratigraphy and river morphology using sedimentological lithofacies analysis. Previous research has suggested that the floodplain deposits represent two distinct depositional processes which should be regarded as separate units, namely the 'Holme Pierrepont Sand and Gravel', and the 'Hemington Terrace' deposits.

Six distinct lithofacies types have been identified and interpreted in terms of origin and position in the riverine tract. Three gravelly (A, B, C), two sandy (D, E) and one fine grained (F) lithofacies have been distinguished as representing deposition by a braided river system. Lithofacies A, B and C are proposed as the deposits of horizontal gravel sheets and midchannel bars; Lithofacies D and E as downstream and lateral accretion elements, and Lithofacies F , as overbank fines (the presence of a large organic palaeochannel within the classification of Lithofacies F provides a radiometric date of 13060± 90 CAL yr BP).

The deposits, interpreted as best-fitting with the shallow gravel braided river (Scott type) model of Miall (1996) are suggested to have formed as a result of the outwash from the retreating Late Glacial (Devensian) Ice Sheet situated, at its maximum extent, to the west of the River Dove — River Trent confluence. The transition from a high-energy proglacial braid-stream environment during Late Devensian times through reduced energy conditions, interspersed with flood events during the Flandrian, and finally, to the low-energy nature of the current sinuous channel is suggested to be more complex than previously thought.

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1.1 Introduction

Covering an area of 10435 square kilometres and draining much of the Midlands of England, the Trent River Basin (Fig.1) is one of the major contemporary river systems in the U.K. The Pleistocene deposits of the Middle Trent have been well documented (Straw & Clayton, 1979; Jones & Charsley, 1985; Brandon & Sumbler, 1988; Brandon, 1997) and, although there is no universal agreement on the precise nature and timing of events, the general glacial history of the area is in little doubt. The original mantle of Pleistocene deposits, underlain in most part by Triassic sediments, particularly Mercia Mudstone (Fig.2) has been dissected by subsequent erosion during Late Quaternary and Recent times. The products of this erosion are partly preserved in the terrace gravels and alluvial deposits lying in the main valleys. The solid and drift geology of the Middle Trent is provided with reference to figure 3.

The floodplain deposits of the River Trent, previously regarded as the 'floodplain terrace', have been differentiated into an upper and lower terrace. The lower 'Holme Pierrepont Sand and Gravel' (after Charsley et al, 1990) is regarded as Late Devensian in age whilst the upper 'Hemington Terrace' deposits (after Brandon, 1997) are interpreted as having been accreted during 'possible Late Glacial times into the Late Flandrian' (Brandon, 1997).

Much of the previous work on the floodplain deposits of the River Trent has focused on the nature of the sediments testifying to the overall environmental and climatic regime at the time of aggradation. Latest research (Brandon, 1997) has suggested a cold-climate braided system (Holme Pierrepont Sand and Gravel) followed by a meandering planform during a warmer period (Hemington Gravel).

This research will consider more closely the palaeohydrological relationship between floodplain stratigraphy and river morphology, with particular regard to sedimentary facies analysis. Floodplain sediments are examined with a view to the identification of distinct lithofacies and the interpretation of these lithofacies in terms of origin and position in the riverine tract. Environmental conclusions are drawn based on this information.

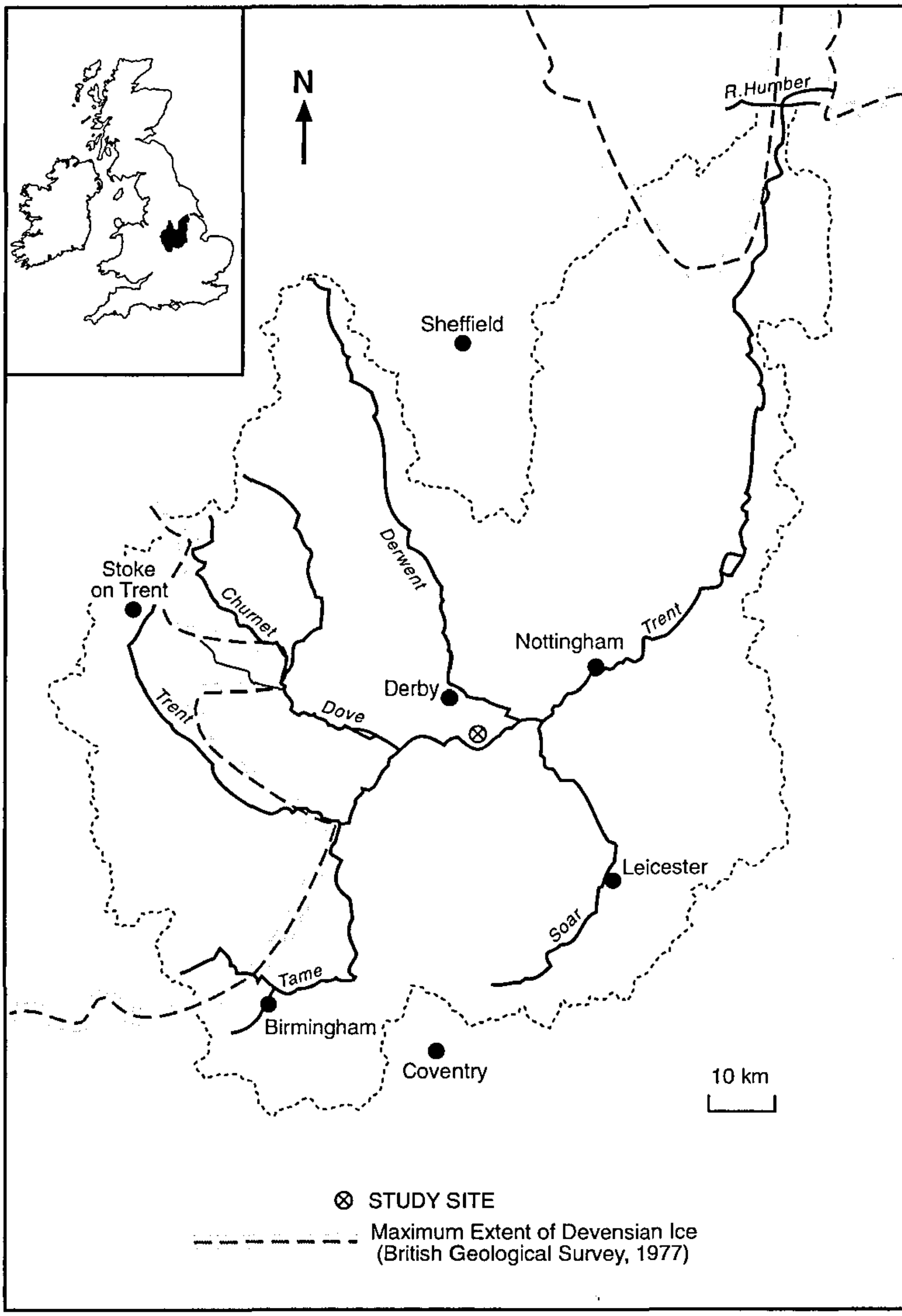

Fig. 1 Trent River Basin

A systematic palaeohydrological investigation was undertaken on the floodplain deposits at Barrow-upon-Trent, the site of an ongoing sand and gravel operation (Lafarge Minerals Ltd.). The findings of this investigation are presented in this report.

1.2 Previous Research on British River Systems

1.2.1 Context

To fully understand the relevance of previous research to this study it is necessary to outline the general climatic and environmental conditions that prevailed during the Late Pleistocene.

The Quaternary period, or Pleistocene began approximately 1.8 million years before present (BP) and is conventially subdivided into colder Glacial and warmer Interglacial stages. These stages are further divided into relatively short-lived Stadial (cold) and Interstadial (warm) episodes. These episodes are extremely important since they are associated with ice readvancement/retreat and have a marked effect on depositional regimes (see below). In Britain it is widely accepted that the last cold stage, the Devensian (Weichselian in Northern Europe), is divided into a Lateglacial (or Windermere) Interstadial and a Loch Lomond Stadial (Younger Dryas in Northern Europe). The end of the Loch Lomond Stadial, at approximately 10000 ^{14}C years BP, marked the close of the Devensian and the return to the warmer conditions of the Flandrian (Holocene in Europe).

The maximum expansion of Devensian ice occurred during the latter part of the cold stage between 26000BP-13000BP (Dimlington Stadial) and although ice sheets covered much of northern England, the Trent Basin was largely unglaciated (Fig.1). Previous glacial sediments originally deposited over large areas of the Trent Basin are now found as a dissected and eroded mantle largely occupying higher ground and thought to be Anglian in age (over 430000BP). Withdrawal of the Late Devensian ice sheets would of course led to an increase in meltwater to headstreams and promoted massive

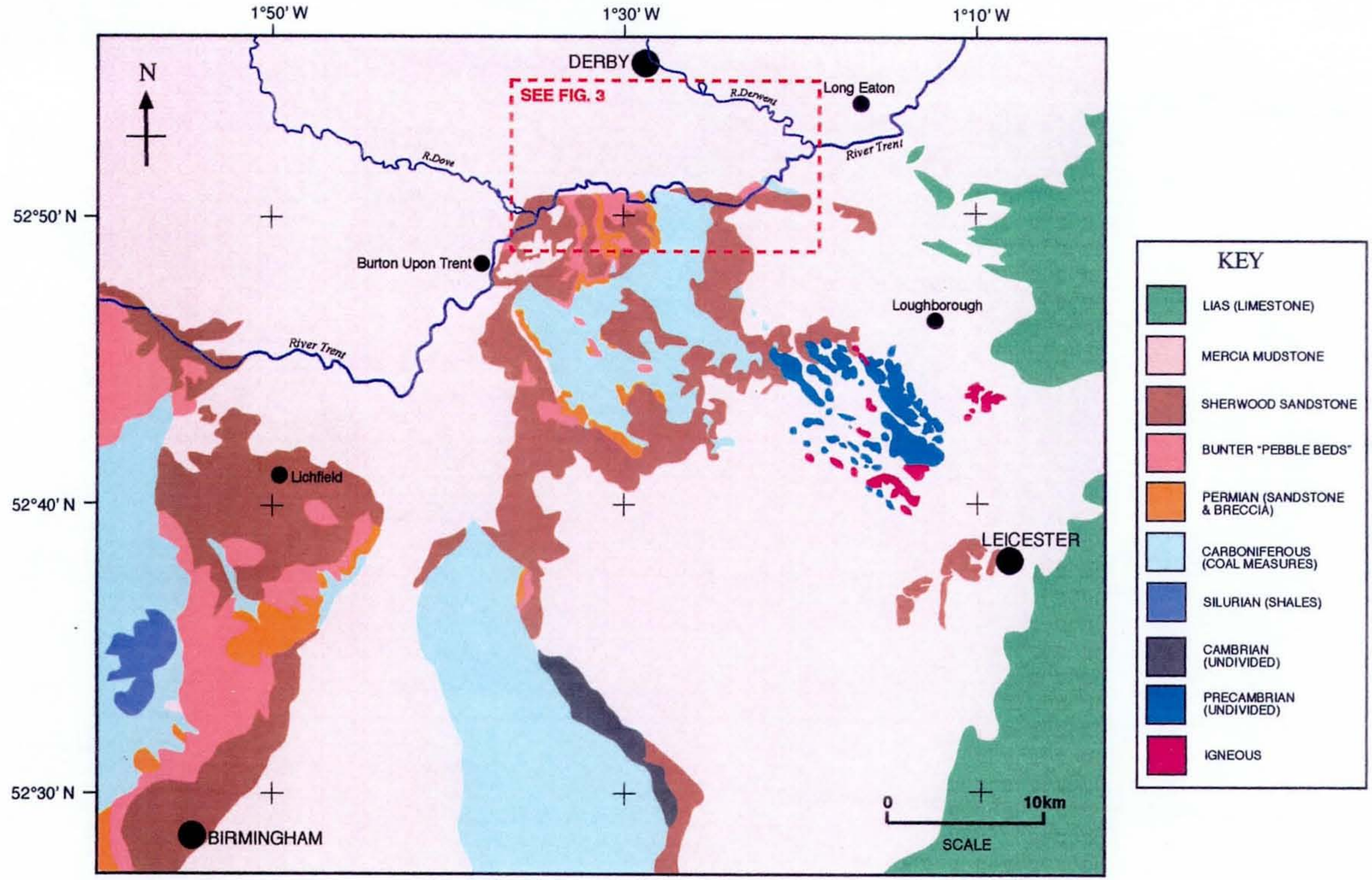

Fig. 2 Generalised regional geology (solid)

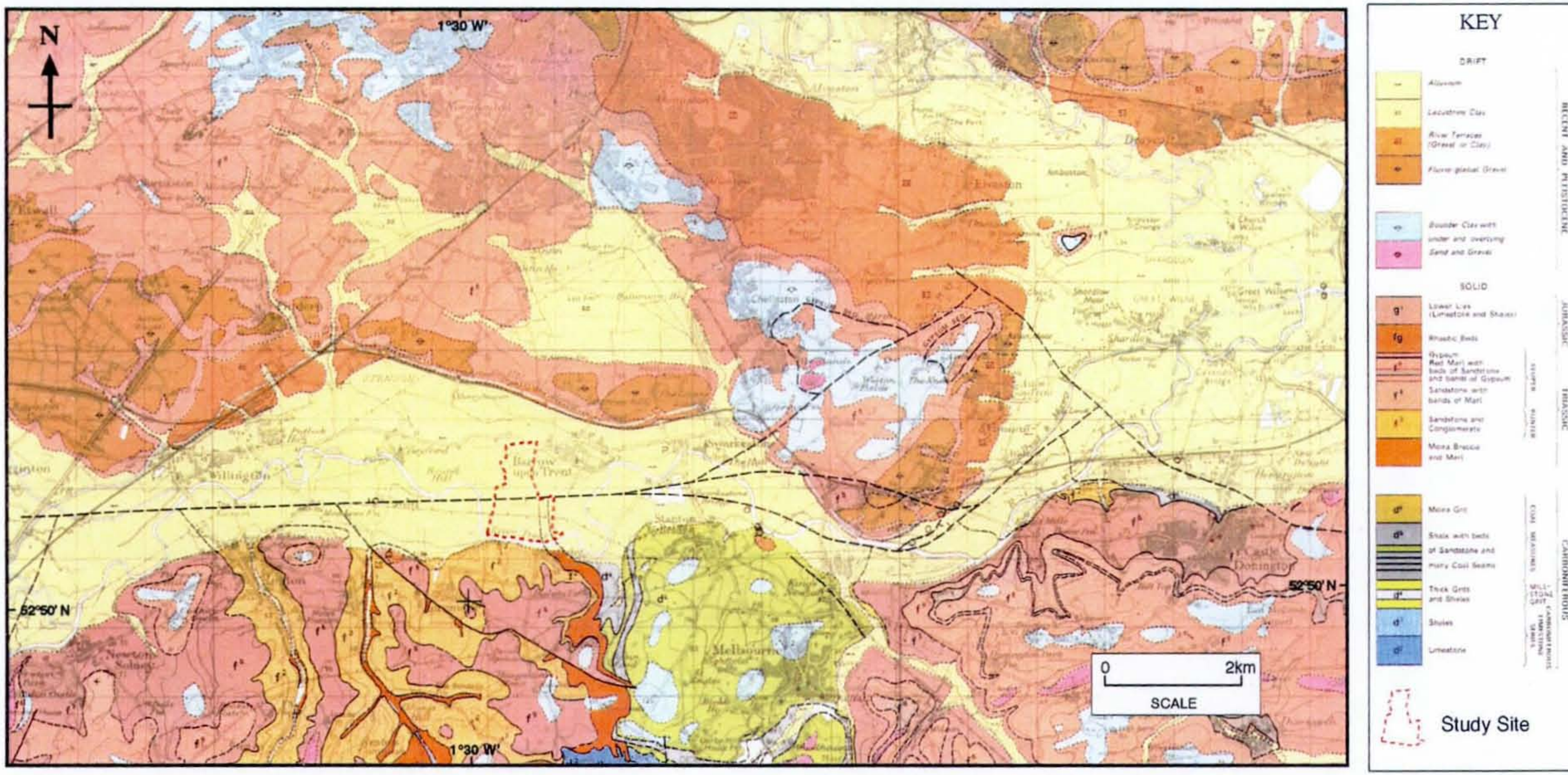

Fig. 3 Middle Trent - Solid and Drift Geology

sediment supply into the valley basin. Of course the depositional environment was very much more complex than outlined here. The Younger Dryas (11000 – 10000 BP) for example, illustrates the complexity of change that can occur within a relatively short period of time. This period was characterised by glacial readvancement which had a marked effect on the erosional/depositional regime, and consequently, channel stability.

Readvancement would very likely have led to a reduction in meltwater, however, it is probable that the Trent was subject to major channel instability as a result of periods of high, short-lived discharges caused by seasonal snowmelt. The change from erosion to depostion is related to the relative magnitudes of the peak discharges and the available sediment supplied to the channel from adjacent hillsides (Rose et al, 1980). An added complication arises when a system, such as the Trent, is subject to periglacial activity (as was the case during the Younger Dryas) leading to different depositional mechanisms (i.e. hillslope gelifluction) and additional discharge sources (i.e. spring and summer melt of seasonally frozen ground).

1.2.2 Research

A significant part of this research concerns the sedimentological features of the floodplain terrace gravels of the Middle Trent. Other aspects of palaeohydrology equally as important, include palaeoclimatology, basin analysis, geomorphology and geochronology.

Research relevant to this palaeohydrological investigation involve that which examines river basins and terraces thought to have experienced similar depositional processes. Many of these studies are based on historical examples, however modern-day analogies are available and include research carried out in Arctic, Canadian and European river systems (Bryant, 1983; Cant and Walker, 1978; Yamskkikh, 1994,1996).

Palaeohydrological and associated studies have been carried out on a number of British rivers. Popular areas for investigation include the Severn, Avon and Thames Basins, although the deposits of the rivers Soar, Nene, Ouse and Gipping amongst others, have also been subject to scrutiny (Brown, 1994; Rose et al, 1980).

The largest British river basin, the Severn, has been subject to a great deal of research. Brown (1991) has studied the Late Glacial and Flandrian development of the Severn, suggesting that the change from a high-energy braided system to a low energy anastomosing-meandering system could be interpreted at a scalar level. This suggests that the river system is a function of the changes brought on by different magnitude events such as a major climatic change at one level ('first order changes') or single flood episodes ('fourth order changes') at another. Similarly, Rose et al (1980) have looked at the form and development of the channel of the River Gipping over the past 13,000 years and concluded that from a relatively stable condition in the early and middle parts of the Windermere Interstadial (Coope and Pennington, 1977) the river became unstable in the later part of the Interstadial and whole of Younger Dryas, and then stabilized again through the Flandrian. This interpretation, derived from geomorphological (i.e. incised channels), sedimentological and palaeoenvironmental (i.e. macroscopic plant remains, coleoptera, mollusca) evidence suggests that the instability is attributed to high, short-lived discharges caused by seasonal snow melt during this time. The floodplain evolution of the River Nene during this Lateglacial-Flandrian period has also been studied. Brown (1994) has suggested that the evolution of the Nene is fundamentally similar to that of the River Soar and has proposed a model that may be suitably applicable to other temperate lowland rivers. Whilst recognising that there are local variations (such as basin size and geology) the model suggests that many lowland rivers experience the braided - active meandering – stable anastomosing - sinuous change during this climatically active period

The terrace deposits of a number of British rivers have been investigated. Dawson and Gardiner (1987) give a review of the terrace deposits of the Severn with particular regard to palaeohydrological modelling, whilst Sumbler (1995) has proposed a correlation between the terraces of the upper Thames with the River Thame on the basis of long profiles. This has in turn suggested a two-phase Anglian glaciation for the region.

Sedimentological and facies analysis has been carried out on many river terraces. Research by Maddy et al (1994) has reappraised the Middle Pleistocene fluvial deposits of the River Avon and looked at their significance for the Wolston glacial sequence. The sedimentological study identifies lithofacies types and, combined with environmental and climatic information gathered from organic fills, suggests that the sequence is much more complex than previously thought.

Many palaeohydrological studies examine the sedimentological properties of terrace deposits (such as maximum clast size) in combination with channel dimensions and morphology. Dury et al (1983), studying the Severn, placed particular emphasis on the reconstruction of palaeo-discharge based on former channel dimensions. Maizels and Aitken (1991) examined deglaciation in upland areas of N.E. Scotland and, whilst recognising the limitations of palaeodischarge methodologies, agreed that valley terrace systems contain 'significant evidence of the nature of hydrological changes occurring within these catchments during, and since, deglaciation'.

The environment and processes which led to the development of ancient terrace deposits are clearly very different to those now operating in the U.K. Comparison with modern day analogies allow us to better understand this depositional environment. The floodplain gravels of the River Nene were studied extensively by Castleden in 1976. Using modern day analogies from the Arctic, Castleden proposed that the composition, morphology, stratigraphy and age of gravel flooring the Nene Valley provided evidence of ancient river behaviour very different than that at present. He suggested that the Middle Devensian River Nene was certainly braided and carried a coarse and 'excessive' load which led to aggradation of the floodplain deposits. The adaption to the meandering habit in a large single channel occurred during Flandrian times. Analogy with Arctic river systems has also been attempted for a number of other British river systems (Bryant, 1983).

Many other smaller river systems have been studied over recent years. These studies have looked at specific but ultimately overlapping aspects of palaeohydrology including sedimentation (River Tywi; Smith, 1989), palaeoflow (Kennet Valley; Cheetham, 1976), stratigraphy (Upper Axe Valley; Macklin, 1988), valley floor development (River Swale; Taylor and Macklin, 1997) and climatic reconstruction (Wye Valley; Hey, 1991).

1.3 Middle Trent Valley

The Pleistocene deposits of the Middle Trent Basin have been the subject of much discussion by various writers.

An account provided by R.M. Deeley in 1886 gives descriptions of the 'Interglacial river' terrace gravels and presents a basic summary of the Pleistocene successsion in the Trent basin. Swinnerton (1937) differentiates the terrace deposits into two, namely the 'Flood-plain' Terrace and the Beeston Terrace (equating with the Second Terrace of the Pocock (1929)). Later added to these was the Hilton Terrace as proposed by Clayton (1953). Posnansky (1958) suggested that the deposits, recognised as the 'Floodplain Terrace', were certainly post-Eastern Glaciation and, although Rice (1968) said that 'no general agreement on their (the terraces of the Trent) interpretation has emerged', it is now widely accepted that the 'Floodplain' deposits of the Middle Trent are a Devensian, or later phenomena.

Previously referred to by many authors as the 'Floodplain Terrace' (after Swinnerton, 1937) and then the 'Floodplain Sand and Gravel' (Brandon and Sumbler, 1988) the floodplain Sands and Gravels of the Middle Trent are now differentiated into two distinct horizons, each representing different depositional processes (Table 1). The underlying 'Holme Pierrepont Sand and Gravel' (after Charsley et al, 1990) is said to form the bulk of the deposits across the floodplain and consists of sediments reworked from earlier depositional events. The deposits of the 'Hemington Terrace' (after Brandon, 1997) are thought to represent a post-glacial phase and consist of sediments reworked from the underlying Holme Pierrepont Sand and Gravel.

