The use of uplift modelling in the reconstruction of drainage development and landscape evolution in the repeatedly glaciated Trent catchment, English Midlands, UK

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Abstract
The Trent Valley Palaeolithic Project has recently investigated the Quaternary evolution of the River Trent, the northernmost river system in western Europe with a documented long-timescale terrace staircase. The uppermost and lowermost reaches of the Trent, which drains the English Midlands, were glaciated during Marine oxygen Isotope Stage (MIS) 2, but older fluvial terraces dating back to MIS 8 are preserved in the remainder of the catchment, delineating the former course through the Lincoln Gap and across the Fen Basin (the modern course to the Humber estuary dating only from the latest Pleistocene). Numerical modelling enables lateral variations in uplift across the catchment to be deduced from differences in height of these fluvial terraces above the modern valley floor. Uplift rates thus indicated over the last two climate cycles attain values of ~0.08 mm a^{-1} around Nottingham and Derby in the middle reach of the Trent, but are significantly lower elsewhere in the catchment; these variations are shown to relate to lateral variations in crustal properties, primarily variations in radioactive heat production in the underlying continental crust. Glaciation during the late Middle Pleistocene (MIS 8) caused significant changes to the Trent catchment, including the integration of the modern Upper Trent with the rest of the system. Older sedimentary evidence is much more fragmentary, but is used along with the results of the uplift modelling to reconstruct the earlier drainage. It is thus inferred that between the Anglian (MIS 12) and Wragby (MIS 8) glaciations the Trent already flowed into the Fen Basin via the Lincoln Gap, but the smaller-than-present catchment, indicated by gravel lithology, resulted in a much steeper longitudinal gradient, such that during interglacials (MIS 11 and 9) an elongated estuary would have developed, extending inland almost to the present location of Newark. Prior to the Anglian, much of the modern Trent catchment, including the rivers Derwent and Dove, drained into the former Bytham River. The modern Middle Trent catchment downstream of Nottingham was drained by a relatively small 'Ancaster Trent' river, which flowed above the Ancaster Gap; analysis of gravel lithology suggests that it probably joined the Bytham in the area that now forms the Fen Basin.

Key words: Trent, Pleistocene, fluvial terraces, uplift, glaciation

1. Introduction
The Trent (Fig. 1) is the third longest river in Britain, after the Thames and Severn. At ~53°N, near the northern limit of land that remained unglaciated during Marine oxygen Isotope Stage (MIS) 2, it is the most northerly

river system in western Europe for which a long-timescale terrace staircase has been documented (Fig. 2). It is also the most northerly fluvial system from which assemblages of Lower Palaeolithic artefacts have been amassed, this providing the rationale for the multi-disciplinary Trent Valley Palaeolithic Project (TVPP), completed recently under the auspices of English Heritage, aimed at improving understanding of the Quaternary geological and landscape context of this archaeological material (White et al., 2007a; Bridgland et al., 2014).

Being also located at the boundary between 'lowland' to 'upland' Britain, the record from the Trent system is also of potential value for understanding landscape development in response to long-timescale regional uplift across this transition between topographic provinces, as well as for identifying the cause in terms of lateral variations in the physical properties of the Earth's crust. This synthesis will use numerical modeling to establish the uplift history since the Early Pleistocene at selected (representative) localities, including Willington, Nottingham, Derby, and Leicester, as well as Tattershall, which lay within the Lincoln Trent catchment that persisted until the latest Pleistocene (White et al., 2007b; see below; section 2.1; Fig. 1). The basis of this analysis is age and height data from fluvial terrace deposits, including new data acquired during the TVPP, which has also provided significant new insights regarding the chronology of the Trent terrace deposits and their disposition in relation to glacigenic sediments (Bridgland et al., 2010; 2014; White et al., 2010; Fig. 3). Older data have also been synthesized, in particular geological mapping by the British Geological Survey (BGS; e.g., Brandon and Sumbler, 1988, 1991) and reports on Mineral Assessment Review (MAR) boreholes.