In contrast to many earlier authors, Brandon, (1997) has distinguished between the terrace features per se and the deposits that comprise the terrace. Accordingly, this form of classification will be followed in the coming chapters.

Quaternary Stage	Approx. Age (commencement)	Conditions	Terrace
Flandrian	10k	Warm	Alluvium Hemington Terrace
Younger Dryas	11k	Cold	Holme Pierrepont Sand and Gravel
	26k		
Devensian	65k		
	80k		Beeston Sand and Gravel
	115k		
Ipswichian	128k	Warm	
Wolstonian	195k	Cold	Eggington Common Sand and Gravel (Lower Hilton Terrace Deposits)
	240k		
	297k		Etwall Sand and Gravel (Upper Hilton Terrace Deposits)

Table 1

1.4 Objectives

This contribution aims to describe and interpret the floodplain deposits of the Middle Trent using palaeohydrological techniques, particularly sedimentological and facies analysis, and provide a review of the environmental conditions that prevailed at the time of aggradation. The discussion will consider the following:

- During the adaptation of the Trent Basin from an ice marginal system to an interglacial regime, how and when did the change occur from a braided to a meandering reach?
- What is the relationship between floodplain stratigraphy and river morphology? Can the deposits of the Lateglacial-Flandrian / braided-meandering systems be confidently distinguished?

- If, as suggested, there was reworking of Lateglacial sediments during Flandrian times is it possible to determine exactly when this happened and what pre-empted this?
- How does the nature of the sediments at Barrow-upon-Trent compare with other reaches of the River Trent?
- How relevant is the Trent basin to established depositional models?

2. Methodology

2.1 Understanding River Channel Changes

2.1.1 Palaeohydrological Overview

The evolution of the fluvial environment is largely a function of the interactions between water discharge and sediment transport conditions. These fluvial processes are continuously changing through time with the influence of climatic and other variations. The changing rates and character of fluvial activity inevitably leads to changes in morphology and sedimentology of the river system.

The science of 'Palaeohydrology' has been defined by a number of authors since the 1950's (Leopold and Miller, 1954; Schumm, 1965; Gregory and Walling, 1973; Cheetham, 1976; Dury, 1977). The most widely quoted definition, provided by Schumm (1977) and amended by Gregory (1983), is that 'Palaeohydrology can be defined as the science of the waters of the earth, their composition, distribution and movement on ancient landscapes from the occurrence of the first rainfall to the beginning of continuous hydrological records'. The major aim of fluvial palaeohydrology lies in achieving the reconstruction of former river systems based on the reconstruction of the conditions and processes that operated in such environments. The reconstruction of the palaeohydrological cycle is achieved by utilising a wide range of techniques and approaches that reconcile data from a number of different scales (Gregory, 1983). The purpose of this section is to give a brief review of the methodologies available for palaeohydrological reconstruction. A number of techniques have been deliberately omitted from this review because, whilst they are recognised as very important tools in environmental reconstruction, they are not immediately relevant to this research. These include sea level analysis, palaeopedology, thermoluminescence and magnetic studies.

As an interdisciplinary field, palaeohydrology has foundations in geomorphology, palaeohydraulics, palaeoclimatology, and palaeoecology and archaeology. Geomorphological methods aim to distinguish channels, floodplains and terraces, and estimate palaeodischarge. The division into braiding, meandering and transitional rivers is based on the detailed characteristics of channel parameters: width (w) depth (d) length of meander (l) river gradient (s) and sinuosity. Formulae and equations involving those parameters (Leopold et al, 1964. Dury, 1977a; Schumm,1977) can be used for the reconstruction of the palaeohydrologic regime (Starkel, 1983). Differences in sedimentary textures, bedforms and facies types reflect changes in hydraulic conditions, particularly in velocity, bed shear stress and stream power (Gregory and Maizels, 1991). Gregory and Maizels have defined these sedimentary parameters within a 'hierarchical scalar structure'.This ranges from individual clastic particles at a single point or cross-section of the palaeochannel, through to aggregates of particles as bedforms and bars, described stratigraphically in terms of their lithofacies characteristics and, at a larger scale where the intrinsic processes, channel pattern and sedimentology of a river system can be defined in terms of a particular facies model. The palaeohydraulic significance of fluvial sediments therefore varies depending on the scalar level at which the sediments are considered. Smaller scale features are widely used as a means of interpreting the flow hydraulics whilst larger scale features can indicate former fluvial conditions and long term changes in both fluvial activity and depositional history.

The prediction of former fluvial discharge is a primary goal of palaeohydrology since it influences the formation of a particular channel pattern (Leopold and Wolman, 1957) the hydraulic geometry and the accumulation of sediments. Most methods of palaeoflow determination are based on direct analogies either with hydraulic relations between sediment and flow parameters in present day rivers or flumes, or with geomorphic relations between channel morphology, channel sediments and discharge measures observed in present day rivers (Maizels, 1983). The development of these two approaches has produced a large number of possible methods for palaeovelocity and palaeodischarge determination. Theoretically, it is possible to determine past river discharge on the basis of its effects, firstly through analysis of deposits of former streams (e.g. Dawson and Gardiner, 1987) and secondly through analysis of preserved morphological effects caused by a given discharge (e.g. Maizels and Aitken, 1991). Unfortunately, the relationships between the deposits, structures and statistical parameters are so complex and ambiguous that they cannot be expressed in any current quantitative models. None of the methods of palaeovelocity or palaeodischarge analysis (e.g. Manning, Chezy, Darcy-Weisbach equations) provide particularly accurate results due to the difficulties in determining former channel conditions and because of the procedural assumptions which have to be introduced during the calculations. However, they may at least indicate the order of magnitude of palaeoflows.

The climatic regime, in particular air temperature, evapotranspiration, precipitation and their annual variability, is one of the main parameters causing change in the fluvial system. Intimately related with these climatic variables is vegetation (cover and type) which has a significant effect on runoff characteristics, sediment yield and erosivity of the channel banks. Temperature change can be deduced from palaeobotanical information and coleopteran remains, together with the presence of periglacial phenomena (i.e. ice wedge casts) during colder cycles. Such features have been recognised by Brown (1992) who suggests that the Devensian and Holocene sediments of the River Trent are only distinguished at one location by the presence of ice wedge pseudomorphs truncated by medieval gravels.

The presence of organic remains are also extremely important for age correlation, and radiocarbon dating probably remains the most commonly applied geochronological tool in fluvial palaeohydrology (Baker,1991). The value of the radiocarbon method stems from both its accuracy and the wide range of materials that can be analysed, including charcoal, wood, fine grained organic detritus (twigs and seeds) organic rich palaeosols and peaty deposits. The delicate nature of these deposits, however, allows that they only develop in sites protected from channel erosion for a sufficient period.

One of the most important controls on channel adjustment during the Flandrian must be that of human impact. Our understanding of this impact on the hydrological landscape is mainly deduced from archaeological evidence (e.g. Knight and Howard, 1994). Human activity may be classed as direct or indirect, and either way, these effects may have a catastrophic consequence for the fluvial regime. There is no doubt that the impact of human activity is more important in Holocene times than earlier. Deforestation and cultivation of soils have a direct effect on runoff rates and sediment load variation, as do the modifications of drainage networks (Starkel,1991a).

2.1.2 River Terraces

Representing abandoned floodplains, River terraces form a fundamental part of the fluvial system. They may occur as corresponding sets on both sides of the valley, termed 'paired' terraces, or as a single matchless terrace on either side of the valley. Paired (or 'cyclic') terraces form as a result of rapid incision whilst unpaired (or 'non-cyclic') terraces are formed as a consequence of lateral channel migration and erosion such as experienced in a slower meandering reach (Schumm, 1977).

River terraces can develop by direct incision into bedrock, or more commonly, from episodes of aggradation and incision of unconsolidated alluvial sediments. The destruction and reworking of alluvial deposits by subsequent fluvial activity leads to older terraces being only preserved in fragments, usually at higher levels.

The development of terraces by either river incision or aggradation may result from a number of factors. These include changes in climate, sea level (base level) and tectonism, and the effects of glaciation and human activity. A model of fluvial terrace formation has been recently proposed by Bridgland (1984) and revised by Bridgland and Allen (1995). The model considers the development of the terraces of the Thames and suggests a multi-phase cycle of aggradation and incision activated initially by climate set against a background of localised isostatic adjustment. Incision, controlled by eutectic changes, under cold climatic conditions is superceded by aggradation as erosion is exceeded by sediment supply. A phase of reduced deposition, under temperate conditions, is then followed by renewed incision and aggradation as climate deteriorates. As discharge exceeds sediment supply, incision again becomes the predominant activity.

The importance of river terraces lies in their inherent ability to provide valuable evidence of former fluvial conditions. The sedimentology of terrace deposits provides essential information with which to reconstruct palaeochannel morphology and dynamics, and further increases our understanding of the processes that operate in the fluvial environment. River terraces also represent a reference level in a river system and give a relative chronology which can be related to other palaeohydrological events and

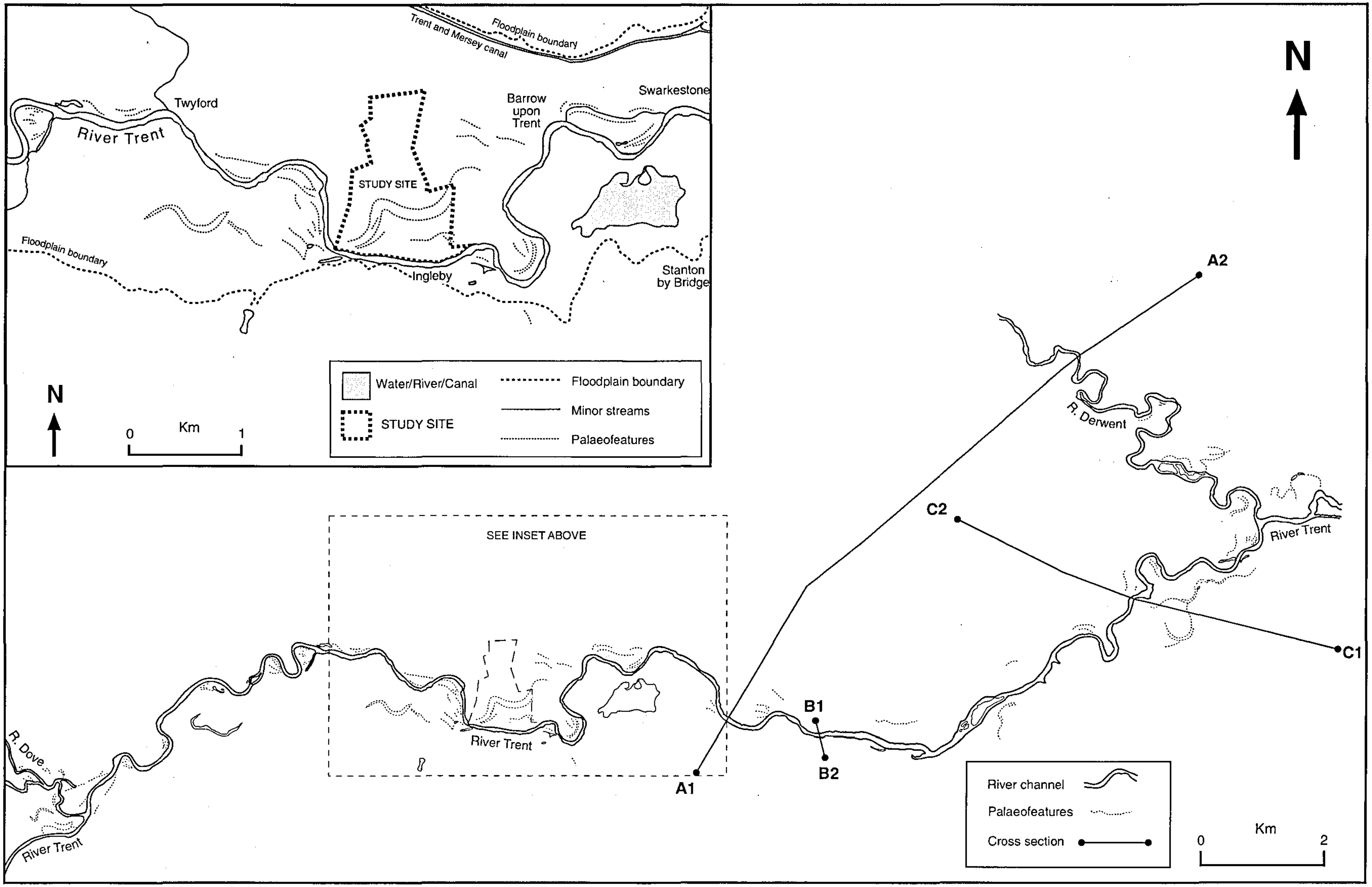

Fig. 4 Middle Trent - Location of study site and geological cross-sections

provide a correlation between sequences in different areas. If terrace deposits contain relict organic material then radiocarbon dating is possible, in addition to the inferences that can be made as to the former climatic and hydrological conditions. In Britain the terrace deposits of the Thames, the Severn and the Trent have been designated into glacial/stadial and interglacial/interstadial stages partly on the basis of organic remains. (eg Bridgland, 1994; Gibbard, 1994).

2.2 River channel changes on the Middle Trent

Palaeohydrological reconstruction relies on evidence from a number of sources including aerial photography, ground surveys (topographic, geological and geomorphological), stratigraphical and sedimentological analysis, and age determination. The purpose of the section following is to briefly explain the relevant methodologies that were used in this research.

2.2.1 Regional context

Providing an accurate picture of the prevailing Late Glacial and Flandrian conditions of the Middle Trent necessitates that palaeohydrological evidence is considered over the full reach. The approach taken in this research is typical of a number of palaeohydrological studies on British and continental rivers (e.g. Lower Severn basin – Brown, 1983; Rivers Twymyn, Dee and Teme – Lewin, 1983; Esk, Dee and Don valleys – Maizels and Aitken, 1991; Rivers Nene and Soar – Brown, 1994).

Valley floor morphology has been determined from the interpretation of aerial photographs and the comparative analysis of Ordnance survey maps (1:10000 and 1:25000). Palaeochannel scars, representing previous courses or incursions of the Trent, were identified by means of aerial photographs (Fig.4) whilst floodplain delineation (Fig.5) and topographical variation was established from the combined photographic and map information.

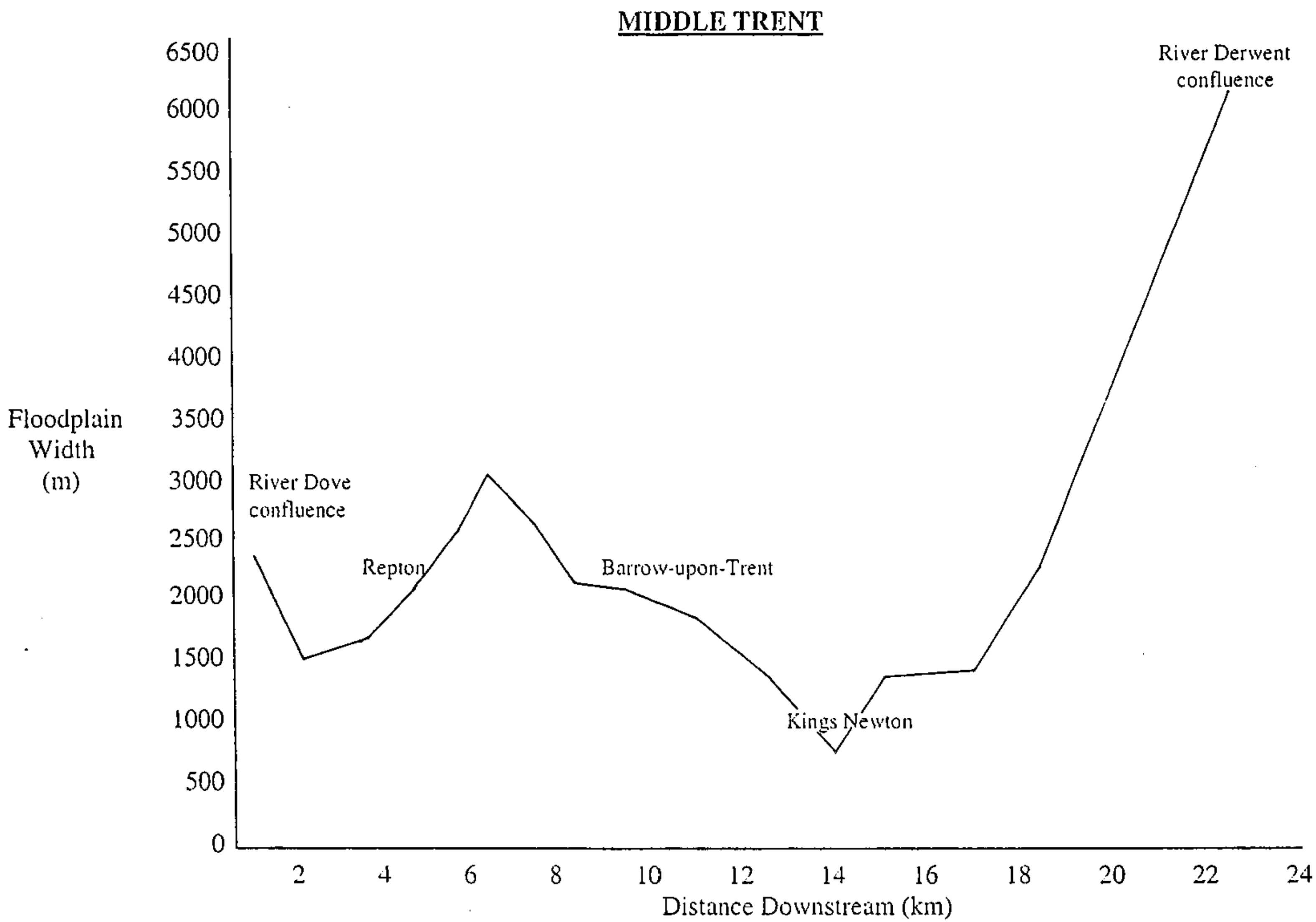

Fig.5 Floodplain Width vs. Distance Downstream

To fully understand the geomorphological and sub-surface structure it was necessary to examine borehole evidence along the valley. Stratigraphical correlation from boreholes and hand augering is a technique that has proven invaluable in palaeohydrological investigations of lowland river valleys (e.g. Severn valley – Brown,1983; River Soar – Bradley and Brown, 1992). Apart from studying working gravel pits, the examination of existing borehole/trial pit evidence provides a wealth of information that would otherwise not be available.

Many thousands of boreholes have been drilled on the floodplain between the confluence of the River Trent with the Rivers Dove and Derwent over the last 50 years. They have been drilled for a multitude of reasons including mineral exploration, hydrological evaluation (including natural water and sewage), civil engineering schemes and stratigraphical correlation. The majority of these records are held by the British Geological Society under the general agreement (and to protect economic investment) that they are not released to the public without the prior permission of the company or government agency concerned, or until a period of 25 years has elapsed. Despite the problem of accessibility, the borehole records reveal many important features upon the surface and subsurface geomorphology of the River Trent floodplain. Before concentrating on these geomorphological features, it is worth briefly considering the relative usefulness of the borehole data.

The borehole logs provide valuable information on Ordnance Datum (O.D.) levels and enable morphological features to be identified on the floodplain, in addition to reflecting the long profile of the river. Essential information is provided on the geology of the deposits, their depths and whether any special features exist within the deposits. It should be acknowledged, however, that various agencies are used to record different boreholes, some of which describe and class the deposits differently. This causes a problem in interpretation and correlation between boreholes since only the written log remains and inevitably this leads to 'sifting out' of data. Most boreholes are drilled with reference to particular schemes, i.e. motorway link, mineral extraction, groundwater levels, and are therefore logged appropriate to that scheme. On many occasions this does not provide all the relevant information for correlation between cores, i.e. a certain project may only be interested in the thickness of 'overburden' above bedrock and not the division of units. The uneven spatial distribution of boreholes also means that many boreholes are often superfluous to requirement whilst in other areas a scarcity of information exists. Additionally, borehole distribution is purely a reflection of a particular drilling scheme (i.e. gravel pit, road link, etc) and the available floodplain area. This may explain why a predominance of borehole information exists on the northern floodplain of the River Trent (greater floodplain area).

2.2.2 Local context : Barrow-upon-Trent site

Detailed investigations of single sites provide valuable information locally and, by inference, valuable data for regional analysis. The main study site at Barrow-upon-Trent (Fig.4) covers an area of over 75 hectares on the northern bank of the River Trent at one of the floodplains widest points (2km). The site is bounded to the south by the contemporary channel of the Trent and at its furthest point is 1.25km away.

Over 150 boreholes have been drilled at Barrow-upon-Trent over a number of years. This large database, together with the regular distribution of the boreholes provides extremely useful information on the subsurface geology of the deposits. A comprehensive topographical map complements the borehole data and provides additional information on geomorphological features. A total of 38 boreholes were drilled at the Barrow-upon-Trent study site in 1986 (refer to Fig.9) in order to assess the future economic viability for sand and gravel production. These boreholes were relatively evenly distributed across the site and detailed the basic geological structure of the site. Reference was also made to other features such as the presence of cobbles, coal/charcoal fragments and organic deposits. Recording the height of distinct geological units within each borehole allowed a three-dimensional image to be created of that unit across the site. This highlights any geological and geomorphological variation in an easily viewed format. A similar approach, looking at fine alluvial sediments, has been used previously on the River Nene (Needham and Macklin, 1992).

The images created were of a relatively simplistic nature and it was considered that the GIS IDRISI package was most appropriate. The particular application used was INTERPOL, a data entry module that interpolates a full surface from point data. The interpolation relies on data input as vector format, which in this case meant that 'X' and 'Y' co-ordinates represent the location of the borehole on the grid and the 'Z' co-ordinate was the Ordnance Datum height of the particular geological unit. The profiles that were generated are discussed in the results section.

It is important that study sites contain the required full sedimentary sequence in order to assess changing conditions over time. The presence of laterally extensive working quarry faces at Barrow-upon-Trent facilitated this. Sedimentological study included detailed logging of the lithographical units, taking account of thickness, colour, grainsize and structural variation and the relationships of particular units with others. Thirteen locations (refer to Fig.10) were chosen for detailed study. The locations provided accessible, exposed sections selected mainly on the basis that they illustrated particular lithofacies types and showed the vertical relationships of the lithofacies with each other. Studying laterally extensive faces allowed any lateral variation in sedimentology to be recorded and highlighted the overall relationships between lithofacies types and sedimentary sequence (eg. such as matrix-supported orange gravel overlying clast-supported pink gravel). Other locations were selected because they contained palaeoenvironmental data or recorded the presence of special sedimentary features, eg. clay drapes.

Eight detailed sedimentary logs were carried out along three major quarry faces, with field sketching and photographic documentation providing additional support. Palaeocurrent analysis was employed where possible, data being largely obtained from the measurement of cross-stratified sand sets (5 locations) and pebble imbrication (1 location). In-situ bulk samples of the various lithological units were taken for further facies analysis, the laboratory tests including particle size, clast composition and shape, and relative palaeodischarge estimation. Fifteen samples, selected from recognisable lithofacies types, were taken for coarse grainsize analysis (i.e. greater than silt grade). The samples were dried, weighed and screened through sieves ranging from –5phi (32mm) to +5phi (0.032mm) and the results plotted as standard grading curves. Additionally, seven samples were selected from the major palaeochannel deposit at location 1 for fine grading analysis (clay/silt/sand content) Samples were taken at 10cm intervals from the upper surface of the channel to its base and submitted for grainsize analysis by sedimentation (Pipette method, B.S.1377, Part 2). This was carried out primarily to distinguish sediment grading throughout the palaeochannel and to give an indication of relative deposition. The method involved mixing approximately 25 grams of sediment with a water and Calgon solution (antiflocculent) and then passing through a 4phi sieve to separate off the sand fraction. The silt and clay fraction was further mixed with water and made up to 1000mls. By pipetting off 25mls of sediment-water mix at varying time intervals, and taking into account the Calgon percentage, it was possible to calculate the clay and silt fraction of the sample (applying Stokes Law). Organic carbon analysis was also carried out on the fine-grained channel sediment samples in order to establish whether conditions became more or less anaerobic over time. The samples were boiled in hydrochloric acid to remove any carbon due to carbonates and then washed with distilled water. The resultant sediments were then tested for carbon using a Leco carbon and sulphur analyser, the results considered to represent organic carbon content. The presence of the organic fill deposits also allowed palaoenvironmental reconstruction and dating (both particular and relative) of lithological units. Approximately 10 grams of sample from the 10 – 20cm interval (from top) of the palaeochannel was submitted for radiometric dating to 'Beta Analytic' in the USA. The sample, taken by Malcolm Greenwood of Loughborough University Department of Geography, was analysed using the conventional beta-counting method rather than the AMS (Accelerator Mass Spectrometry) system, which is widely acknowledged to give a greater reading sensitivity (Ramsey, 1999). For the purpose of this research, and given that the samples may have been subject to 'hard water effect' (i.e. dissolved carbonates) and the presence of extraneous carbon from coal, the conventional method provides a valuable reference level.

3. Stratigraphical Correlation of Borehole Data

3.1 Stratigraphy of the study area based on regional borehole data

The spatial distribution of the boreholes dictated where the geological sections could be drawn. The sections, are of greater value where they cross the River Trent and reveal the floodplain to both the north and the south.

In all, three major transects were drawn (refer to Fig.4) each of which will be dealt with in detail later. The first transect (A1-A2) covers a distance of approximately 10km, following a SW-NE trend, crossing the River Trent 1km to the east of Stanton by Bridge and proceeding to cross the floodplain of the River Trent and River Derwent before crossing the Derwent 1km to the north of Ambaston. Section B1-B2 crosses the River Trent 1.5km downstream of section A-A1 across a relatively narrow stretch of floodplain approximately 1km NNE of Kings Newton. This section trends SSE-NNW and covers a stretch of only 400m. The final transect, C1-C2, lies a further 5.5km downstream, approaching the Trent-Derwent confluence. This section covers a distance of 6.5km across the floodplain, trending ESE-WNW and crossing the Trent approximately 1km to the south west of Cavendish Bridge. The borehole data from which the three transects were constructed is presented in appendix I.

It is clear, from the location of the 3 transects that there is a predominance of borehole information in the lower reaches of the Middle Trent floodplain, i.e. more towards the River Derwent than the River Dove.

In addition to the information gained from the three major transects, there are also particular areas of the floodplain which, although having limited borehole data, do provide information on special features (such as deep channels or the presence of organics).

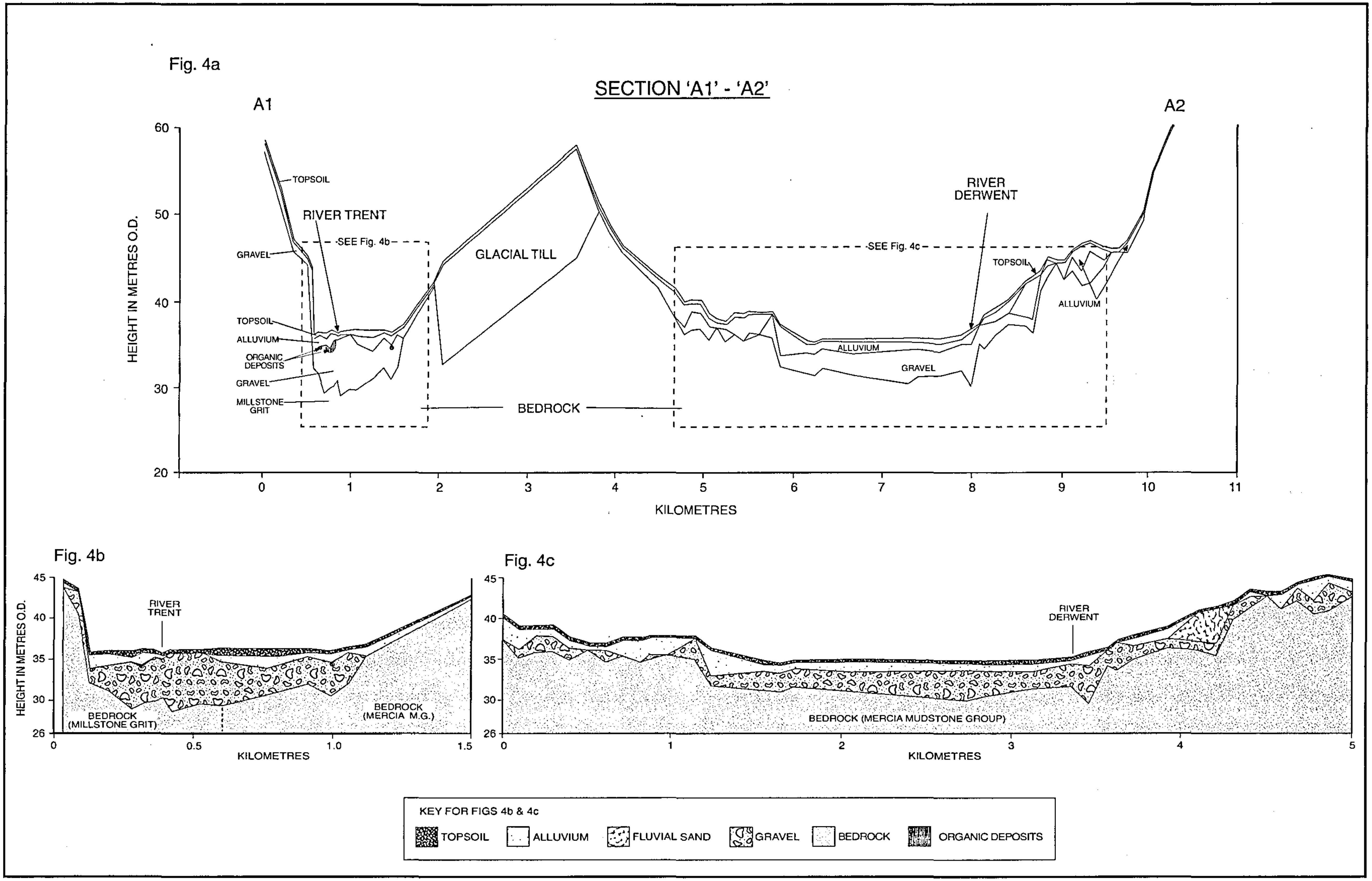

Fig. 6 Section A1 - A2

Of particular interest to this research is the borehole data relating to the main study site at Barrow-upon-Trent. This obviously requires special attention and will be considered later.

3.1.1 Section A1-A2

Section A1-A2 (Fig.6a,b,c) is the longest of the transects, following a line which crosses the floodplains of both the Trent and the Derwent. This provides a useful comparison between the relatively restricted floodplain of the Trent near Stanton-by-Bridge and the extensive floodplain of the River Derwent as it approaches its confluence with the River Trent.

The River Trent to the east of Stanton-by-Bridge is bounded to the south by a very narrow floodplain width. A rise from the floodplain level of 36m O.D. to 43.5m O.D. over a distance of less than 40m clearly illustrates this boundary. The total floodplain width of the Trent at this location is approximately 1.5km and merges to the north into a hinterland comprising of Boulder Clay overlying Mercia Mudstone. Moving in a north easterly direction, the level drops from 57.54m O.D. at its highest Till capped peak back to the Derwent floodplain at around 36m O.D. The relatively wide floodplain area (~ 4.5m) close to the confluence of the Derwent with the Trent then rises more gently to the north into Mercia Mudstone bedrock.

There is tremendous variation in the thickness of overbank fines and gravel sediments even in proximal boreholes. This variation illustrates the complexity of the depositional regime. The irregular nature of bedrock levels, whilst perhaps a reflection of zones of weakness and competence within the strata, may indicate channel incision.

Topsoil thickness remains relatively constant over the floodplain (approx. 0.3-0.5m), however, alluvium cover, whilst generally between 1-2m thick, (occasionally up to 3m) in many places is absent altogether. Similarly gravel thickness varies enormously. The borehole information reveals a marked difference between the floodplain deposits of the River Trent and the River Derwent along this line of section. The sand and gravel thicknesses on the extensive Derwent floodplain tend to be in the region 2-4m

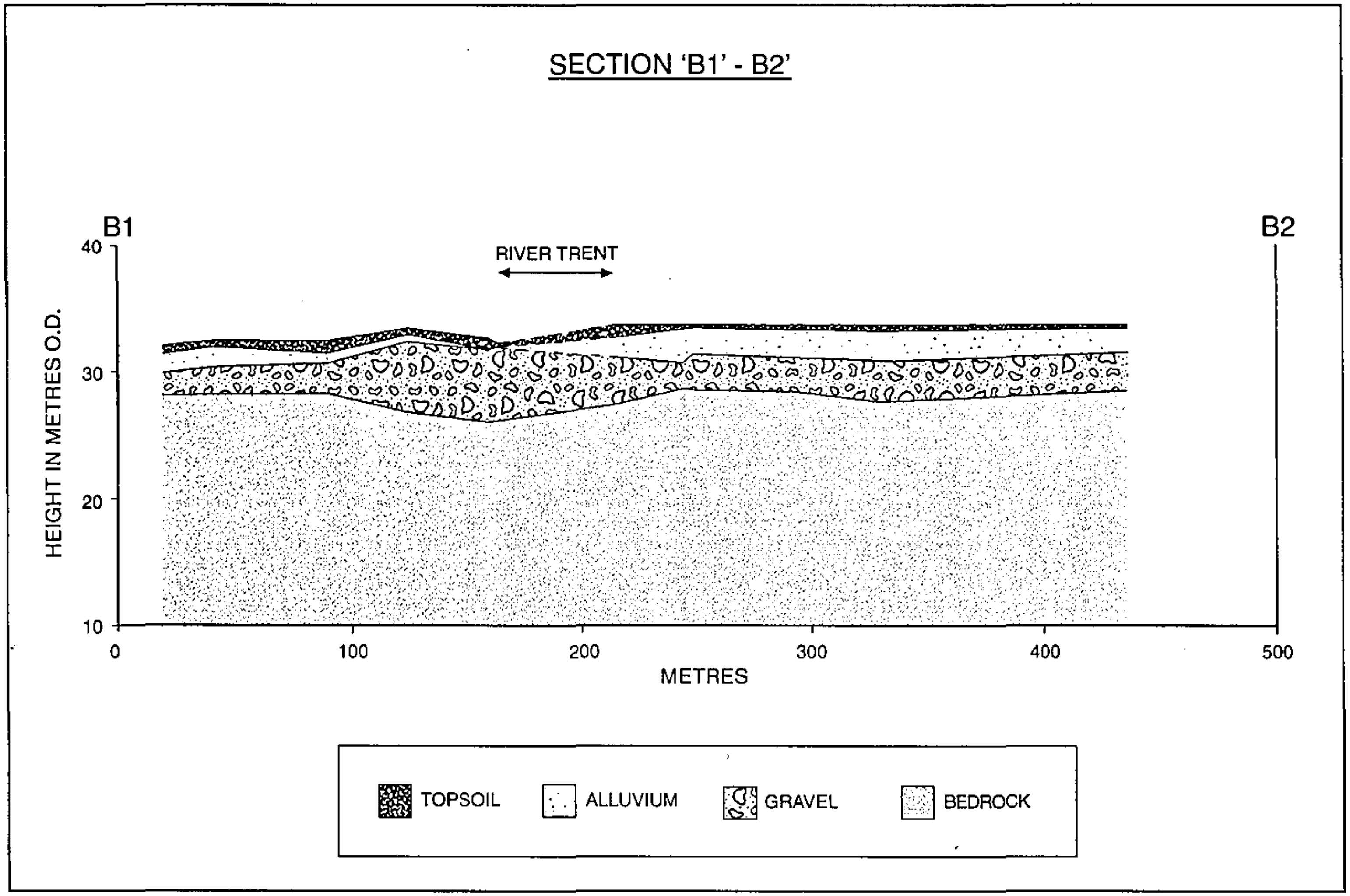

Fig. 7 Section B1 - B2

(occasionally thicker in some channels) whilst the thicknesses over the floodplain area of the Trent around Stanton-by-Bridge are typically in the region of 3-6m. This might be expected considering the relatively restricted floodplain at this point. In places, on both floodplains, there is distinct incision into the bedrock below.

The section also indicates the presence of higher elevated gravel horizons particularly noticeable flanking the Derwent floodplain. On the northern flank it appears that two distinct gravel elevations, or terraces, are present, one starting at a level of 36.5m O.D. above bedrock and the higher starting at approximately 41.5m O.D. above bedrock. On the southern flank only the lower terrace seems present. It is not clear from the boreholes, however, if there is a morphological distinction between the gravel terraces.

Occasionally the presence of clays and organic deposits is noted towards the top of the gravel particularly recorded in proximity to the existing channel of the River Trent around Stanton-by-Bridge. These deposits are upto 2m thick in places and are a clear indication of much reduced activity in the river system at the time of deposition.

3.1.2 Section B1-B2

Section B1-B2 (Fig.7) follows a line of boreholes which were drilled sometime ago for a potential motorway crossing. The section covers a distance of approximately 400m, crossing the River Trent 1km to the NNE of Kings Newton. At this point the Trent floodplain is at its narrowest at less than 1km across. North of the River Trent the floodplain is only 180m wide before being sharply bounded by a higher gravel terrace overlying Mercia Mudstone. From a floodplain elevation of approximately 34m O.D. at this point, there is a sharp rise of over 5.5m within a distance of 30m or so.

There is a clear difference in floodplain elevation to the north and south of the river. To the north the floodplain rises from an elevation of 34.6m O.D. to 35.6m O.D. within 35m of the river before dropping back again just as quickly and then gently reducing to a level of 33.9m O.D. before the floodplain boundary. South of the river the floodplain occupies a constant elevation of 35.7m O.D..

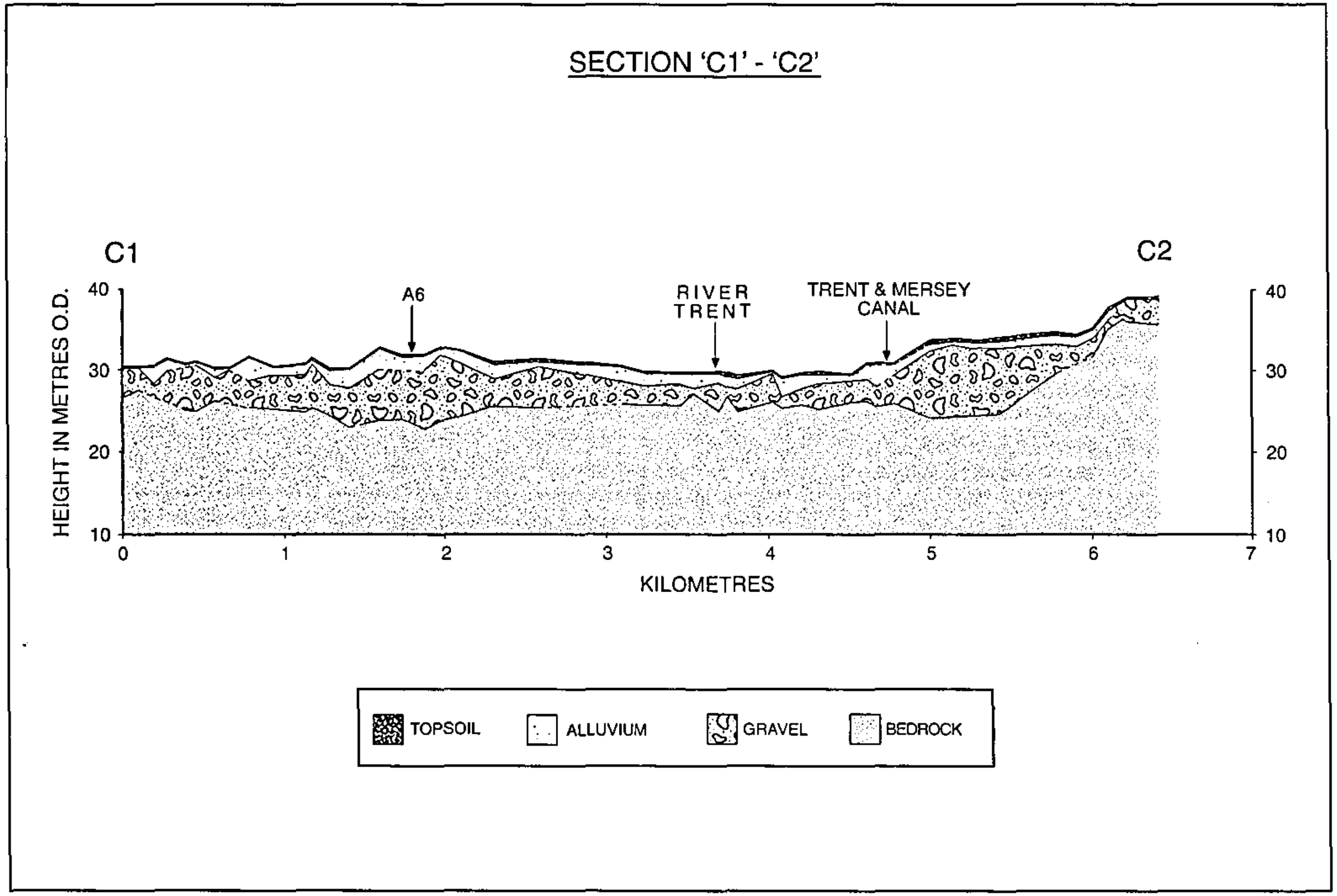

Fig. 8 Section C1 - C2

Fine-grained alluvial deposits show greater development to the south of the River Trent – thicknesses of 2.5m seem common - whilst to the north the alluvium appears wedge-shaped, starting from less than 2m furthest away from the river and becoming non-existent, 20-40m before the river bank.

Gravel thicknesses south of the river are also constant at a level of approximately 3.5m. There is a distinct increase in thickness proximal to the existing river channel as would perhaps be expected, however, it is interesting that the maximum depth of sand and gravel (approx. 6.5m) is not directly beneath the existing channel but is skewed slightly to the north. This may indicate possible channel migration over time.

3.1.3 Section C1-C2

Section C1-C2 (Fig.8) covers a distance of over 6km, almost all of which is across the Trent floodplain. The transect trends ESE-WNW from north of Lockington to north of Alston-upon-Trent and crosses the River Trent approximately 2.5km upstream of its confluence with the River Derwent. A number of features are worth comment, particularly regarding the enormous variation in floodplain elevation and thickness of deposits.

Apart from the irregular nature of the floodplain surface to the south and east of the River Trent, the general elevation of this floodplain area is much lower than to the north and west of the river, and much lower than is seen in either of the other transects. Along the line of section the mean elevation to the north and west is approximately 30.0m O.D. proximal to the river, however after a distance of 1km this rises to around 34m O.D. To the east, mean elevation is approximately 30.5m O.D. for at least 3.5km. This may be evidence of the reduction in the long profile of the River Trent, however it seems clear from the borehole data that this reduction is sudden and not gradual as might be expected.

Alluvium thickness, although highly variable, is generally between 1-2m. Occasionally this increases to 3m whilst in other areas is virtually non-existent. Similarly the gravel horizon varies between 0.8m (BH221) and 8.3m (BH148). It is interesting that the lowest gravel depths lie immediately beneath the existing river channel of the Trent. This is in direct contrast to the information from sections A and B. The greatest gravel depths can be found flanking the existing channel on both sides of the Trent at a distance of approximately 1.5km. How significant this may be is not clear. As the section progresses westwards it is again possible to identify a higher gravel terrace moving into a clay/till deposit.

3.1.4 Other Features of borehole data

In addition to the information gained from the three major sections crossing the River Trent, there are also other features present and logged in other boreholes that warrant special mention:

a) Deep channels/pockets

In locations where borehole information is spatially distributed over an area, rather than in a linear fashion, it may be possible to identify features that exist on a three-dimensional scale. This is especially apparent in the main study area, but also seems to be the case on the floodplain north of the river around Willington power station. A number of boreholes suggest relatively deep gravel deposits (6-7m) rising gradually to shallower levels on both sides. The boreholes also suggest the presence of isolated deep gravel pockets within the floodplain. These features may represent braided-river scour holes

b) Cobbles

The presence of large cobbles seem to be associated with increased gravel depths. A number of boreholes record the presence of large stones or cobbles towards the base, especially in the deeper gravels. The presence of these larger stones are important and prove invaluable (e.g. Maizels, 1991) when it comes to reconstructing the hydraulic regime at the time of deposition i.e. greater size equals greater stream power.

c) Organic silts/clays

A number of boreholes record the existence of dark organic silts, clays or peats within the strata. In some cases the presence of these deposits is recorded very close to the existing river channel (as in section A-A1) whilst in other boreholes they are more distal. These deposits are also found at various depths within the sequence, whether it being the top, the base or within the sands and gravels. The presence of these deposits is important since they represent periods of relative inactivity (low energy) in the river system. This is in addition to any information that might be provided from the organic content of the material i.e. coleopteran remains, macro/microfossil data.

d) Molluscs

Very occasionally, reference is made in the borehole logs to the presence of mollusc shells within the strata - one such borehole lies approximately 1km north of the existing Trent channel and 500m east of Twyford. Because of their fragile nature these tend to be associated with the presence of fine-grained deposits (silts/clays etc). Unfortunately, very little information is often provided regarding the type of shell found, although the presence of molluscs themselves do reveal valuable information about the alluvial regime.

3.2 Summary of regional borehole data

The analysis of the borehole evidence has clearly provided valuable information on the floodplain of the Trent between its confluences with the River Dove and River Derwent.

There is tremendous variation in alluvial and gravel thicknesses, even in very close proximities, and the presence of deep channels and pockets within the strata emphasize the enormous variability that exists, and has existed within the river system. The presence of glacial deposits, organics, molluscs and extreme grainsize variation and grading (i.e. coarsening to cobbles at base) add to this palaeohydrological variation.

Consistently low alluvial thicknesses on wide floodplain areas is somewhat unexpected and there does not seem to be any direct correlation between alluvium cover and gravel thickness. Nor does any clear relationship exist between thicknesses of various strata and proximity to the existing river channel. It is perhaps generally the case that gravel thickness does increase towards the river but often this increased thickness is not directly under the contemporary channel but skewed to one side, perhaps indicative of channel migration.