Figure 1 here: Location and drainage map

Figure 2 here: Summary long profile of the Trent terraces

Figure 3 here: Schematic west–east profile through the Newark-Lincoln area

The entire Trent catchment was overridden by ice during the Anglian (MIS 12) glaciation, as a result of which little or no sedimentary evidence survives from earlier times. Prior to the Anglian, the main W–E drainage in this region took the form of the much larger Bytham River, which lay >30 km to the south of the modern Middle Trent (e.g., Rose, 1989; Lee et al., 2004; contra Gibbard et al., 2013; Fig. #4(a)); the limited evidence from the Trent catchment from this period will thus be interpreted in relation to the more extensive sedimentary evidence from the Bytham. The Late Devensian (MIS 2) glaciation was much less extensive; during this glaciation the Cheshire ice lobe, derived from the Irish Sea region, impinged upon the upper reaches of the Trent, whereas the more easterly North Sea ice lobe, which appears to have developed somewhat later (e.g., Clark et al., 2012), adjoined its lower reaches. However, most of the Trent catchment remained unglaciated (Fig. 1), as a result of which it has a terrace record extending back to the Middle Pleistocene. Most of this record can be shown to date from and since MIS 8; indeed, it believed, from TVPP findings, that the Trent catchment was glaciated during MIS 8, when the extent of glaciation was larger than in MIS 2 although smaller than in MIS 12 (e.g., Westaway, 2010b; White et al., 2010; see section 2.2). The sedimentary evidence pre-dating this glaciation is extremely limited, although fluvial terraces that formed during and after deglaciation in late MIS 8 are well preserved (Bridgland et al., 2014). Although interpretations of the Trent sequence in the mid-20th century envisaged what would now be described as a post-Anglian/pre-Devensian glaciation (e.g., Clayton, 1953; Posnansky, 1960), recent received wisdom has recognized only Anglian (MIS 12) and Late Devensian glaciations in central England, with the exception of small enclaves in north Birmingham and Lincolnshire (e.g., Lewis, 1999; Maddy, 1999; Howard et al., 2007; cf. Straw, 2000, 2002a, 2011).

It is a well-established principle that river terrace heights can provide a record of uplift, subject to the assumption that palaeoflow and sediment-transport regimes have been consistently repeated, for example at corresponding phases within successive ~100 ka Milankovitch climate cycles (e.g., Maddy, 1997; Maddy et al., 2000; Westaway et al., 2002; Bridgland and Westaway, 2008b). This requires the catchment area of a river (which determines its downstream gradient, these quantities being inversely related) and downstream course length (which determines the height reached by a river of a particular gradient in any given upstream locality) to have remained roughly constant. An additional requirement for fluvial incision to serve as a direct proxy for uplift, apparent from previous studies of other British rivers (e.g., Maddy and Bridgland, 2000; Bridgland and Schreve, 2009; Westaway, 2011a), is for the terrace heights to be unaffected by local effects of glacio-isostasy; sediment emplaced when the underlying crust is depressed due to loading by an adjacent ice sheet will rebound following deglaciation to a higher level than would otherwise be expected. Therefore, in order to establish which Trent terraces can be used to infer uplift, one needs to know which terrace deposits were emplaced under equivalent conditions, unaffected by glacio-isostasy or by course or catchment changes. Distinguishing between these different effects in the Trent is difficult, and requires careful assessment of the evidence; indeed, preliminary solutions have been published (e.g., by Westaway, 2007) that can now be seen to be incorrect and are, therefore, superseded by the present work. The present study uses TVPP data and interpretations (Bridgland et al. (2014), along with other evidence, to deduce, using numerical modelling of the heights of river terraces, variations in uplift and related differences in crustal properties across the Trent catchment. Key aims are to use the available fragmentary evidence and the uplift modelling to reconstruct the geometry of the Trent catchment both before MIS 12 and between MIS 12 and MIS 8; the principal changes to the Trent catchment thus identified are summarized in Fig. 4.

Figure 4 here: Sequence of palaeogeographic maps

2. Key findings of the Trent Valley Palaeolithic Project

The state of knowledge of the Trent sequence, including the consensus view on the evolution of that river system, was summarized by Howard et al. (2007) as a baseline statement at the outset of the TVPP. They noted that previous research had established that the Trent formerly flowed through the Lincoln Gap to the Fen Basin, by way of the modern Lower Witham valley, and that all but the lowest terrace followed this route (Brandon and Sumbler, 1988, 1991; Fig. 2). Another putative early course of the Trent, through the now dry Ancaster Gap, favoured in earlier literature (e.g., Swinnerton, 1937; Fig. 4(a)), was considerably more dubious, although early work during the TVPP revealed deposits in this vicinity that threw light on the issue (White et al., 2007a; Bridgland et al., 2014; see below).