3.3 Statigraphy of Barrow-upon-Trent based on borehole data

Analysis of aerial photographs reveal the presence of possible palaeochannels cutting through the southern half of the site (refer to Fig.4). Unfortunately, the topographical data does not provide evidence of increased or decreased elevations matching any such channel development although this is not particularly unusual. There are areas of raised elevation within the site, however these 'islands' seem to be fairly random in nature. Floodplain elevation shows a distinct decrease in a downstream direction from approximately 39m O.D. to just over 37m O.D. eastwards as would be expected from long profile.

The alluvium cover over the site is relatively thin given the area of open floodplain. Thickness of alluvium range between 0.1m - 1.9m, increasing towards the existing river channel. Gravel thicknesses, when compared with other areas of the Trent floodplain, tend to be consistently high over the whole site and show a marked increase in the southern section. Thickness of 6m+ are not uncommon. Even deeper gravel channels and pockets can be identified running through the site, and the highest recorded gravel depths of 11.5m are present in the south western corner close to the existing river channel.

Other features, such as the presence of large stones, coal/charcoal fragments and organic silts are recorded in boreholes throughout the site.

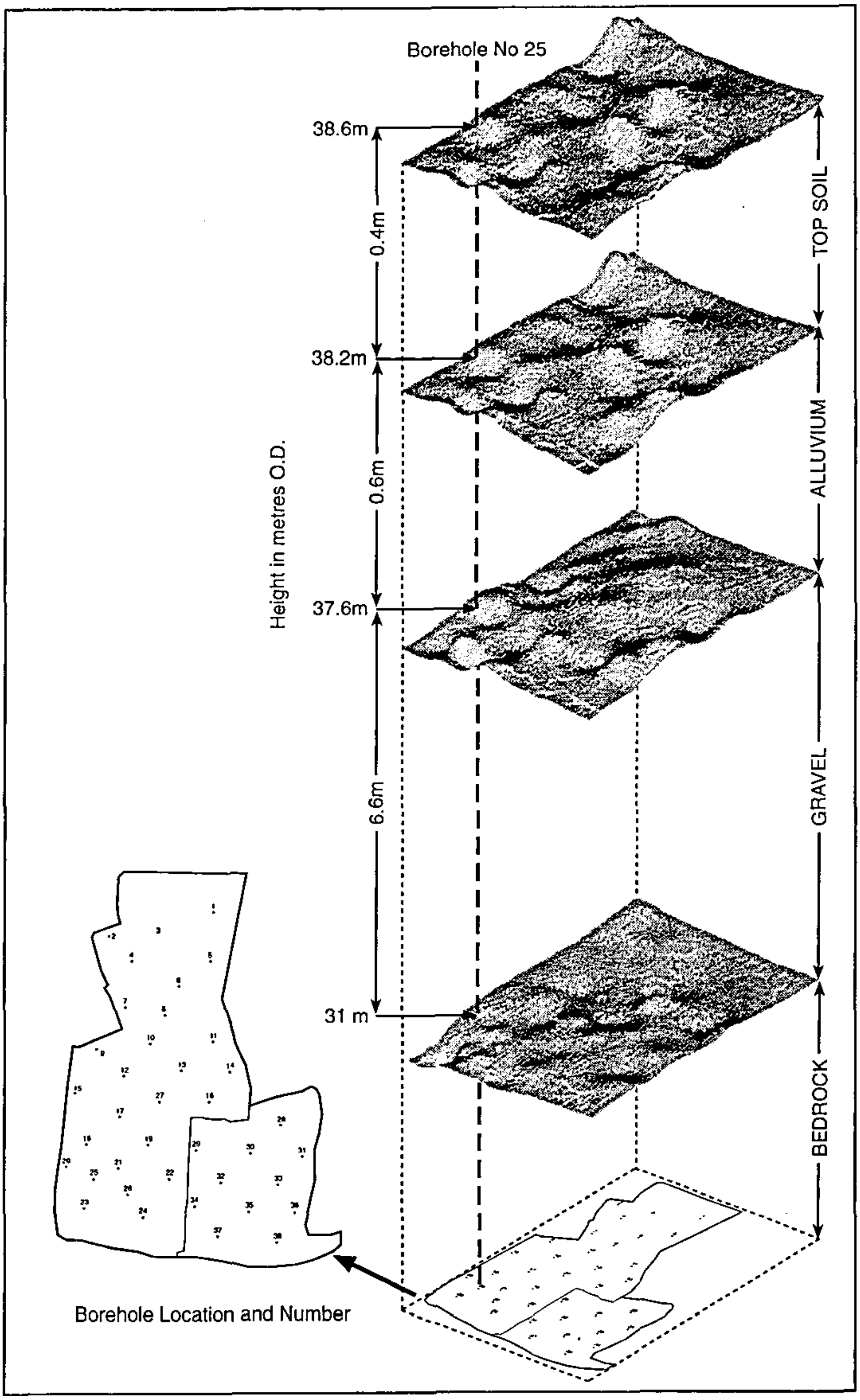

Fig. 9 Three-dimensional profiles

3.3.1 Three-Dimensional Profiles

Four profiles were generated, these representing the current surface level across the site, the topsoil/alluvium interface, the alluvium/gravel boundary and the bedrock contour. These images can be seen with reference to figure 9. The image created for each unit shows a degree of vertical exaggeration in order to illustrate and exacerbate any variation present within the strata and highlight the presence of high/low points and channel alignment.

The profiles should be considered in conjunction with each other to be of most value though it is clear even from a cursory glance that a distinct alignment exists in all the deposits showing an approximate South West - North East trend. This is interesting when a comparison is made with earlier aerial photographical analysis which clearly shows palaeochannel scars/haloes progressing across the southern half of the site in a very similar orientation. Each profile is considered briefly below.

1. Surface Topography

The profiles indicate a fairly variable surface topography with noticeable high points in the central area of the site and, most markedly in the far northwestern corner. A distinct 'channel' alignment exists in a SW - NE trend which is particularly pronounced in the southern section.

2. Topsoil/Alluvium Interface

This interface is almost a mirror-image of the surface topography which is more or less to be expected given that topsoil thickness is approximately 0.3m across the site and shows very little variation.

3. Gravel Surface

The upper gravel surface is distinctly different to the surface and alluvium profiles and, whilst still showing depressional linear features, these are now much more pronounced.

There is a greater preponderance of high points in the sands and gravels towards the southern half of the site. Comparatively lower contours are found at the southwestern edge of the site and this, when compared with the alluvium surface profile, indicates greater thicknesses of alluvium in this vicinity.

4. <u>Bedrock Contour</u>

The bedrock contour is the least variable of all of the profiles although some channel alignment is clearly visible. The most noticeable feature of this contour is the marked low point or 'pocket' at the southwestern edge. When compared with the upper gravel contour this indicates a much increased thickness in the sands and gravels at this point. Channel development seems almost exclusively restricted to the southern half of the site.

Whilst borehole analysis is an extremely valuable tool, it can only provide a somewhat limited picture in palaeohydrological reconstruction. For a more complete investigation it is necessary to examine the deposits at first hand.

4. Field Investigation

The field investigation at Barrow-upon-Trent was carried out at the Redland (since 'Lafarge') Aggregates sand and gravel pit (G.R. SK32NW 344284) and focused on the working aggregate faces as quarrying operations progressed. Figure 10 illustrates the position of these faces over a period of time and gives the location of the sedimentary logs referred to later. Representative sections were drawn where appropriate allowing the recognition and analysis of sedimentary facies and their change (vertical and lateral) over distance. Although the textural and structural monotony of many gravel deposits make the recognition and correlation of bounding surfaces more difficult (re: Smith, 1990) the sedimentary units referred to are laterally extensive both locally and over the site as a whole.

Typically, the stratigraphy at Barrow-upon-Trent consists of medium to coarse-grained sand and gravel deposits, overlying Mercia Mudstone bedrock. These coarser sediments often contain minor sub-horizontal planar sand horizons and occasional sand lenses that exhibit cross-bedding (allowing palaeocurrent determination).

The presence of organic deposits is noted at a number of localities and the discovery of a substantial organic palaeochannel (location 1, Grid Ref:34394,28264) is of particular relevance on both a palaeoenvironmental and dating perspective. This will be discussed in a separate section to follow.

Alluvial sand and gravel depths are typically in the range 3.0-4.0m although depths of upto 5.4m (and as thin as 2.0m) have been recorded. The terrace deposits are overlain by approximately 1.0-1.5m of clayey/silty alluvium containing occasional coarser fragments.

Classification of the deposits into particular terrace formations i.e. 'Hemington Terrace deposits', 'Holme Pierrepont Sand and Gravel' is attempted in chapter 6 and will be discussed as relevant.

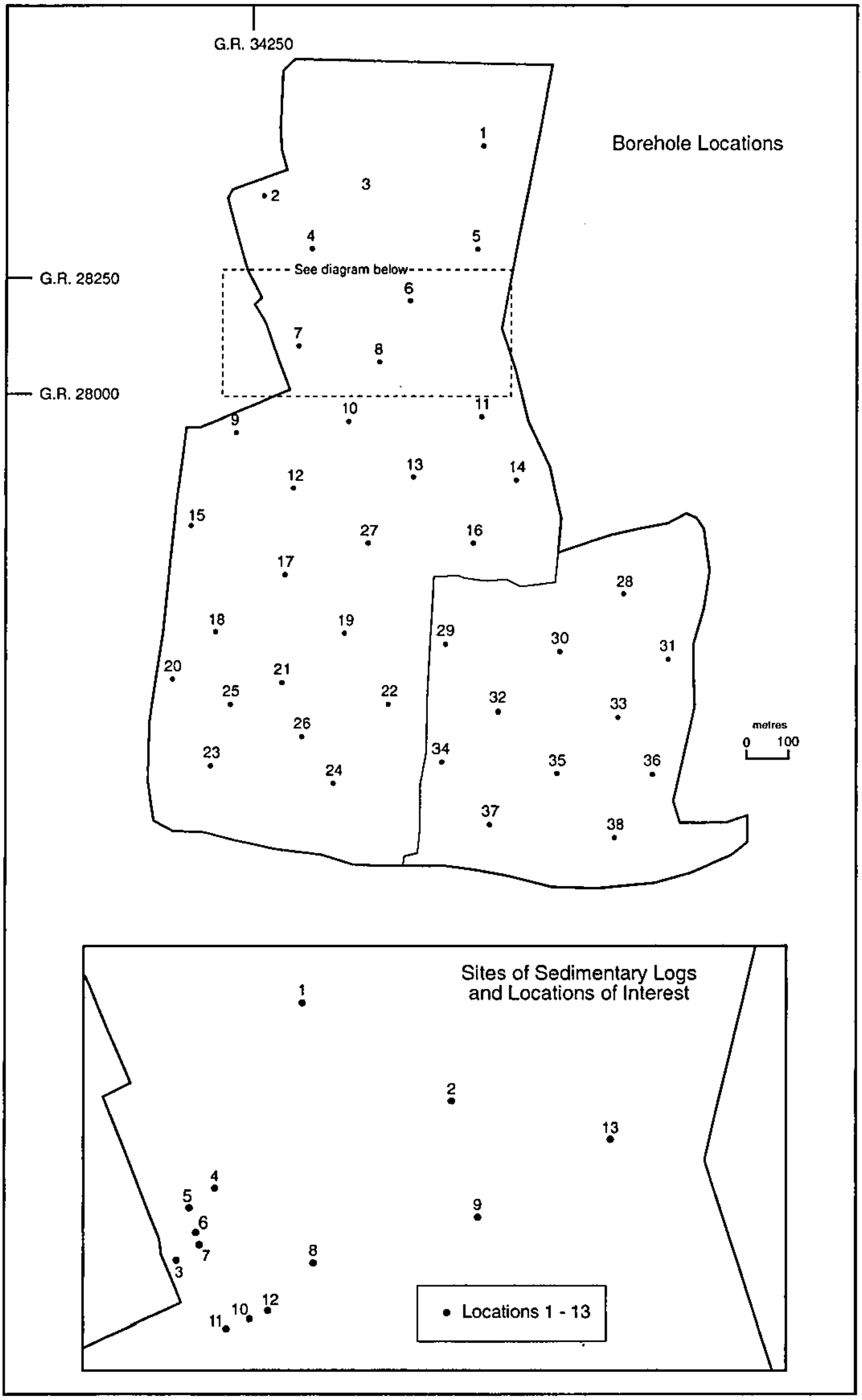

Fig. 10 Location of boreholes and sedimentary logs

Lithologically the gravels are dominated by quartz and quartzites, and minor components that include sandstone/siltstone, resistant acid and basic igneous lithologies, cherts, flints and limestones. Table 2 details the clast composition of the pebble grade from different levels in the gravel substrate (low, medium, high). The majority of the clasts are well rounded, well shaped (34.3% equant/spherical, 31.4% oblate/discoidal, 25.7% prolate/roller, 8.6% triaxial/bladed; based on Zingg,1935) smooth textured and generally testify to the development of considerable maturity within the deposits. The association of clast compositions (particularly the dominant types) suggest that the gravels are most likely derived, in the main, from upstream Bunter deposits (refer to Fig. 2). Forming part of the Sherwood Sandstone Group the Triassic 'Bunter' pebble beds consist of well –rounded, conglomeritic quartzites with some carboniferous limestones, cherts and igneous lithologies. It is always dangerous to make assumptions on clast inheritance or provenance, particularly in a glacial outwash/alluvial context which involves the continual reworking of deposits, however it is proposed that the gravel sediments are most likely from this source. The ultimate origin, or provenance, of the lithologies is a separate issue altogether since the Bunter deposits are themselves a result of the simultaneous deposition of sands and gravels from a previous inheritance (Steel and Thompson, 1983).

	Quartz	Quartzite	Sandstone/ Siltstone	Iron-stone	Chert	Flint	Lime-stone	Chalk	Acid Igneous	Basic Igneous	Blackened Quartz	Clast Count
High Level Gravels	33.3	38.3	14.0	0	1.6	2.1	0	0	3.7	6.6	0.4	243
Medim Level Gravels	32.0	39.3	11.1	0	2.5	0.8	0.4	0.8	5.3	7.0	0.8	244
Low Level Gravels	26.4	39.1	14.1	0.4	2.5	1.1	2.1	2.5	4.6	4.9	2.1	284

Table 2 Lithological Analysis (%)

Six major lithofacies are distinguished (A-F), each representing a distinct location in the depositional environment. Sub-facies have not been so formally designated but an important range of variation does exist. An outline description of the lithofacies and their relative frequency is presented in Table 3.

Lithofacies	Outline Description	Relative Frequency
A	Clast-supported, medium-coarse grade gravel with medium coarse interstitial sand (<20%). Gravels are largely poorly sorted and are massive or exhibit crudely sub-horizontal bedding.	V.common
B	Moderate-well sorted, clast-supported gravel. Largely medium-fine grade gravel with medium-coarse interstitial sand (20-30%). Crude sub-horizontal bedding may or may not be apparent.	Common
C	Matrix-supported, unsorted medium-fine grade gravel with medium-coarse interstitial sand (>30%). No observable structures or bedding.	Less common
D	Horizontally bedded medium-coarse grade sand, laminated to massive. Facies thickness usually less than 30cm.	Common
E	Cross-bedded sand (Planar), medium-coarse grade. Facies thickness usually less than 30cm.	Less common
F	Clays and silts, mainly organic, occurring as thin drape deposits or distinct palaeochannels. Rootlets and desiccation cracks common	Uncommon

Table 3 Lithofacies

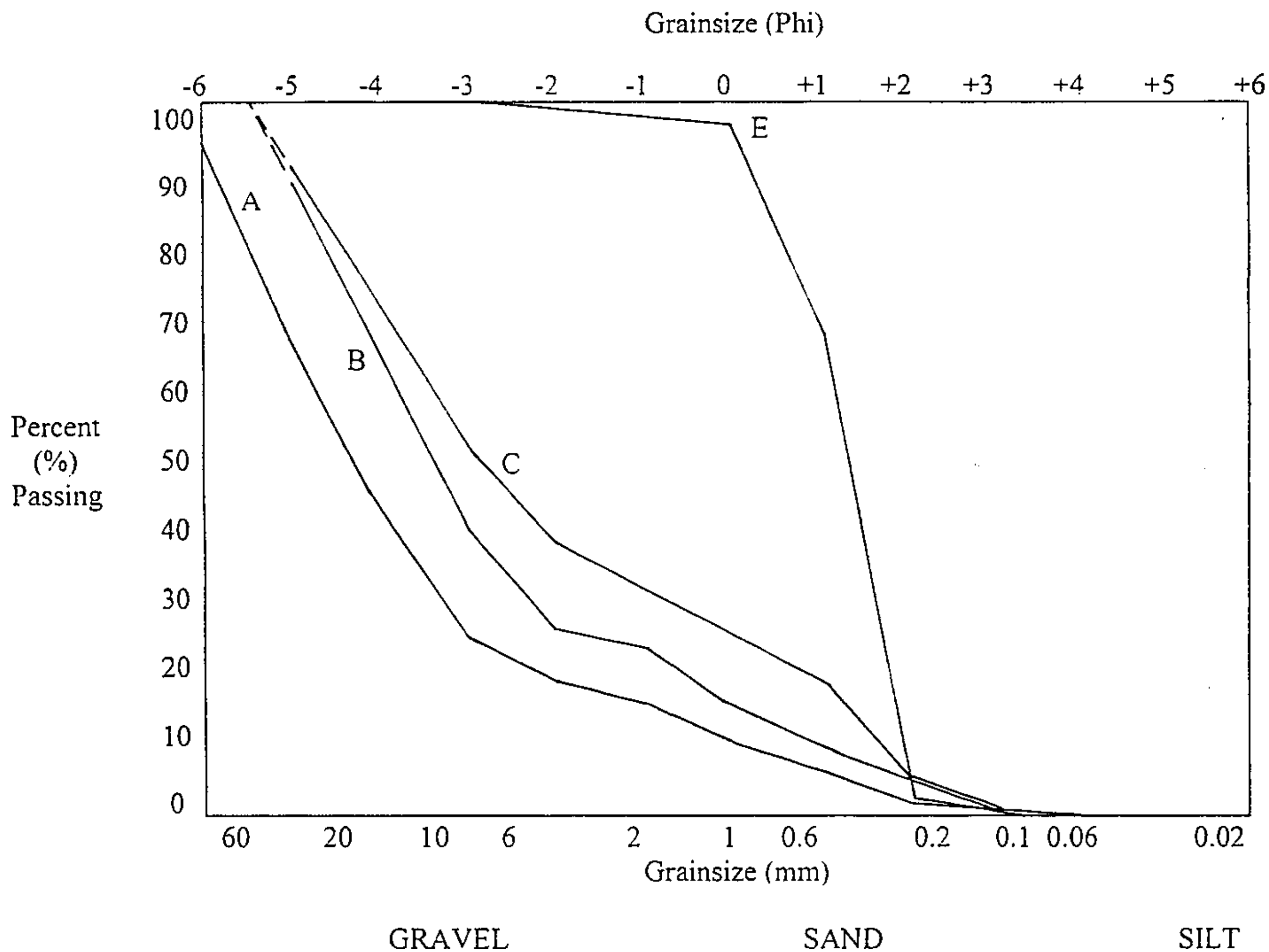

Fig.11 Grading Curves for Lithofacies A, B, C & E (semi-log plot)

4.1.1 <u>Lithofacies A</u> (Plate 1)

This is the most common type of lithofacies at Barrow-upon-Trent and is comparable to facies Gcm/Gh as proposed by Miall (facies classification, 1978c). The deposits are clast supported and typically consist of medium-coarse grade gravels with a medium-coarse interstitial sand content of less than 20% (Fig.11, Appendix II). The clast framework suggests that the major gravel deposition occurred prior to the influx of the finer (sand-silt) matrix. The gravels are poorly sorted and exhibit massive or basic sub-horizontal bedding, perhaps indicative of successive bar core sedimentation. The gravel is polymodal, clasts are well rounded and show no marked orientation. Individual gravel clasts are typically in the range 2-3cm although clast sizes up to 7cm diameter (∅) are common and clasts in excess of 20cm∅ are recorded in several logs. The lithofacies may also contain basal lag deposits (as recorded at location 1) where larger clast sizes (up to 20cm∅) are common. Relative palaeodischarge estimates based on clast size will feature later in the chapter, however, it is clear that overall sediment grading suggests high energy fluvial conditions at the time of aggradation.

Units generally occur as horizons less than 1m thick and have a lateral extent of over 100m in some cases. Contact with other lithofacies tends to be highly variable and ranges from distinctly sharp (particularly with sand-silt horizons) to gradational boundaries which may or may not be undulating. The nature of the contact between lithofacies is clearly indicative of differing erosional regimes which will be discussed later.

4.1.2 <u>Lithofacies B</u> (Plate 1)

Lithofacies B is a moderate to well sorted, clast-supported gravel which can be described as showing a greater degree of 'organisation' (after Steel and Thompson, 1983) than Lithofacies A. Deposits consist largely of medium grade gravels with a medium-coarse interstitial sand content of between 20-30% (Fig.11, Appendix II). Clast size is typically in the 1-2cm∅ range (with occasional larger pebbles up to 8cm∅)and there may be a further distinction into different common gradings i.e. groups averaging 2cm∅, 1cm∅, 0.5cm∅, etc. Occasionally the clast matrix is absent, perhaps indicating the winnowing of fines

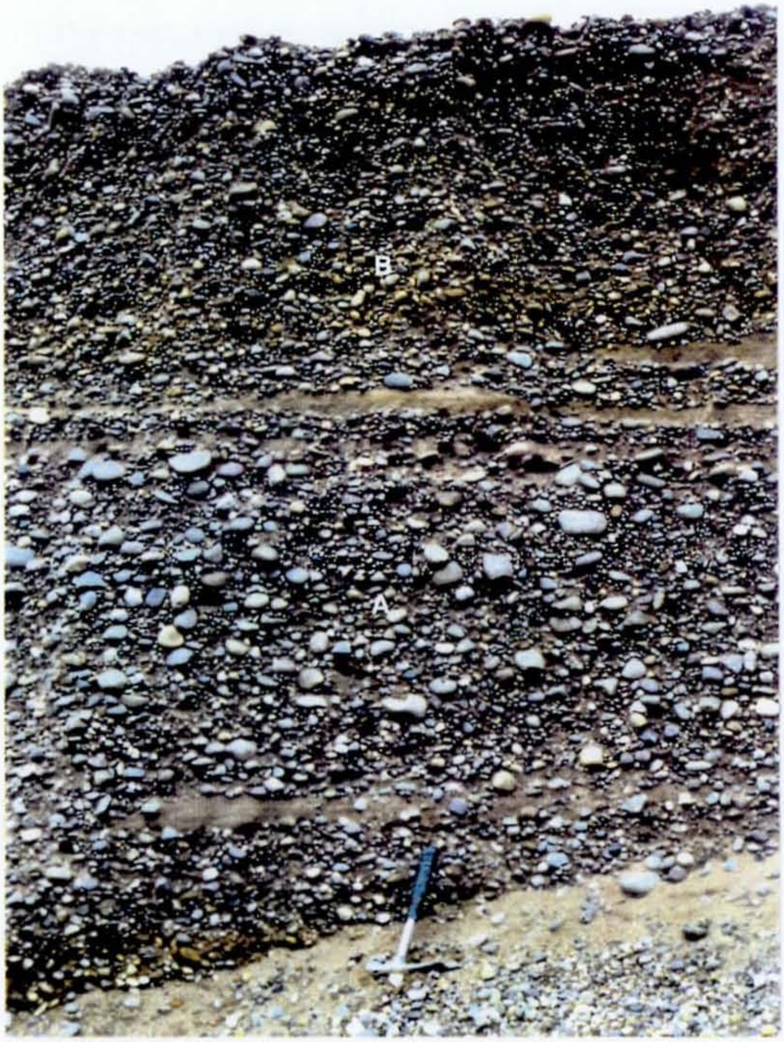

Plate 1
'A' – Lithofacies A, clast-supported gravel (poorly sorted)
'B' – Lithofacies B, clast-supported gravel (moderately sorted)
(hammer measures 30cm)

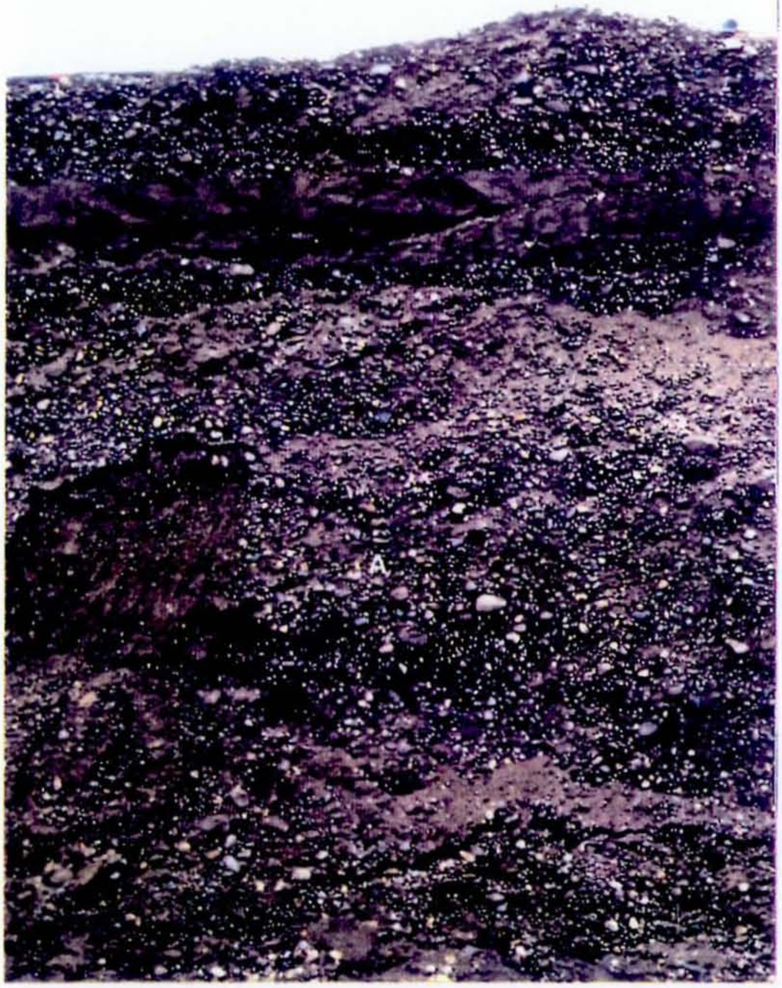

Plate 2
'A' – Lithofacies C, matrix-supported gravel
(face is 3m high)

during discharge decrease, and clasts tend to be well-rounded, occasionally showing alignment (allowing for possible palaeocurrent determination).

As with Lithofacies A, crude sub-horizontal stratification may or may not be apparent although the individual units do not appear as laterally extensive as the former. Unit thickness is typically less than 1m and horizons tend to alternate with units from Lithofacies A in many sequences, occasionally forming the only lithofacies present. Contact with other facies is variable except with Lithofacies A when the contact is almost always gradational. It seems likely that the more organised deposits of Lithofacies B were formed at times of more continuous flow than the 'disorganised' former.

4.1.3 <u>Lithofacies C</u> (Plate 2)

Lithofacies C is a disorganised matrix-supported deposit which is less common than Lithofacies A or B. Grading is that of an unsorted, medium-fine gravel with medium-coarse interstitial sand content in excess of 30% (Fig.11, Appendix II). Clasts are sub-rounded and usually small (less than 1cm ∅) but may include some larger pebbles upto 8cm in diameter. No observable clast orientation, structures or bedding is apparent and individual units rarely approach substantial vertical or lateral extent (thickness < 1m). Occurring more usually at the top of a sedimentary sequence, Lithofacies C commonly has a gradational contact with other gravel lithofacies types.

The nature of Lithofacies C would certainly suggest that deposition of the sediments occurred both rapidly and simultaneously (Steel and Thompson, 1983).

4.1.4 <u>Lithofacies D</u> (Plate 3)

Lithofacies D is a horizontal, or sub-horizontal, bedded sand facies that may appear laminated or massive. The sand is largely medium grained and may contain small, included pebble or coaly fragments along bedding planes. The unit may take the form of a thin lense, measuring between 3cm and 30cm in thickness and upto 10m in lateral extent, or may appear as a very thin (usually less than 5cm) sheet-like feature many metres across. This lithofacies tends to have a very sharp contact with other horizons although the

Plate 6

'A' – Lithofacies F (deposit is 0.5m thick)
'1' – Sand lenses at base of deposit
'2' – Orange (iron) stained band
'3' – Lithofacies B

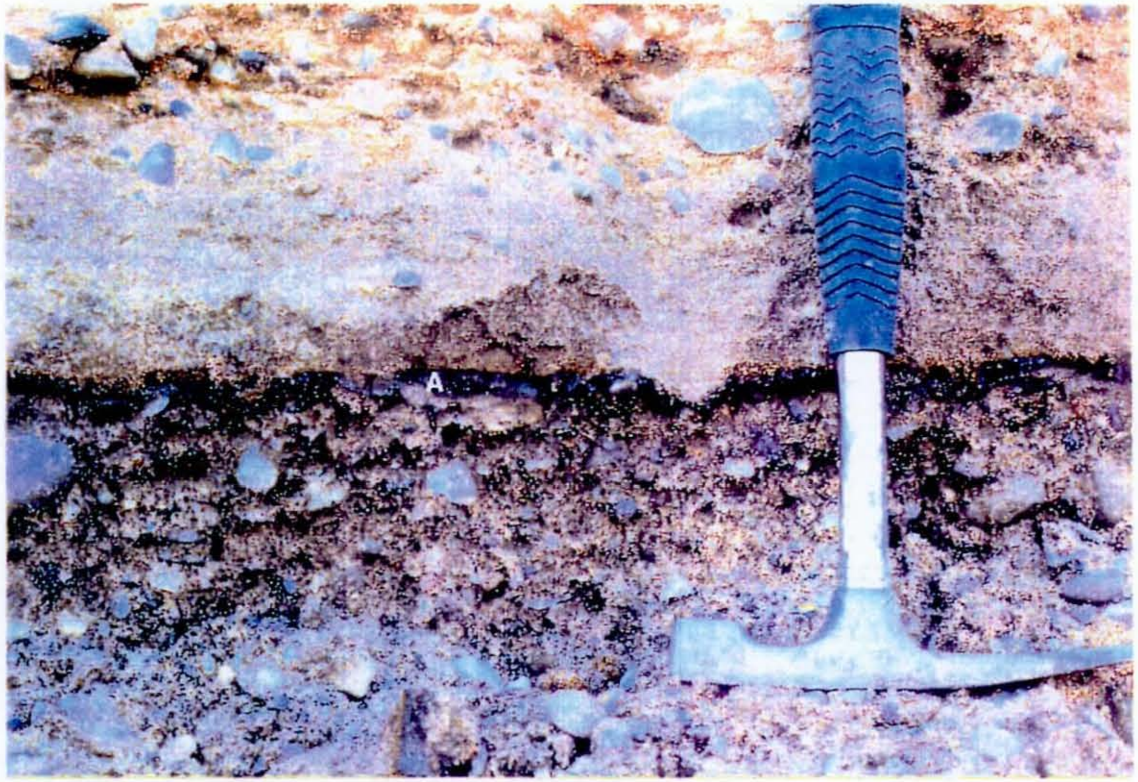

Plate 7

'A' – Lithofacies F, clay banding

Plate 8
'A' – Clay involutions
'1' – Orange medium sand with pebble inclusions
'2' – Lithofacies B, small clasts

(organic) and contain small occluded pebbles. Interbedding or grading of fine and coarse (upto medium sand) sediments may occur (Plate 6) and the deposits may also contain shell fragment and rootlets. Desiccation cracks can also be present.

Lithofacies F may occur as thin lenses or drapes a few millimetres to a few centimetres thick (Plate 7) and have a lateral extent of over 20m (location 3). Contact with other lithofacies is usually distinctly sharp and may show signs of post depositional periglacial disturbance (Plate 8). The deposit also occurs podlike or indeed as distinct palaeochannels (Plate 9) containing organic debris. At location 1 such a palaeochannel was present and this provided valuable information on the depositional environment and dating of the deposits. This is discussed in a separate section to follow.

4.2 Organic Palaeochannel (Plate 9)

At location 1 (refer to Fig.10) the presence of a large palaeochannel was recorded. The following section examines the stratigraphy and sedimentology of the deposit and gives special reference to the contained organic debris that provide age and environmental information.

4.2.1 Stratigraphy and Sedimentology

The palaeochannel was located at an exposed section (Grid Reference 3439 2826) sandwiched between gravel horizons. Much of the right hand side of the channel was obscured by a gravel bund, however, the overall dimensions were clearly recognisable. Measuring 21.6 metres across with a maximum thickness of 0.75 metres the channel was configured in the form of a shallow dish with a sharp sub-horizontal upper boundary. The palaeochannel occupied an elevation within the gravels of 36.98m O.D. at the upper surface mid-point. To the north-west of this channel at another exposed section (Grid Reference 34325, 28311) a smaller dish-shaped organic deposit was recorded at an elevation of 36.87m O.D. (upper surface mid-point) and measuring 6.0m x 0.5m. Information provided by excavation personnel suggested that this was part of the same deposit on the opposite quarry face i.e. the deposit was followed during the excavation of the void.

Plate 9
'A' – Lithofacies F, organic palaeochannel
(note orange stained layer at base of deposit)

At the maximum thickness, the palaeochannel consisted of a black/dark brown clay and silt infill with sandy lenses or bands contained within. These bands, appearing pale grey or brown in colour, were concentrated towards the base of the channel (refer to Plate 8). Occasional pebbles and small shell fragments were present within the infill. The maximum clast size encountered was 8cm in diameter although the vast majority were less than 1.5cm in diameter. The base of the palaeochannel was distinct though slightly undulating in places and approximately 3 metres of pinkish gravels lay beneath. At the very top of these clast supported gravels and immediately underlying the palaeochannel, the gravels were stained orange over a 6cm interval following the curvature of the channel. The palaeochannel was overlain directly by 0.7m of clast supported (orange) gravels and then 1.0m of matrix-supported gravels (increasingly sandier towards the top) and 1.3m of alluvium. The sub-horizontal contact between the upper surface of the palaeochannel and the gravels was distinctly sharp and level.

4.2.2 Grainsize Analysis and Carbon Content

Fine-grained sediment and carbon analysis was carried out on the palaeochannel over 10cm intervals, as detailed in the methodology, the results of which can be seen with reference to Table 4.

Sample Interval (from top)	Sample Weight (grams)	% Clay	% Silt	% > Silt	% Organic Carbon Content
0 – 10cm	16.4271	14.38	53.40	32.22	2.28
10 – 20cm	18.0135	2.74	60.49	36.77	2.28
20 – 30cm	17.8222	5.04	50.32	44.64	2.21
30 – 40cm	18.3533	5.31	38.69	56.01	1.73
40 – 50cm	18.3365	0.93	50.61	48.46	2.10
50 – 60cm	18.8032	0.56	59.20	40.24	1.65
60 – 70cm	18.8549	0.92	43.47	55.61	1.65

Table 4 Sediment Analysis for Organic Palaeochannel

A number of observations are worth comment:

- Increasing clay content towards the top of the channel – clay content shows a marked increase above the 40-50cm horizon and reaches a maximum of 14.38% at the top of the palaeochannel. This variation indicates a progressive decrease in depositional energy over time, perhaps representing increased abandonment from the main channel or a decrease in energy of the overall riverine system. Fluctuations that do exist, such as in the 10-20cm horizon (clay content drops to 2.74%), reflect normal fluctuations in flow that exist in such an environment.

- Fluctuating silt and sand content – silt content throughout the channel depth is variable and bears little relationship with clay content. Sand content is also variable although a distinct decrease in the top 20cm of the channel is apparent. The highest sand concentration occurs in the middle of the channel (30-40cm = 56.01%) and suggests a relative increase in depositional energy at that point (perhaps a small-scale overbank flood?).

- Upward increase in organic carbon content – organic carbon content is clearly higher in the top 30cm of the palaeochannel than elsewhere. The higher concentration above 30-40cm horizon suggests that, overall, the conditions were more anaerobic in this depositional period. This conclusion is supported by the inherent faunal assemblage (refer to section 4.2.3).

4.2.3 Dating and Correlation

Samples taken from the organic palaeochannel, as part of research carried out by Mr. Malcolm Greenwood, revealed the presence of macrofossils which could be used for age and environmental correlation. The Caddis/faunal assemblage indicated a distinct change in environmental conditions over the depth of the palaeochannel. From the base of the channel to a height of 30cm the assemblage represented a more riverine system whilst above this level the assemblage pointed to a more ponded environment (re. Malcolm Greenwood). This information supports the interpretations made from carbon and grading analysis.

Plant macrofossils (i.e. stems, leaves etc.) sampled from the 10-20cm interval (from top) were submitted for radiometric dating using the conventional beta-counting method. The results indicated a radiocarbon age of 13060 ±90 (CAL yr BP) for the sediments. Although this information relates only to one point of deposition in the palaeochannel it is nonetheless valuable. As well as providing a definitive age, it provides a relative dating for the underlying/overlying deposits i.e. gravels below the palaeochannel are at least older than 13060 BP and gravels above are younger than this date. This proves an important point in the interpretation of the deposits and is discussed later.

4.3 Palaeocurrents and Palaeohydraulics

4.3.1 Palaeocurrents

Palaeocurrent directions, taken from forset dips (Lithofacies E) at five locations (locations 2, 3, 8, 10, 12 – refer to Fig. 10) and pebble long-axis orientation at one location (location 2) are compared. Two distinct alignment patterns emerge. At locations 2, 3 and 8 forset dip directions are 112°, 114° and 106° respectively with angles of repose at approximately 21° (Table 5, Appendix III). Pebble imbrication, recorded in the form of a rose diagram (Fig. 12, Appendix III) also suggests a mean palaeocurrent direction of between 105° and 120°. Although pebble imbrication can often be a function of 'rolling' along the stream bed the readings, supported by those from forset dips, do suggest alignment parrallel to flow direction. At locations 10 and 12 measurement from planar cross-bedding suggests a mean palaeocurrent direction of approximately 80°. The reason for this variation may become clear when the mechanisms of deposition are considered. It is well established that a wide dispersion of palaeocurrents is typical of meandering stream deposits whilst lower dispersions are typical of less sinuous streams (e.g. Kelling, 1968; Thompson, 1970). The individual bedforms which migrate in river channels and which give rise to cross-bedding are extremely complex in their behaviour (Collinson, 1978). Cross-bedding directions most commonly relate to different patterns of bar movement rather than channel type (Smith, 1972; Collinson, 1978) and palaeocurrents should therefore be considered with regard to the type of sedimentary structure and their position in the channel sequence. With this in context it should be noted that locations 10 and 12, recording palaeocurrent directions of 79° and 80° respectively, are further south than

Location	Dip	Direction
2	21°	112°
3	21°	114°
8	22°	106°
10	20°	79°
12	20°	80°

<u>Table 5</u> Palaeocurrent data from planar-crossbed forset dips (Lithofacies E)

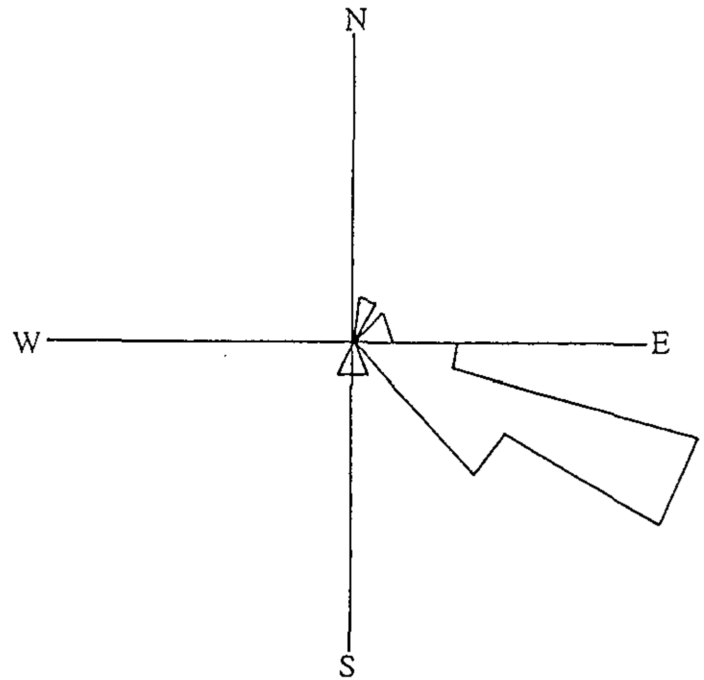

Fig.12 Palaeocurrent data from pebble orientation (long axis)

locations 2, 3 and 8 and are within 15 metres of each other in the sedimentary sequence.Locations 2, 3 and 8, recording palaeoflows to 112°, 114° and 106° respectively, are relatively spread-out over the site and may suggest a more reliable picture of overall palaeoflow direction. It should be noted that the deposits from which all the palaeocurrent data is obtained show no relationship with regard to height in the sedimentary sequence. The planar cross-bed and pebble alignment horizons vary from the base of the gravels (location 2) to the top (location 3) perhaps suggesting that the overall flow direction remained relatively constant throughout aggradation of the deposits. As a point of comparison, and discussed later, Brandon (1997) recorded a palaeocurrent direction of 140° in the base of the gravels at the same site (G.R. 3444 2835).

4.3.2 Palaeohydraulics

The methods used in palaeodischarge reconstruction are fraught with difficulties (refer to section 2.1.1) and rely on the input of data regarding former channel dimensions, such as channel width and gradient, and the measurement of mean and maximum particle size. Complications can also arise in glacial systems which are subject to the sporadic release of vast quantities of stored water (known as glacier outburst floods, or 'Jökulhlaup's) leading to a marked change in fluvial dynamics.

Channel slope was calculated at 0.00076, using the contemporary floodplain surface gradient as an approximation. Average and maximum clast size was recorded for each lithofacies type. Lithofacies A recorded maximum clasts upto 20cm in diameter, sometimes present as a basal lag, whilst gravel lithofacies B and C recorded maximum clast size upto 8cm in diameter. Unfortunately, as is common for many coarse gravel rivers, especially braided rivers (e.g. Miall, 1977) channel width could not be established. This is because distinct channel banks are often absent in the complex (and mobile) bar and channel system of a braided reach.

It is theoretically possible to determine shear stress and probable flow depths using clast size and gradient data. However, given that the range of error in calculating palaeoflow is substantial, even when many of the variables are known (i.e. methods used by Maizels, 1991, suggest errors of between –80 and +350%), it seems clear that any proposed estimates using these methods would be subject to criticism. Additionally, it should be emphasised that floodplain deposits, such as those of the Trent, represent deposition over a considerable period and it is perhaps oversimplistic to regard localised sediments within the alluvial sequence as representative of the overall hydraulic conditions.

To illustrate the inherent complexity of deposition within a system, such as that which prevailed within the Trent Basin, it is possible to consider the issue of catastrophic burst events. These flood events are triggered by a number of mechanisms although the most common glacier outburst floods occur as a result of the sudden drainage of an ice-dammed lake situated behind, or below the ice dam (glacier/ice-sheet). Commonly, as in the case of the Trent (refer to section 6.1.1) ice damming may lead to temporary diversion of a river course until the ice dam is breached, usually as a result of ice retreat. Apart from the major glacial readvancement that occurred during the Younger Dryas (Benn and Evans, 1998) the retreat of the Devensian ice sheet across Cheshire and Shropshire, for example, is suggested to have been complex 'involving halts and perhaps minor readvances' (Jones and Keen, 1993). It is probable, therefore, that the Trent Basin experienced a number of glacier burst episodes which are represented, and preserved with the deposits.

Accepting the above, it may be possible to reconstruct palaeodischarge using ancient and modern analogies, although as Miall (1996) points out there is a 'gross lack of data'. Research carried out by Maizels and Aitken on Lateglacial deposits of Northeast Scotland provided palaeoflow estimates for a number of catchments. The valley floor width and clast size data for the North Esk palaeochannel system is comparable with the data from this research and suggests outwash plains with peak flows of 18,000m^3s^{-1}. The mean gradient, however, is substantially higher (0.0037) than for the Trent which suggests that this analogy is perhaps less relevant than the clast size data initially indicates. Dawson and Gardiner (1987) studied the terraces of the Lower Severn using a analogue approach which relied on a number of assumptions. Their palaeoflow estimates for the Lower Severn during the Late Devensian varied between a mean discharge of 202 - 1502$m^3s^{-1}$and a maximum of between 2015 - 14992m^3s^{-1}. Based on channel widths of between 62m and 173m, channel velocity was calculated between a range of $1.8ms^{-1}$ and $2.5ms^{-1}$ with flow depths of 1.05m - 2.5m. It may be appropriate from a palaeohydraulic perspective to compare the Main Terrace deposits of the Lower Severn with the deposits at Barrow-upon-Trent. Channel gradient is of the same order of magnitude (0.00063 for the Severn, 0.00076 for the Trent) and the sedimentology of the terraces (i.e. Lithofacies types) are similar. Using this analogy may suggest channel depths ranging from 1.05m – 1.3m and velocity at $1.8 – 2.1ms^{-1}$.

It should be re-emphasised however that any absolute measure of palaeodischarge should be taken as purely indicative. It may be more valuable to consider palaeohydraulic reconstruction by using a relative approach and clearly the inherent sedimentology of the floodplain gravels (i.e. clast size, structure, lithofacies type) at Barrow-upon-Trent suggests a high powered depositional environment very different to the contemporary system.

4.4 Other Sedimentological Features

A number of sedimentological features were observed at Barrow-upon-Trent which merit special attention. In particular, the presence of clay drape structures, iron and manganese staining and overall sediment colour indicate that the deposits were subject to post-depositional alteration.

4.4.1 Clay Drapes (Plate 8)

At location 6, the presence of a clay drape structure was recorded. Measuring approximately 3cm in thickness with a lateral extent of 13m, the structure consisted of a sub-horizontal irregular base with uneven tongue-like protrusions into the overlying sediments. Immediately underlying the structure, and moulding to the basal surface, a 17cm orange-stained, medium grade sand horizon was present. Immediately above the clay drape the sediments consisted of matrix-supported gravels. The base of the clay structure fluctuated between 12 – 32cm of the gravel top surface.