The original classification of named terraces in the Trent, established by Clayton (1953) and utilized by Posnansky (1960) and others, envisaged three terraces in decending sequence: Hilton, Beeston and Floodplain. The highest of these was later subdivided into Eagle Moor, Etwall and Egginton Common in the Middle Trent and Eagle Moor and Balderton in the Lower Trent (Brandon and Sumbler, 1988, 1991; A.S. Howard et al., 2009). In addition, the Floodplain Terrace was renamed Holme Pierrepont and local names were given to the Beeston Terrace: Bassingfield in the Trent Trench and Scarle in the Lower Trent (Fig. 2). The TVPP classification uses the local name 'Sandiacre' for what has appeared on BGS maps of the Middle Trent as Eagle Moor (Bridgland et al., 2014), retaining the name 'Eagle Moor' within the Lower Trent; the Eagle Moor, Balderton and Scarle terraces of the Lower Trent in turn project into the Martin, Southrey and Tattershall Castle terraces in the Lower Witham valley downstream of the Lincoln Gap (Figs. 2 and 3).

The terraced reaches of the Middle Trent and Lower Trent are separated by the Trent Trench, an incised gorge-like reach, bounded by cliffs or steep bluffs up to ¬ 60 m high (Fig. 2), cut across the bedrock dipslope into relatively erosion-resistant divisions (the Radcliffe and Gunthorpe formations) of the Triassic Mercia Mudstone (e.g., Bridgland and White, 2007). The Trent Trench forms most of the reach of the river between Nottingham and Newark (Fig. 1), being most clearly developed between Radcliffe-on-Trent (National Grid reference ¬ SK645396) and East Stoke (~SK746500). It has a near-constant northeastward orientation, its ~2–3 km wide floor being formed in Holocene alluvium and the Holme Pierrepont and Bassingfield terrace deposits, with no preservation of any older Pleistocene fluvial sediment (Bridgland et al., 2014). Within this relatively constricted reach, the modern river has a longitudinal gradient of ~0.4 m km^{-1}; for example, from Edwards (1966) it falls by ~6.4 m in the ~15.5 km straight-line distance between Holme Pierrepont (~SK630396) and Fiskerton (Nottinghamshire; SK737512). Measured along the meandering course of the river within the Trent Trench, the gradient would be less, ~0.3 m km^{-1}, the distance increasing to ~22.5 km. As Fig. 2 indicates, the longitudinal gradient of the Trent increases upstream from the Trent Trench towards its headwaters. Conversely, downstream of the Trent Trench, where the valley is less constricted, the gradient of the Trent decreases. For example, the river falls from ¬ 8 m O.D. near Winthorpe in the northern outskirts of Newark (¬ SK802576) to ¬ 5 m O.D. near Low Marnham (~SK818684), a gradient of ~0.27 m km^{-1}. Furthermore, the modern valley floor of the Witham (palaeo-Trent) downstream of Lincoln falls from ~3 m O.D. in the eastern outskirts of Lincoln (~TF000712) to ~2 m O.D. near Fiskerton (Lincolnshire; ~TF063712), a gradient of ~0.16 m km^{-}1. The typical longitudinal gradient of the palaeo-Trent downstream of the Trent Trench can thus be tentatively estimated as ~0.2 m km^{-1}.

The fluvial terraces are disposed subparallel to these modern longitudinal profiles, the height differences between the terraces increasing gradually upstream from the Lower Witham valley to the Lower Trent and

Middle Trent (Fig. 2). However, unlike many other rivers, the vertical separations of these terraces are not simple consequences of uplift. As will become clear, the height difference between the Sandiacre and Etwall terraces is inferred to result from glacio-isostatic perturbation of the river (Section 2.2), whereas that between the Beeston and Holme Pierrepont terraces includes adjustment in response to diversion of the river to the shorter Humber course in the latest Pleistocene (Section 2.3). The key findings of TVPP relevant to the present study fall within three main areas: development of age-control evidence (Section 2.1); resolution of relations between Trent terrace deposits and glaciation (Section 2.2); and improved stratigraphic documentation and correlations (Section 2.3).