The injection of tongues of finer sediments (clays, silts) into overlying sands and gravels is well documented (Sekyra, 1961; French, 1976). Termed 'involutions', these are among some of the most widespread features thought indicative of Pleistocene frost action. The presence of periglacial involutions do not, however, necessarily suggest the former existence of permafrost, but only that seasonally frozen ground developed (French, 1976).

4.4.2 Iron Staining (Plate 10)

The feature recorded as iron staining in this section refers to the more localised orange-stained horizons occurring as distinct shallow bands within the sediments. Although a function of the same depositional processes, the overall colouration of the deposits are discussed separately (section 4.4.4).

Orange stained bands are recorded at most of the locations studied. Varying from pale to vivid orange in colour, the stained horizons are commonly sub-horizontal, usually following a particular lithofacies unit. Individual stained horizons show a restricted vertical thickness, usually less than 10cm, but may have a lateral extent of many tens of metres. A common feature of these iron stained bands is that they usually mark an interface between different lithofacies types.

The main influence on the colour of sediments is the level of the water table and groundwater movement. The presence of orange-stained horizons suggests iron oxidation as a result of lower saturation levels and exposure. Caution must be taken, however, when drawing environmental conclusions from colour since oxidation state is commonly set during early diagenesis rather than at the time of deposition (Miall, 1996).

4.4.3 Manganese Staining (Plate 10)

The black discolouration of sediments caused by manganese staining can be seen with reference to plate 10. Whilst not as widespread a feature in the deposits as iron staining, it is commonly associated with the latter and forms thin sub-horizontal banding (usually less than 10cm thick). The manganese staining, or coating of grains

Plate 10

'A' – Iron-staining

'B' – Manganese-staining (band is 3-5cm thick)

'1' – Lithofacies B

Plate 11

'A' – Orange colouration of Lithofacies B

'B' – Pink colouration of Lithofacies A and B

'1' – Lithofacies D with clay banding (Lithofacies F) at base in some cases (forming small manganese nodules usually less than 2mm in diameter) forms as a result of the precipitation of manganese ions from streams and rivers. The precipitation of the manganese ions is a function of both reduction and oxidation in depositional, and diagenic, waters (Leeder, 1982).

4.4.4 Sediment Colouration (Plate 11)

A distinct feature of the sand and gravel deposits at Barrow-upon-Trent is that, in many places (locations 1, 3, 4, 5, 6, 7, 8) the deposits are clearly distinguishable on the basis of colour. In the majority of instances there is a clear sub-horizontal boundary between overlying, predominantly orange, gravels and underlying, predominantly pink, deposits. This division occurs at varying levels in the sedimentary sequence and there does not appear to be a simple relationship between lithofacies type and colour.

Differences in sediment colouration have been well documented along the Trent valley. Brown (1996) suggested that the only way of distinguishing between Devensian and earlier deposits at Hemington was by colour ('the red iron staining of the Devensian gravels is replaced by yellow-brown staining in the Medieval gravel'). More recently, Brandon (1997) suggested that the Late Glacial – late Flandrian Hemington Gravel had lost 'the pinkish hue imparted by the Holme Pierrepont Sand and Gravels primary Triassic component and is generally more brown to pale orange-brown due to oxidation'.

5. Interpretation of Lithofacies

5.1 Braided River Interpretation

The floodplain gravels of the Trent are recognised as representing deposition in a braided system. The dominance of gravel lithofacies indicate deposition by a coarse bedload river. Although meandering rivers can carry a gravelly bed load (Bluck, 1971; Gustavson, 1978; Jackson 1978) it is suggested that the lithofacies and lithofacies assemblages mainly represent the deposits of braided streams. Meandering stream deposits are present - the very nature of the contemporary River Trent suggests that there must be, and must have been, meandering deposition - and a combination of braided and meandering sediments are also interpreted.

Collinson (1978) suggested that in order to ascertain channel type consideration should be given coarse/finer member ratio, shape of sediment bodies, palaeocurrent evidence, internal organisation of grainsize and sedimentary structures, and the overall palaeogeographic and palaeoclimatic setting of the sediments. The following considerations support a braided river interpretaion for the deposits:

i. Coarse/fine member ratio

Coarse gravel deposits predominate the study area and there are a lack of fine sand-silt-clay sediments. Sediment grading therefore suggests high energy fluvial conditions at the time of aggradation.

ii. The shape of the sediment bodies

The majority of individual sand and gravel lithofacies are sub-horizontal or massive, and relatively thin (less than 1m). The sediment bodies are likely to be sheet or bar-like with very low depth to width ratios.

iii. Palaeocurrent evidence

Palaeocurrent evidence, although showing two distinct alignment patterns (105-120° and 80°) shows low variance and indicates relatively unimodal transport.

iv. <u>The internal organisation of grain size and sedimentary structures</u>

The absence of clear epsilon cross-bedding in the gravels, and the lack of evidence that associated fine sediments were transported obliquely to the gravels, suggests that point-bar deposition was largely absent (re. Ori, 1982). The dominance of 'framework' (after Rust, 1978) gravels, represented by Lithofacies A and B, is characteristic of braided streams and suggests that major gravel deposition occurred prior to the influx of the finer matrix. Greater organisation in Lithofacies B (than Lithofacies A) suggests more continuous flow conditions whilst occasional fining-upwards units are interpreted as medial bars growing under waning flow conditions rather than as point bars. Matrix-supported gravels (Lithofacies C) indicate rapid and simultaneous deposition possibly as a result of a flood event and the alternation of clast- and matrix-supported lithofacies suggests discharge variation. These features, similar to those recorded by Steel and Thompson (1983) on Triassic braided system deposits, suggest 'medial or mid-channel braid bars dominated the river system'.

v. <u>The general palaeogeographic and palaeoclimatic setting of the deposits</u>

Previous research (Clayton,1953; Posnansky, 1958; Rice, 1968; Brandon and Sumbler,1988; Brandon, 1997) has already recognised that the floodplain deposits of the Trent are a Devensian, or later, phenomenon and therefore ideal conditions for the development of a braided system.

5.2 Braiding – Causes and Characteristics

'Braiding' in a river system has been defined by a number of authors. Rust (1972) suggested that braided rivers consist of two or more channels divided by bars or islands, with one channel usually being dominant, 'although there may be several principle channels'. More recently, Knighton (1998) has provided a definition which is useful; 'Braided rivers consist of flow separated by bars within a defined channel, bars which may be inundated at higher discharges to give the appearance of a single channel close to a bankful. The degree of bar development can vary considerably, both horizontally and vertically, with, at one extreme, occasional, low-amplitude bars and, at the other, intense bar formation almost to the level of the surrounding floodplain'. If bars do develop into stable islands then the term anabranching is used. Braided rivers are characterised by low depth/width ratios, possibly exceeding 300, (Miall, 1977) steep slopes and low sinuosities. Braided streams deposits are usually

Classification of river types

Type	Morphology	Sinuosity	Load type	Bedload per-cent (of total load)	Width/ depth ratio	Erosive behaviour	Depositional behaviour
Meandering	single channels	>1.3	suspension or mixed load	<11	<40	channel incision, meander widening	point-bar formation
Braided	two or more channels with bars and small islands	<1.3	bedload	>11	>40	channel widening	channel aggradation, mid-channel bar formation
Straight	single channel with pools and riffles, meandering thalweg	<1.5	suspension, mixed or bedload	<11	<40	minor channel widening and incision	side-channel bar formation
Anastomosing	two or more channels with large, stable islands	>2.0	suspension load	< 3	<10	slow meander widening	slow bank accretion

Table 6 Classification of river types (after Miall, 1977)

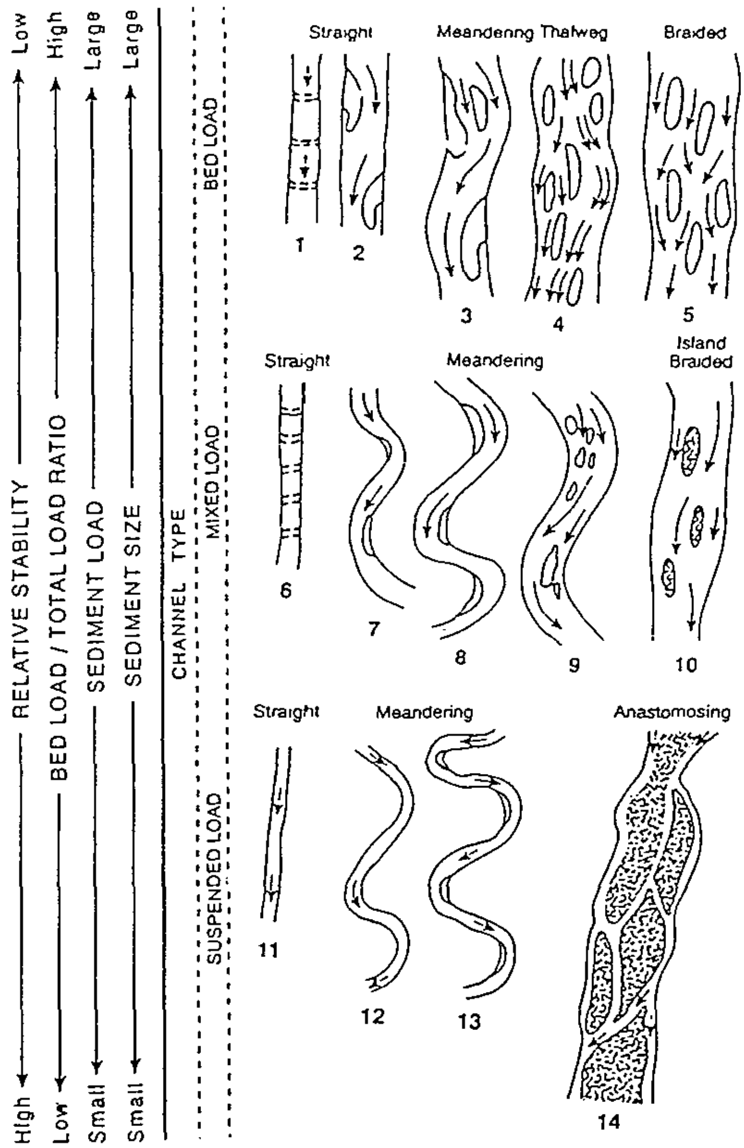

Fig.13 Classification of channel pattern
(after Schumm 1981, 1985)

coarser than those of other river types and show a dominance of gravel and/or sand. There are exceptions, such as the contemporary Amite River in Louisiana, which has typical braided river deposits and bedforms in a largely meandering setting. Table 6, adapted from Miall (1977), provides a simple classification of river types based on sinuosity, load, width/depth ratio and erosive/depositional behaviour.

Channel pattern, classified by Schumm (1981, 1985) as having 14 distinct types (Fig.13) depends on a combination of hydraulic and sedimentary factors some of which have already been referred to. Rivers alternate between meandering and braided conditions depending on at least nine variables as identified by Leopold and Wolman (1957). These include slope, depth, width, discharge amount and variability, sediment load, velocity and roughness of stream bed.

Although braiding is not systematic of overloading, the availability of large amounts of sediment is a necessary requirement. In addition the load should be of a certain calibre. As stated by Leopold and Wolman (1957) 'braiding is developed by sorting as the stream leaves behind those sizes of load which it is incompetent to handle...if the stream is competent to move all sizes comprising the load but is unable to move the total quantity provided to it, then aggradation may take place without braiding'. Deposition of the coarser bed load initiates mid-channel bar formation and as the bar develops, flow is diverted towards the channel banks thus providing the mechanism for bank erosion which results in the development of wide, shallow channels (Fig.14). Bank erodibility is an important factor in channel development. If no erosion occurs, channels remain straight, and whilst a meandering channel requires localised bank erosion, braiding involves extensive bank retreat (Knighton, 1998). Mackin (1956) attributed a meandering-braided-meandering sequence along the Wood River, Idaho to changes in bank resistance as the river passes through a corresponding sequence of forest-prairie-forest environment. Rapid discharge fluctuations, often associated with high sediment supply, are especially relevant in proglacial areas. Whilst it is accepted that braiding can develop under steady flow conditions (re: Leopold and Wolman, 1957; Hong and Davies, 1979), fluctuations in discharge contribute to bank erosion and irregular bed-load movement which lead to bar formation.

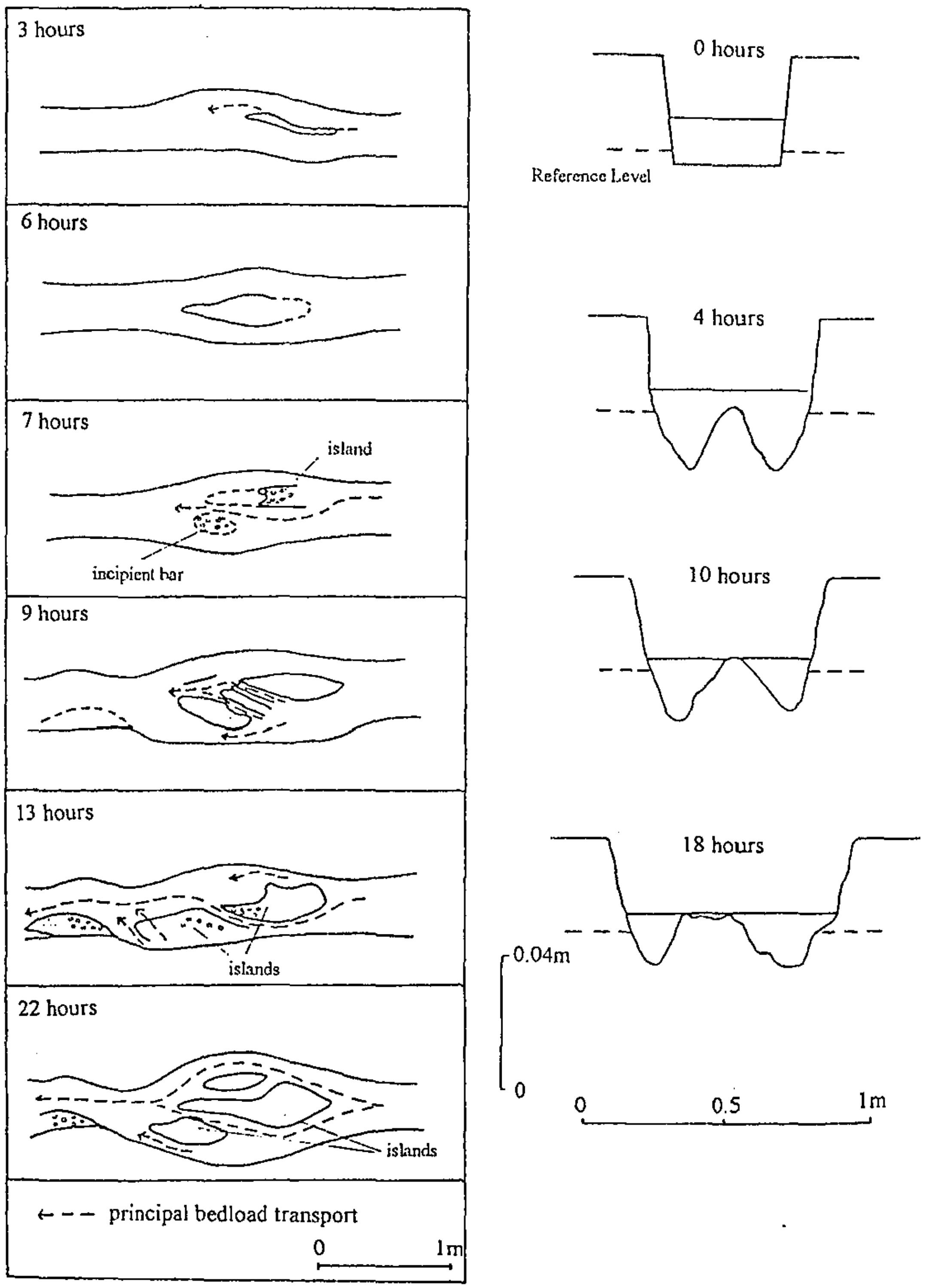

Fig.14

Stages in the plan and cross-sectional development of longitudinal bars in a flume
(after Leopold and Wolman, 1957)

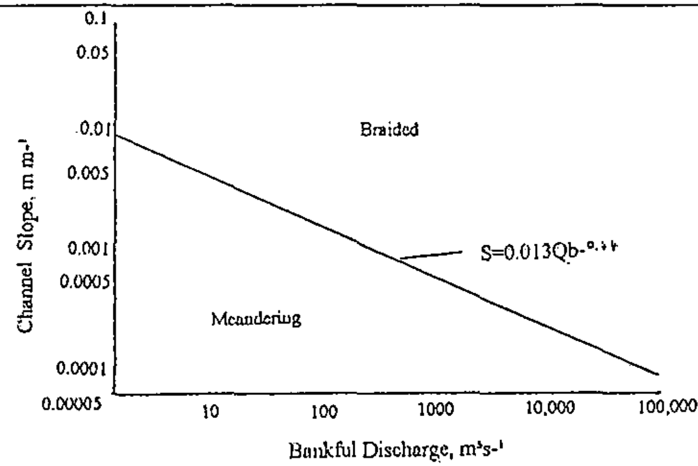

Fig.15 Distinction between braided and meandering channels on the basis of slope-discharge relationships (after Leopold &Wolman, 1957)

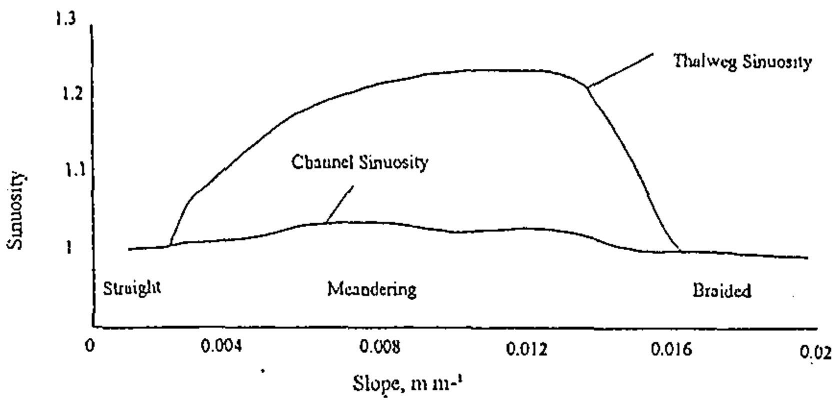

Fig.16 Channel pattern thresholds defined in terms of sinuosity-slope relationships from flume exp.s.(after Schumm and Khan, 1972)

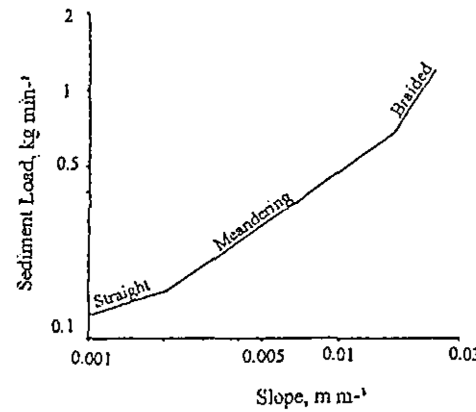

Fig.17 Channel pattern thresholds defined in terms of sediment load-slope relationships from flume experiments(from Schumm and Khan, 1972)

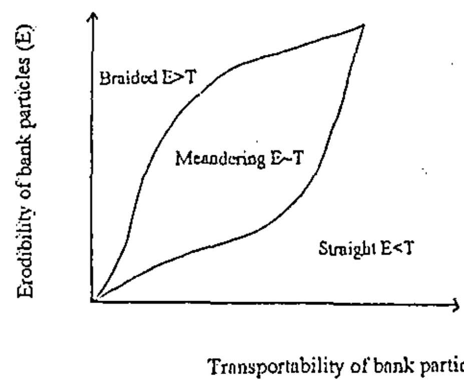

Fig.18 Stream pattern defined in terms of the erodibility and transportability of bank particles(after Brotherton, 1979)

Research by Leopold and Wolman (1957) suggests that braided and meandering streams can be distinguished on the basis of an empirical relationship between discharge and slope, $S=0.013Qb^{-0.44}$, where 'S' is slope and 'Qb' is bankful discharge (m^3s^{-1}). For a given discharge braided streams should occur on slopes steeper than given by the equation and meandering streams on gentler gradients. The degree of braiding has been shown to increase as slope increases (Chang, 1979b), however, it may be that it is the increase in stream power associated with higher gradients that is the governing factor since braiding can persist at lower slopes in large rivers (Leopold and Wolman, 1957).

Figures 15-18 illustrate some basic relationships between channel pattern and hydraulic sedimentary variables, however, it should be accepted that straight-meandering-braiding types are a function of a combination of these factors. Knighton (1998) suggests this sequence should be regarded as the association of the increase in stream power (from either an increasing discharge on a fixed slope or an increasing slope and fixed discharge, or a combination), an increasing width:depth ratio (associated with increasing bank erodibility and increasing bed-load transport) and an increasing bed-load (amount and calibre).

5.3 Morphology of Braided Rivers

Several distinct levels are recognisable in many braided channels ranging from the deepest and most active channels to elevated, abandoned areas (Kessler & Cooper, 1970).

Numerous terms have been used to describe bars in braided rivers. An older classification by Miall (1977) of bar types (Fig.19) has now been replaced by a simplified version which relates directly to depositional processes. Hein and Walker (1977) have indicated that certain bar types are members of evolutionary sequences and that the morphology of a particular bar type may be very ephemeral.

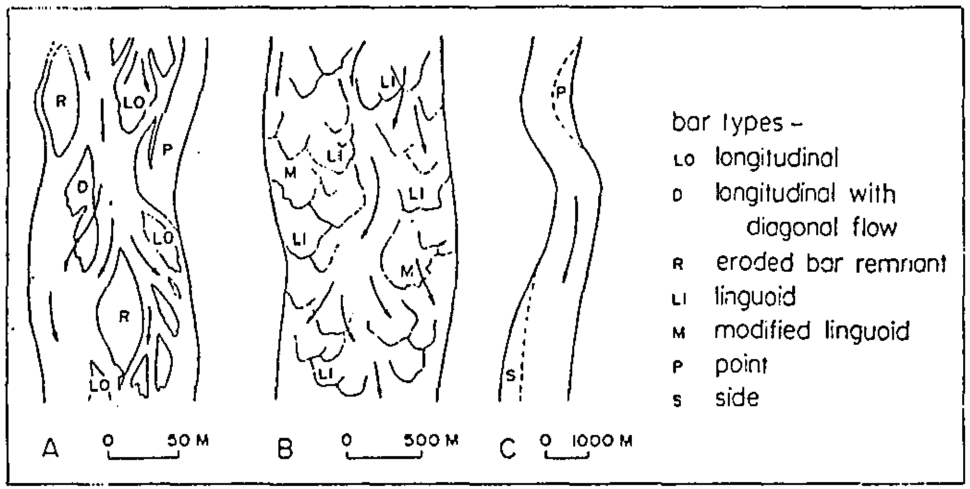

Fig.19 Principal bar types (from Miall, 1977)

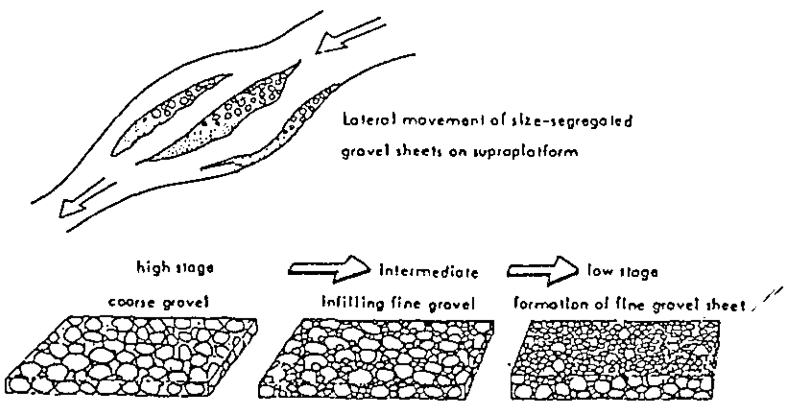

Fig.20 Vertical and downstream grainsize variation (from Steel and Thompson, 1983)

5.3.1 Horizontally stratified gravel sheets

Previously termed longitudinal bars (after Rust, 1972) these deposits are the simplest of the channel forms and are formed by clast-by-clast accretion over an obstruction or a channel lag deposit. During high water and sediment discharges these sheets grow upward and downstream by the addition of gravel clasts. Horizontally stratified gravel sheets are now regarded as channel 'mesoforms' and are amalgamated into 'macroforms' (bars) in several ways (Miall, 1996). The sheets, reaching approximately 1m in height, may show evidence of upward-fining as a result of reduced water levels over the sediments and are also shown to fine downstream (Fig. 20). Bar types associated with this macroform can either be mid-channel or attached.

5.3.2 Midchannel bars

Midchannel or medial bars are lozenge-shaped in plan, elongated parallel to flow direction and are bounded by active channels. Gravel bars are most commonly of this type (N.D.Smith, 1974; Boothroyd & Ashley, 1975) and are the classical braid-bars of Leopold and Wolman (refer to Fig. 14). Waning flow/reduction in competency or a channel obstrution results in the deposition of the coarsest bedload as a small, submerged bar over which the flow forms a riffle. Finer particles become trapped in the interstices of the initial deposit enabling growth (Miall, 1977). Bar length may reach several hundred metres with the coarsest sediments concentrated along the central bar axis (Boothroyd & Ashley, 1975). The internal structure of the bars is massive or crude sub-horizontal bedding (N.D. Smith, 1970) possibly indicating transportation in planar sheets (refer to section 5.3.1) under very high flow conditions (Rust, 1972). Variations in gravel sheet growth may result in, what have previously been refered to as, linguoid and transverse bars. Typical of sandy braided rivers these bars are suggested to form in relatively deep channels that are confined by narrow banks (Rust, 1975; S.A. Smith, 1990). Such conditions are not typical of an aggraded braiding environment but they may occur in fan-head trenches (McGowen and Groat, 1971). Linguoid bars are characteristically lobate (refer to Fig. 19) with upper surfaces dipping gently upstream towards the preceeding bar and sinuous, avalanche-slope terminations facing downstream (Miall, 1977). Commonly occurring in trains showing an out-of-phase relationship, linguoid bars vary in width, up to 150m, length, upto 300m and have a typical height of between 0.5m and 1.0m. Internally the main structure of linguoid bars is planar tabular cross-bedding representing avalanche-slope progradation. Transverse bars are genetically similar to linguoid bars except that they have straighter crests (N.D. Smith, 1972). Identification in the geological record would prove difficult, possibly only from detailed palaeocurrent data, and their recognition in modern streams tends to be arbitrary (Miall, 1977).

5.3.3 Attached bars

Bank-attached bars are termed 'point' or 'lateral' bars depending on their shape and position in the channel. 'Alternate' bars develop as a result of turbulence in an initially straight channel and evolve into point bars (Miall, 1996). Point and lateral bars are genetically similar, forming in areas of relatively low fluvial energy such as the inside of meanders where the main flow is diverted to the opposite, outside bank. Most typical of meandering streams, point bars can also occur in braided environments (such as the Kicking Horse River – N.D. Smith, 1974 and some Scottish rivers – Bluck, 1976). Developing by lateral growth, point and lateral bars may be of a much greater magnitude than midchannel bars (Collinson, 1970 illustrated a lateral bar covering over 6km in the Tana River) and tend to form in braided rivers by the coalescence of smaller bedforms, such as dunes and transverse bars (Miall, 1977). Miall (1996) recognises that the 'internal geometry and lithofacies composition of LA (point and lateral bars) is highly variable and depends on channel geometry and sediment load, but the presence of lateral accretion is the common theme'. Internal structures are complex and may include planar tabular cross-bedding (of linguoid bar origin), trough cross-bedding (of dune or scour origin), various types of ripple marks, coarse-grained lag deposits and fine grained drapes and fills. The term 'compound bar' has previously been used for these structural elements.

5.3.4 Modifications of bar form

Modification of bar form can occur as a result of both increase and decrease in flow regime. These modifications are best summarised by Bristow and Best (1993) 'where bars exist for periods of time in excess of a single flood event they will experience a complex history of erosional and depositional modifications related to changes in stage. At higher flow stages when the largest volumes of sediment are transported, the channels are often scoured, bars may be reduced in height or in some cases completely eroded. However, during falling stage maximum deposition occurs as discharge and flow competence are reduced. Channel beds aggrade, the high stage bedforms may be modified and new bars may be formed or enlarged as sediment is deposited. As discharge continues to fall, bars may become emergent and dissected by low stage channels. Additionally, the nature of the falling limb recession......will be important not only in the reworking of higher stage sediments, but also in the deposition and spatial distribution of the finer grained sediments (silts and clays)...'

5.4 Lithofacies Interpretation

Lithofacies classification is currently the standard method for the interpretation of sedimentary deposits. Miall (1977) provided a review of the braided river environment and suggested that there was a consistency in the lithofacies assemblages that occurred in both modern and ancient sand and gravel sequences. Miall proposed a practical lithofacies classification that could be used for the recognition of all fluvial deposits. The classification has been expanded (Miall, 1978c) and modified and is currently the 'standard field methodology for the examination of fluvial deposits' (Miall, 1996).

A total of nine architectural elements are recognised in the fluvial system (from Miall, 1996). These are: channels (CH), gravel bars and bedforms (GB), sandy bedforms (SB), downstream-accretion macroforms (DA), lateral-accretion macroforms (LA), scour hollows (HO), sediment gravity flows (SG), laminated sand sheets (LS) and overbank fines (FF). Inherent within each of these architectural elements are principal facies assemblages which can be interpreted in terms of their 'hydrodynamic origin and position within the braided-river tract'.

The six lithofacies identified in this research are comparable with a number of lithofacies types as described by Miall (1996) and are interpreted as representing particular architectural components.

5.4.1 <u>Gravel Bars and Bedforms (GB)</u> – Lithofacies A, B, C

Horizontal gravel sheets and midchannel bars, as previously described, develop as a result of clast accretion, growing laterally and vertically in the river channel. The bedforms are of low amplitude and may appear massive in section because of obscure bedding contacts (S.A.Smith, 1990, suggests that the 'textural and structural monotony of conglomerates may make recognition and correlation of bounding surfaces more difficult'). Where bedding is defined it is usually sub-horizontal, as recognised in lithofacies A and B. The coarseness and structure of these deposits testify to high water discharge and relatively high sediment concentration in the flows and indicate that accumulation occurred in areas of low relief. Lithofacies B, showing a greater degree of organisation than lithofacies A, suggests that formation occurred at times of more continuous flow. The clast framework of lithofacies A and B suggests that major gravel deposition occurred prior to the finer matrix influx. Occasionally, the clast matrix is absent, indicating the winnowing of fines during discharge decrease, or alternatively, perhaps, increased reworking of earlier sediments. Grading within lithofacies A and B may suggest that the deposits represent part of a fining upward or downstream sequence within the bar. The inherent variability and relationship of lithofacies A and B can be explained in terms of their position on the midchannel bar and the amount of modification that resulted during grainsize seggregation, and variations in flow stage. Steel and Thompson (1983) suggested that where disorganised beds (Lithofacies A in this research) occur in thin sequences, it is likely that they are of bar-head origin, largely during high-flow stage. The more organised beds (lithofacies B in this research) are likely to have accumulated at lower levels on the bar (e.g. bar margins or tail) where exposure to more prolonged reworking and winnowing produced a clast supported 'armour'.

The structure and content of deposits that resemble Lithofacies C have previously been ascribed as representing debris flow deposition (re: Miall, 1996, facies Gmm). Debris flow deposits are characterised by usually massive, unstructured, matrix-supported clasts suspended in a matrix of sand, silt and mud. The formation of debris-flow deposits has been described and interpreted by A.M.Johnson (1971). Johnson explains that debris flow 'is a process by means of which granular solids, sometimes mixed with relatively minor amounts of entrained water and air, move readily on low slopes'. The movement of these flows is likened to that of large waves moving steadily through channels with superimposed smaller, and more rapid, waves riding on top. Whilst recognising that the deposits of Lithofacies C are indeed of a similar nature to those of debris flow deposition, it is suggested here that this interpretation is probably not valid. It has already been suggested that debris flows operate on relatively gentle gradients, however, it is also true that most flows are as a result of the accumulation of sediments from higher and steeper slopes (such as valley sides or canyons, e.g. Titus Canyon and Wrightwood, California – A.M.Johnson, 1971). Moreover, most debris flows are relatively proximal deposits (i.e. restricted range) and contain high clay/silt/sand content. Lithofacies C contains a higher sand content (30%) than other gravel lithofacies but suffers from a distinct lack of clay and silt grade sediments (refer to Fig.11) that would be expected from debris flow deposition. Although it is possible that some fines may have gone into suspension, or been winnowed out, the overriding fact that the research area occupies a relatively flat tract in a lowland valley suggests that debris-flow was not a major depositional factor. Debris flows can occur as a result glacial outwash, althiugh it is suggested here that Lithofacies C more likely represents (debris) flood events. Dawson and Gardiner (1987) have recognised that disorganised matrix-rich gravels may occur as the 'surface of complex bars or on lateral within-channel bends'. More likely, however, Steel and Thompson (1983) suggest that matrix supported conglomerates in the 'Bunter' Pebble Beds are as a result of simultaneous deposition of sand and cobbles in a 'flow in which there was high sediment concentration and rapid deposition', such as a debris flood. It should also be acknowledged that Carling (1990) has used fluvial modeling experiments to show that different textural variation, such as clast and matrix support in gravels, are part of the same depositional process. If nothing else, the research by Dawson and Gardiner, Steel and Thompson, and Carling, demonstrates that matrix-supported assemblages, such as lithofacies C, are complex and may represent a number of different depositional mechanisms. As a point of comparison, Brandon (1997) has interpreted the matrix-supported gravels of the Holme Pierrepont Sand and Gravel as a 'braid plain distal sandur deposit' (therefore characteristic of rapid aggradation) from the 'Devensian ice front situated above Burton-upon-Trent'.

5.4.2 Sandy Bedforms (SB), Downstream Accretion (DA) and Lateral Accretion (LA) – Lithofacies D, E

Miall (1996) has made the distinction between predominantly 'aggradational' sandy bedforms and predominantly 'accretional' deposits ('DA' and 'LA') on the basis of accumulative style and geometry. For the purpose of this research Lithofacies D and Lithofacies E are regarded as sandy bedforms, which inherently represent the products of downstream, lateral or vertical, accretion, or a combination.

Accepting that downstream and lateral accretion elements are very common in braided environments, Miall (1996) recognises that most sandy lithofacies develop as a result of both vertical aggradation and lateral accretion. The difference between downstream and lateral accretion is fairly self-explanatory (i.e. accretion in the direction of flow or more perpendicular to flow, respectively) and there are various methods of determining these styles (re: Miall, 1996). Since there is a distinct gradational relationship between 'DA' and 'LA' sedimentation, it is very difficult to identify these depositional processes in the field. The designation 'DA/LA' is used is when this so. Lithofacies D (horizontally-laminated sand) equates with Lithofacies 'Sh' of Miall (1977) and develops as a result of one of two conditions. Most commonly, this occurs during flood stage when plane-bed conditions develop at critical flow and the channel floor becomes a traction carpet, with virtually continuous particle flow parallel to water movement (Harms and Fahnestock, 1965).

Miall (1996) suggests that, for this condition to develop, water depths are usually between 0.25 and 0.5m, with velocities in the order of 1ms-_. Plane bed conditions also develop at lower velocities at shallower depths, and in coarse sands at low flow speeds (~0.4ms-_), although the latter condition is rarely preserved (Miall, 1996).

The interpretation of Lithofacies D as representing plane-bed flood conditions is distinctly different to that suggested for the second sand lithofacies. Consisting as solitary or grouped planar cross-beds, Lithofacies E occurs as distinct small lenses within gravel lithofacies. The occurrence and form of these units suggests relatively restricted ripple and dune development, with the angle of cross-bedding (18-22°) representing avalanche deposition on forsets. Comparable with lithofacies Sp of Miall

(1977), these features are a dominant component of linguoid and transverse bars and are interpreted as such. Variations in the basic form of Lithofacies E can be attributed to different flow conditions i.e. forsets flatten out as separation eddies decrease (re: Miall, 1997).

5.4.3 Overbank Fines (FF) – Lithofacies F

Lithofacies F can be regarded as comprising distinct sub-facies (refer to section 4.1.6). Each sub-facies represent similar low-energy conditions of formation in different depositional areas of the floodplain. Occurring as thin clay bands, 'pods' or distinct palaeochannels, these sediments commonly show grading features, illustrating fluctuations in discharge, and tend to be associated with lithofacies D (horizonatlly bedded sand). The sub-facies comprising Lithofacies F are interpreted as the deposits of overbank or waning floods, and abandoned channels and pools. Overbank areas in braided rivers are usually small, in comparison with overall channel magnitude, and this, in combination with the shifting nature of channels (causing erosion), explains why deposits of this type are less evident in fluvial sediments than other lithofacies types. Flood deposition of fine sediments in inter-channel areas ('from suspension or weak traction currents', Miall, 1996) will tend to result in the deposits having interlaminations of clay, silt and fine sand occurring along an undulating bedding plane. Very often the deposits of major depositional episodes are separated by fines deposited by waning floods, but in many cases 'these are scarce owing to their removal by subsequent scour or to an inherent lack of fines in the sediment supply' (Collinson, 1978). If deposition is in 'standing pools of water during low stage channel abandonment' then the lower surface of the clay drape will correspond to the shape of the underlying bedform (Miall, 1977). An extension of this depositional regime produces channel fill sequences in abandoned areas, such as recorded at Barrow-upon-Trent (this research, section 4.2).

5.5 Lithofacies occurrence

The lithofacies have been identified as representing braided river deposition. It has already been acknowledged that meandering streams are able to carry coarse bedloads, and many of the elements and deposits found in braided tracts, such as downstream, lateral and vertical accretion elements, are also common to meandering reaches. The causes and characteristics of both braiding and meandering patterns have also been alluded to (section 5.2). Many braided versus meandering classifications have been attempted (Cant and Walker, 1979; Miall, 1977) although it is now clear that these classifications may be oversimplistic (re: Jackson, 1978; Miall, 1996) and that successions of braided and meandering deposits may look very similar. An added complication arises when a system (such as the Trent) comprises of both braided and meandering deposits deposited in different time frames, or at varying downstream positions along the same riverine tract. In these cases, and without the aid of other data (such as palaeocurrent information) it may be impossible to confidently distinguish between these deposits.

One method of distinguishing between braided and meandering streams is to look at the relative abundance of lithofacies types. Table 7 (adapted from Miall, 1977) illustrates some simple lithofacies abundance comparisons between braided and meandering rivers for the lithofacies recognised in this research (A – F).

<table>
<tr><th>Lithofacies Occurrence</th><th>Braided Rivers</th><th>Meandering Rivers</th></tr>
<tr><td>A</td><td rowspan="5">Common</td><td rowspan="3">Rare to common (generally as a thin lag deposit)</td></tr>
<tr><td>B</td></tr>
<tr><td>C</td></tr>
<tr><td>D</td><td>Common</td></tr>
<tr><td>E</td><td>Generally rare</td></tr>
<tr><td>F</td><td>Rare to common</td><td>Common</td></tr>
<tr><td>Channel – Fill Sequences</td><td>Rarely > 3m</td><td>Commonly > 3m</td></tr>
</table>

Table 7 Lithofacies abundance in braided and meandering streams
(adapted from Miall, 1977)

The recognition of proximal and distal lithofacies assemblages can also aid in the differentiation of different fluvial systems. Although there are no absolute indicators that can be used to estimate proximity to source in braided streams, there are several parameters such as grainsize (mean or maximum), stream power and bar form, that can be used to provide a relative estimation between proximal and distant deposits. Certain lithofacies types are inherently related to these parameters and it is possible to give a relative distinction based on lithofacies distribution, i.e. more proximal deposits are represented by coarse-grained Lithofacies A, B and C, whilst more distal deposits are represented by finer grained facies, and a subsequent lack of Lithofacies A, B and C. A step further is to suggest that meandering streams, in which Lithofacies D, E and F are common, are more likely to appear at more distal sites.

Lithofacies abundance and proximal-distal relationships are clearly very helpful in distinguishing between braided and meandering types, however, it must be emphasised that the distinction of channel pattern in the geological record relies on a combination of evidence. Correspondingly, the evidence provided in this research, apart from the considerations presented in section 5.1, is based on interpretations made on the nature, assemblage and association of lithofacies present.

5.6 Lithofacies Models

Historically, braided river deposits have generally been thought as 'somewhat random in depositional character and lacking recognisable cyclicity' (Miall, 1996). Since 1973 (e.g. Miall, 1973; Cant and Walker, 1976) however, this has been demonstrated not to be the case. A useful tool in the identification of braided river environments is based on the vertical and lateral profiles of the component lithofacies. A number of depositional models have been proposed towards this aim (Picard and High, 1974; Miall, 1977; Cant and Walker, 1978). Perhaps one of the best known models is that for the south Saskatchewan River (Cant and Walker, 1978) which provides a pictorial synthesis of the geomorphological elements of a sandy braided river with that of bedforms, internal structure and stratification sequence. In 1977, Miall presented four empirical facies models based on the descriptions of ancient deposits but named 'after a modern river which appears to typify the interpreted depositional environment'. The 'Scott' type model represented deposition in a shallow gravel braided river in which Lithofacies 'Gm' (equating with Lithofacies A and B in this research) is dominant. Deep gravel braided rivers are typified by the 'Donjek' model whilst the 'Platte' and 'Bijou Creek' types represent sand-dominated environments. Subsequently, Miall (1985) reviewed these models, adding a further 8 examples, and in the light of additional research, has now expanded these to include 16 different fluvial lithofacies assemblages (Miall, 1996). The models are classified into gravel-dominated and sand-dominated fluvial 'styles' depending on lithofacies content. Fig.21 illustrates the six major gravel lithofacies models identified by Miall.

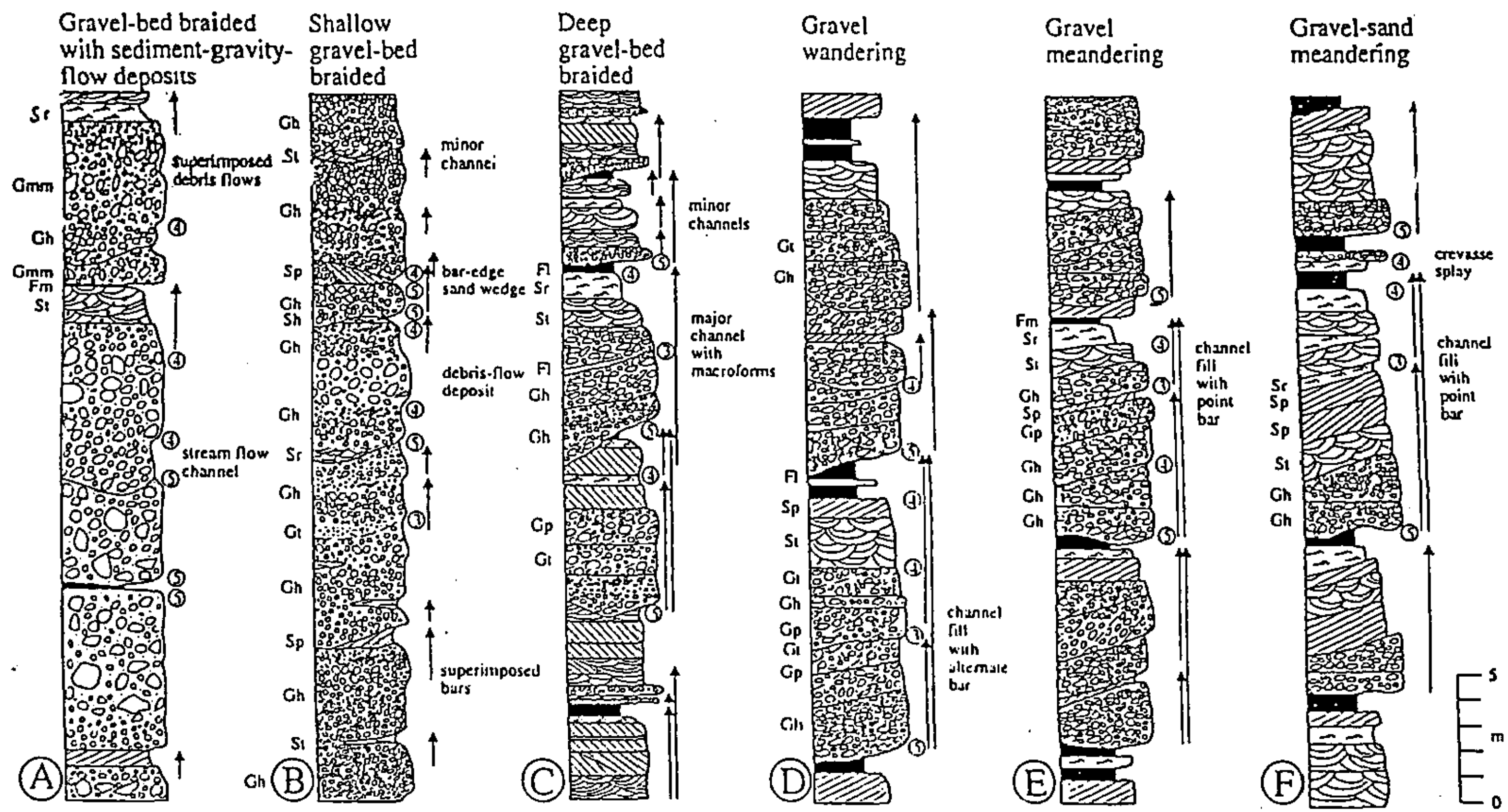

Fig.21 Gravel Lithofacies Models (from Miall, 1996)

For interpretation purposes, Lithofacies A and B in this research, are identified as Lithofacies 'Gh' in the coding of Miall. Lithofacies C, as previously discussed (section 5.4.1), whilst resembling the deposits of Lithofacies 'Gmm', has not been defined as representing the same fluvial processes. Lithofacies D and E identify with codes 'Sh' and 'Sp' respectively, whilst Lithofacies F is represented by the codes with the prefix 'F'. Clearly, it is not possible to identically match a particular fluvial model with a lithofacies sequence in nature, however, it is possible to recognise the model that produces the 'best fit' with the recorded lithofacies and assemblages. Correspondingly, the lithofacies and assemblages recognised in this research can clearly be identified with the shallow gravel-bed braided model (Fig.21), the definition of which Miall (1996) describes as 'proximal gravel bed rivers...in which sediment-gravity flows are rare to absent, consist of a shifting network of unstable, low-sinuosity channels in which a variety of gravel bedforms is deposited. Channel depths on the order of 1m are typical. Channel margins are rarely identifiable in outcrop. Element GB predominates and consists of tabular bodies with numerous minor internal erosion surfaces, and varying assemblages of gravel traction-current deposits. ...channels may be abandoned at low stage, in which case, thin lenses and wedges of sand may be deposited comprising element SB...Typically, element SB comprises about 5% of most fluvial successions formed in these type of rivers'

6. Discussion

At the start of this research a number of objectives were identified (section 1.4). The following section provides a synthesis of the findings of this research and attempts to answer the questions posed.

6.1 Middle Trent Valley

6.1.1 Late Glacial Context

The general Late Glacial history of the Trent Valley is in little doubt. At its maximum extent the Late Devensian ice sheet lay to the east of Sheffield and west of Derby at Burton-upon-Trent (refer to Fig.1). The ice sheet, which left much of the Trent Basin uncovered, is thought to have originated in western Scotland and the Lake District, and then entered the Irish Sea (Shotton, 1977c). Other ice masses were also developing from Ireland and North Wales at this time and this led to deviation of the Scottish-Lake District ice down the Cheshire plain (Jones and Keen, 1993). The retreat of the ice sheet across Cheshire and Shropshire is suggested to have been a 'complex process, involving halts and perhaps minor readvances' (Jones and Keen, 1993) and more significant movements during the Younger Dryas (refer to section1.2.1). The maximum expansion of Devensian Ice occurred during the latter part of the cold stage, between 26000BP – 13000BP (Dimlington Stadial), when ice blocked the Humber Estuary and created a proglacial/glacier-dammed 'Lake Humber' (Fig.22), diverting the Trent through the Lincoln Gap (central East Anglia). It has been suggested (Straw, 1963a) that an earlier Devensian 'Lake Humber I' also formed as a result of ice blockage but this is not widely accepted. Incision, following the re-opening of the Humber Gap, eventually restored the Trent to its present course.

6.1.2 Deposits

Brandon (1997) has interpreted the Holme Pierrepont Sand and Gravel (refer to section 1.3) as a 'braid plain distal sandur deposit from the Late Devensian ice front situated above Burton-upon-Trent in the Trent Valley and near Uttoxeter in the Dove Valley'. The description of these deposits is as pink, poorly sorted, matrix-supported,

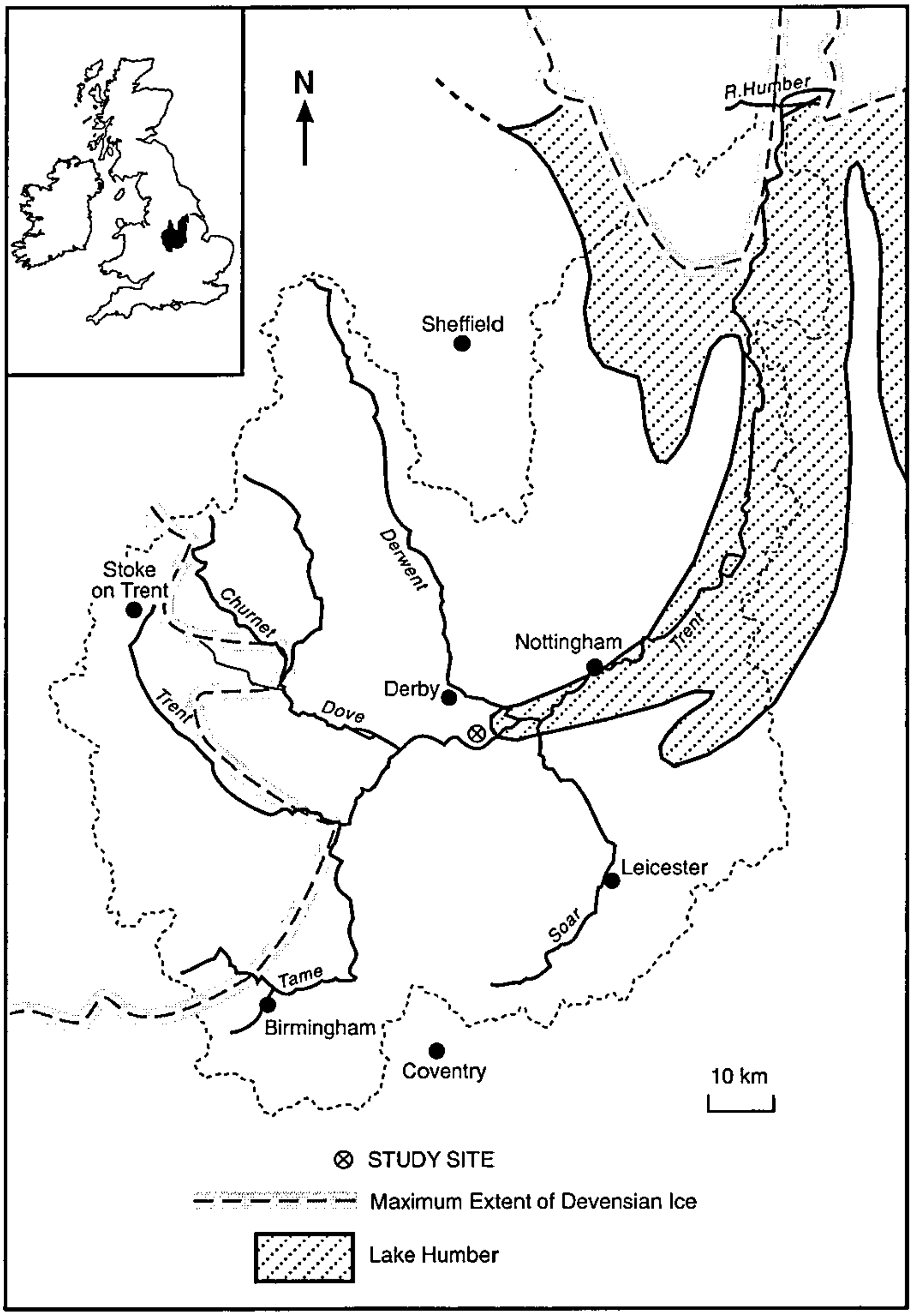

Fig. 22 Lake Humber

trough-crossbedded gravels with numerous sub-horizontal erosion surfaces. Typically, these deposits are recorded as 8m thick (ranging up to 10m). Overlying the Holme Pierrepont Sand and Gravels, Brandon (1997) records Hemington Gravel Deposits (refer to section 1.3). These brown clast-supported planar cross-bedded ('cryptically' bedded or apparently massive at Barrow-upon-Trent) gravels are suggested to be reworked Holme Pierrepont Sand and Gravel deposits. Laterally 'accreted at channel point bars by successive meanderings of the Trent from possible Late Glacial to Late Flandrian times' these gravels are said to be distinguishable from the underlying sediment on the basis of colour and maturity (and the presence of truncated ice wedge casts, when found). Accepting that the 'depositional history of the Hemington Gravel is probably very complex', Brandon suggests that the age of this deposit varies across outcrops.

This research has looked at the floodplain deposits of the Trent very differently to other research carried out along the Trent Valley (e.g. Brandon, 1997; Lillie & Grattan, 1995; Brandon & Sumbler, 1988). Examining the deposits by way of lithofacies analysis, this research has identified the likely sedimentological mechanisms responsible for the deposition of the sands and gravels. This research has identified six distinct lithofacies types, which, together with their associations, are likely to represent deposition in a braided river environment. It is acknowledged that the floodplain gravels do feature elements of meandering deposition (refer to section 5.5) however, for the reasons presented earlier in this research (sections 5.1, 5.5, 5.6) the deposits are regarded predominantly as braided in origin.

Attention has been brought to the internal structure and framework of the floodplain deposits. It has been suggested that the Devensian Holme Pierrepont Sand and Gravel is typically matrix-supported whilst the later Hemington deposits are clast-supported and is 'a more mature deposit which is considerably less sandy than the older gravel' (Brandon, 1997). This research suggests that there is no simple relationship between clast- and matrix-supported framework, and stratigraphical sequence. Clearly the question of 'clast or matrix support' in gravels can be fairly subjective, however, this research has identified many instances where clast- and matrix-supported gravels are interdigitated in a sequence and commonly where matrix-supported gravels lie above clast-supported gravels at the very top of a section. Other research has identified the presence of planar and trough cross-bedding in gravel sequences along the Trent (Brandon, 1997; Lillie & Grattan, 1995). Gravel thicknesses were relatively restricted at Barrow-upon-Trent (3-4m) and this hampered recognition of large scale relationships within and between lithofacies. This research, however, suggests that the majority of gravel sediments are sub-horizontal, representing deposition by gravel sheets and midchannel bars, which would be expected from braided river deposition. Palaeocurrent evidence taken from different height intervals in the sedimentary sequence at Barrow-upon-Trent (refer to section 4.3.1) also suggests relatively unimodal flow (and therefore typical of braided tracts) during aggradation of the deposits. It is suggested that changes in palaeoflow have occurred since original deposition. Evidence from the alignment of 'bedrock channels' (refer to section 3.3.1) suggests scour occurred in a southwest-northeast trend whilst palaeocurrent evidence suggests palaeoflow at 80° and 105-120° at different site locations. Brandon records palaeocurrents of 140° whilst the palaeochannel alignment (refer to section 4.2.1) also records a direction of approximately 120°. The presence of palaeoscars in the southern half of the site, however, suggest a meandering planform and a palaeoflow towards 60° in more recent deposits. Clearly there has been some variability throughout depositional history.

Differentiation between Late Glacial and Flandrian deposits has been attempted on the basis of colour at various sites along the Trent Valley (Brown, 1996; Brandon, 1997). The distinction is made between predominantly pink ('red iron staining'-Brown, 1994; 'pinkish hue...from the primary Triassic component' - Brandon, 1997) Devensian gravels and the more orange-brown (from oxidation) colouration of Flandrian deposits. This distinction of colour (refer to section 4.4.4) can be made at Barrow-upon-Trent, however, it is not clear that the direct assumption can be made as to the age of sedimentation. The palaeochannel identified by this research suggests that at one location the underlying (pinkish) gravels are older than 13060 ± 90 (CAL yr BP) and the overlying (orange) gravels are at least younger than 13060 ± 90 (CAL yr BP). Clearly at this location it is possible to make inferences on colour and stratigraphy, however, this is not always the case e.g. Devensian and Medieval gravels at Hemington (Brown, 1992) are only distinguished by the presence of ice wedge pseudomorphs.

6.2 Questions Answered?

The objectives identified at the start of this research considered the adaptation of the Trent basin from an ice marginal system to an interglacial regime:

- Clearly the relationship between floodplain stratigraphy and river morphology has been established. The floodplain gravels are identified as representing mainly deposition by proglacial outwash braided streams towards the end of the Late Glacial (from the retreating ice front situated to the west of Burton-upon-Trent). Six major lithofacies are recognised at Barrow-upon-Trent and have been interpreted in terms of their origin and position with the braided system. Lithologically, the gravels consist mainly of quartzites, quartz and other durable lithologies (refer to Table 2) and are most likely to represent reworked 'Bunter' pebble deposits (now 'Cannock Chase formation') which are widespread to the west and south-west of Derby (refer to Fig. 2, section 4.). Subsequently, a reduction in sediment and flow conditions associated with the retreating ice sheet and the return to an interglacial environment during the Flandrian, led to more favourable conditions for the development of a meandering system.
- It is possible to distinguish braided and meandering systems on the basis of lithofacies analysis. Complications, however, do arise because certain lithofacies assemblages are common to both braided and meandering rivers, and many river systems such as the Trent, comprise both braided and meandering deposits. The Trent is well documented for its large floods and frequent channel changes and is, in this respect, atypical of British lowland rivers (Brown, 1996). Flood events, such as those recorded at Hemington (represented by Medieval gravels) can deposit unusually coarse deposits within a meandering setting. Apart from the presence of truncated ice wedge casts, and sediment colour, these deposits may well have been interpreted as a Devensian phenomena. Without the aid of additional information, therefore, such as dating information and palaeocurrent data, it is unlikely that any conclusion can be drawn with the utmost confidence.
- Similarly it is possible to distinguish between Late Glacial and Flandrian sediments. Again, colouration of deposits and sediment maturity have been used (Brown, 1996; Brandon, 1997), however, unless this subjective evidence is supported with information from dating, (or relative information from truncated

ice wedge pseudomorphs) then any interpretation must be made tentatively. The presence of cryoturbated structures at Barrow-upon-Trent (section 4.4.1) suggests periglacial activity, however, they may purely be the result of more isolated seasonally frozen ground conditions.

- It is possible to compare the sediments at Barrow-upon-Trent with other reaches of the River Trent. Fieldwork at Barrow-upon-Trent records gravel thicknesses of between 2.0-5.4m (typically 3.0-4.0m). Evidence from borehole data suggests increased gravel thicknesses, up to 11.5m (usually 6m+) present near the contemporary channel of the Trent and a corresponding increase in overbank fines thickness towards the contemporary channel. Evidence from regional borehole data, and previous research carried out along the Trent Valley by various authors (Brandon, 1997; Lillie and Grattan, 1995; Knight and Howard, 1994; Brandon and Sumbler, 1988) has supported the variability and provenance of the floodplain deposits. Surprisingly, whilst gravel thicknesses increase towards the confluence of the River Trent with the River Dove and River Derwent, this increase is not as marked as one might perhaps expect at the intersection of two major streams. In many places gravel deposits cut into the underlying bedrock but show no relationship with regard to proximity of the contemporary channel. Where this occurs, it is suggested that channel migration has contributed.
- The deposits at Barrow-upon-Trent are interpreted as representing deposition by a shallow gravel braided river, the 'Scott' type model of Miall (1996). The investigations of river terrace deposits rely on the proposition of sequence stratigraphy and are therefore regarded as 'models' of terrace formation. Dawson and Gardiner (1987) reviewed the terraces of the River Severn and provided a summary which is equally applicable to the Trent and many other River Basins: 'it is thus unlikely that a simple model of terrace aggradation can be generally applicable, and clearly some assessment of the local depositional environment is necessary prior to using the existence of a terrace for a stratigraphic or palaeohydrological interpretation'.

This contribution has considered the palaeohydrological relationship between floodplain stratigraphy and river morphology, and has in some way fulfilled, or partially fulfilled, the objectives originally presented. This research has also established that River systems are more complex than many authors recognise and further research needs to be an ongoing theme.

APPENDIX I – BOREHOLE DATA

LOCATION	REFERENCE	BOREHOLE DATA
THREE DIMENSIONAL PROFILES	STEETLEY GEOLOGICAL SERVICES DEPT- ATKINS & GOODWINS LAND BOREHOLE SITES (SEE NOTE 1)	AK 1 – AK 27 (BH'S 1 – 27) GW 1 – GW 11 (BH'S 28 – 38)
MIDDLE TRENT	SK22NE	BH'S 1 - 191
	SK32NW	BH's 1 -189
	SK32NE	BH's 1 - 282
	SK42NW	BH's 1 - 241
	SK42NE	BH's 1 - 426
	SK43SE	BH's 1 - 161
	SK43SW	BH's 1 - 301
TRANSECT A1 – A2	SK32NE progressing SW - NE	BH's 86, 14, 15, 88, 89, 17, 91, 92, 93, 94, 18, 19, 103, 105, 107, 109
	then SK435W progressing SW - NE	BH's 51, 52, 53, 54, 55, 56, 57, 58, 60, 61, 69, 74, 77, 80, 81, 82, 91, 97, 98, 103, 107, 108, 109, 110, 111, 113, 116, 117, 118, 119, 120, 121, 122, 123, 124, 128
TRANSECT B1 – B2	SK32NE progressing N - S	BH's 52, 53, 54, 55, 56, 57, 58, 59, 60, 61, 62, 63
TRANSECT C1 – C2	SK42NE progressing SE - NW	BH's 424, 426, 419, 386, 379, 378, 414, 377, 412, 299, 298, 297, 296, 250, 295, 293, 291, 290, 286, 273
	then SK42NW progressing SE – NW	BH's 237, 257, 194, 168, 193, 221, 215, 214, 191, 190, 212, 211, 208, 203, 200, 197
	then SK43SW progressing SE - NW	BH's 148, 228, 227, 146, 145, 223, 144, 222, 143, 286

- Note 1 – Borehole data provided by Mr. C. Camberbach, Redland Aggregates Ltd (now Lafarge Aggregates Ltd), Bradgate House, Groby, Leicester, LE6 0FA.

- All other information provided by the Records Dept., British Geological Survey, Keyworth, Notts, NN12 5GG. Please note that borehole locations are recorded as point data on 1:10,000 scale Ordnance Survey maps and numbered (e.g. 1 to 241) with reference to specific sheet numbers e.g. 'SK42NW'. Grid References are provided on the individual borehole records, which can be accessed by prior arrangement with the British Geological Survey at the above address.

APPENDIX II – PARTICLE SIZE DATA

SAMPLE LOCATION	LITHOFACIES	SEDIMENT GRADING (PERCENT RETAINED)
Location 1	Lithofacies B	-6ø=0%; -5ø=0%; -4ø=21.4%; -3ø=34%; -2ø=14.1%; -1ø=5.4%; 0ø=4.2%; +1ø=9.1%; +2ø=9.5%; +3ø=1.6%; +4ø=0.3%; +5ø=0.4%
Location 1	Lithofacies C	-6ø=0%; -5ø=15.8%; -4ø=30.8%; -3ø=21.6%; -2ø=8.7%; -1ø=4.2%; 0ø=3.0%; +1ø=5.9%; +2ø=7.1%; +3ø=2.4%; +4ø=0.5%; +5ø=0.5%
Location 2	Lithofacies A	-6ø=12.5%; -5ø=18.3%; -4ø=23.8%; -3ø=11.7%; -2ø=8.7%; -1ø=5.3%; 0ø=3.0%; +1ø=7.8%; +2ø=8.1%; +3ø=0.6%; +4ø=0.1%; +5ø=0%
Location 2	Lithofacies D	-6ø=0%; -5ø=0%; -4ø=0%; -3ø=0%; -2ø=0.1%; -1ø=1.4%; 0ø=2.9%; +1ø=24.2%; +2ø=69.9%; +3ø=1.2%; +4ø=0.2%; +5ø=0.1%
Location 3	Lithofacies A	-6ø=3.6%; -5ø=22.1%; -4ø=23.4%; -3ø=17.7%; -2ø=9.0%; -1ø=3.7%; 0ø=2.4%; +1ø=14.6%; +2ø=2.7%; +3ø=0.4%; +4ø=0.2%; +5ø=0%
Location 3	Lithofacies C	-6ø=0%; -5ø=0%; -4ø=22.3%; -3ø=30.6%; -2ø=14.3%; -1ø=2.8%; 0ø=1.6%; +1ø=7.0%; +2ø=19.6%; +3ø=1.4%; +4ø=0.2%; +5ø=0.1%
Location 3	Lithofacies E	-6ø=0%; -5ø=0%; -4ø=0%; -3ø=0.1%; -2ø=0.3%; -1ø=0.5%; 0ø=1.6%; +1ø=30.1%; +2ø=64.4%; +3ø=2.4%; +4ø=0.3%; +5ø=0.2%
Location 4	Lithofacies A	-6ø=0%; -5ø=20.5%; -4ø=36.4%; -3ø=18.6%; -2ø=5.2%; -1ø=1.9%; 0ø=1.5%; +1ø=4.9%; +2ø=9.9%; +3ø=0.8%; +4ø=0.1%; +5ø=0.1%
Location 4	Lithofacies A	-6ø=22.9%; -5ø=37.4%; -4ø=11.3%; -3ø=10.0%; -2ø=5.1%; -1ø=2.4%; 0ø=1.8%; +1ø=4.7%; +2ø=3.3%; +3ø=0.7%; +4ø=0.2%; +5ø=0.2%
Location 4	Lithofacies B	-6ø=4.9%; -5ø=7.7%; -4ø=27.0%; -3ø=22.4%; -2ø=10.5%; -1ø=4.5%; 0ø=2.9%; +1ø=17.8%; +2ø=1.9%; +3ø=0.2%; +4ø=0.1%; +5ø=0.1%
Location 5	Lithofacies A	-6ø=0%; -5ø=34.0%; -4ø=20.1%; -3ø=15.1%; -2ø=8.1%; -1ø=3.8%; 0ø=2.6%; +1ø=14.7%; +2ø=1.0%; +3ø=0.4%; +4ø=0.2%; +5ø=0.1%
Location 5	Lithofacies B	-6ø=0%; -5ø=6.1%; -4ø=34.1%; -3ø=23.2%; -2ø=11.0%; -1ø=4.5%; 0ø=2.5%; +1ø=9.7%; +2ø=7.7%; +3ø=0.6%; +4ø=0.3%; +5ø=0.2%
Location 7	Lithofacies A	-6ø=8.8%; -5ø=38.7%; -4ø=18.2%; -3ø=13.1%; -2ø=5.6%; -1ø=2.5%; 0ø=1.6%; +1ø=5.3%; +2ø=5.2%; +3ø=0.8%; +4ø=0.1%; +5ø=0.1%
Location 7	Lithofacies B	-6ø=0%; -5ø=8.3%; -4ø=24.7%; -3ø=22.0%; -2ø=12.0%; -1ø=5.0%; 0ø=2.8%; +1ø=14.4%; +2ø=9.4%; +3ø=0.9%; +4ø=0.3%; +5ø=0.2%
Location 7	Lithofacies C	-6ø=0%; -5ø=4.0%; -4ø=22.3%; -3ø=27.2%; -2ø=12.5%; -1ø=4.7%; 0ø=2.9%; +1ø=17.7%; +2ø=7.6%; +3ø=0.7%; +4ø=0.3%; +5ø=0.1%
Location 8	Lithofacies C	-6ø=0%; -5ø=14.7%; -4ø=10.5%; -3ø=15.6%; -2ø=10.0%; -1ø=7.7%; 0ø=5.8%; +1ø=16.2%; +2ø=16.5%; +3ø=1.9%; +4ø=0.5%; +5ø=0.5%

APPENDIX III – PALAEOCURRENT DATA

SITE LOCATION	TYPE OF DATA	PALAEOCURRENT (DIP & DIRECTION)
Location 2	Pebble Imbrication (Long Axis Alignment)	23/102°; 15/026°; 07/136°; 24/107°; 15/130°; 19/087°; 31/116°; 23/094°; 24/162°; 15/112°; 13/110°;18/122°; 18/184°; 24/094°; 16/117°; 18/123°; 12/112°; 20/072°; 29/115°; 26/126°; 18/168°; 14/057°; 19/115°; 22/111°; 17/110°
Location 2	Forset Dips	21/112°; 22/110°; 21/114°
Location 3	Forset Dips	21/114°
Location 8	Forset Dips	22/106°
Location 10	Forset Dips	19/076°; 21/077°; 21/082°; 20/080°; 18/080°
Location 12	Forset Dips	21/083°; 19/078°; 19/082°; 21/077°; 20/078°

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