2.1 Chronological control

The oldest Trent terrace deposit for which direct dating evidence exists is formed by the Balderton Sand and Gravel in the reach downstream of Newark. A key site, documented during the TVPP, is Norton Bottoms quarry (SK863588; Figs 2 and 3). Fossiliferous interstadial deposits had been encountered in earlier workings here, yielding Mollusca from which amino-acid dating indicated an age within MIS 6 for the Balderton Sand and Gravel (Miller and Hollin, 1991). Exposures, studied in detail for the TVPP in 2006, revealed previously undocumented organic sediments, evidently filling one or more intraformational channels, near the base of the deposits. They were overlain by 3 m of massive, subhorizontally bedded, clast-supported Balderton Sand and Gravel, disrupted by intra-formational ice-wedge casts, demonstrating emplacement under cold-climate conditions, evidently during MIS 6 (cf. Brandon and Sumbler, 1988, 1991). The Balderton sequence is blanketed by a clay-rich, rubified horizon, interpreted as a palaeosol representing soil formation during the Ipswichian (MIS 5e) and showing subsequent periglacial disruption during the Devensian (Brandon and Sumbler, 1988, 1991). This key site is documented in detail by Bridgland et al. (2014), a summary of their results being provided here.

The fossiliferous channel fills at the base of the Balderton Sand and Gravel in Norton Bottoms quarry represent fully interglacial conditions; they have yielded 22 freshwater and 11 terrestrial mollusc taxa, including thermophilous species, some (*Belgrandia marginata*, *Bithynia troscheli* and *Corbicula fluminalis*) that are absent from Britain at present and one species (*Pisidium clessini*) that is globally extinct. *Corbicula fluminalis* and *Pisidium clessini* are of biostratigraphic significance in that they are not found in Britain in deposits of Last Interglacial (Ipswichian; MIS 5e) age and are therefore indicative of an earlier temperate-climate event (e.g., Preece, 1995, 1999; Keen, 1990, 2001; Meijer and Preece, 2000). Important evidence for the interpretation of these fossiliferous sediments comes from amino-acid dating of calcite opercula of the gastropod *Bithynia tentaculata* from Norton Bottoms. These data (discussed by Penkman, 2007; Penkman and McGrory, 2007, Westaway, 2009a, Penkman et al., 2011, and Bridgland et al., 2014) cluster closely with those from other sites attributed to MIS 7, confirming that age for the basal Balderton sediments. Calibration of the kinetics of serine–alanine decomposition (Westaway, 2009a) suggests the latter part of MIS 7 (i.e., MIS 7c or 7a).

Data from the Southrey terrace at Coronation Farm, Southrey (east of Lincoln; Fig. 2), are entirely comparable (Penkman, 2007; Penkman and McGrory, 2007; Westaway, 2009a; Penkman et al., 2011), confirming the long-established view (e.g., Brandon and Sumbler, 1991) that the Southrey Terrace is the downstream correlative of the Balderton Terrace (Fig. 2). The presence, within the equivalent Bain tributary deposit at Tattershall Thorpe, of *Corbicula fluminalis*, albeit in modest quantities (Holyoak and Preece 1985; cf. Girling, 1977), supports the age assignment for the correlative Balderton and Southrey terrace deposits, this species being an indicator of a pre-Ipswichian age (see above). The Egginton Common Formation of the Middle Trent, the upstream correlative of the Balderton, lacks evidence for dating, the correlation being based on projection of terrace heights (Fig. 2).

Below the level of the Balderton Terrace and its counterparts, the Beeston Terrace, named after a type locality in the SW outskirts of Nottingham, has long been recognized (e.g., Clayton, 1953; Posnansky, 1960), and can be traced throughout the Middle Trent upstream of Nottingham. The Scarle Sand and Gravel, the equivalent of the Beeston in the reach between Newark and Lincoln (Brandon and Sumbler, 1988), has yielded a modest cold-climate mammalian fauna that, on the basis of the stratigraphy, can be attributed to the Devensian (White et al., 2007a; Bridgland et al., 2014). Farther downstream, the Tattershall Castle Terrace (Figs 2, 3) has yielded important biostratigraphical and dating evidence. In particular, at the Tattershall Castle type locality fossiliferous channel-fill deposits yielded temperate-climate molluscs missing from Britain at present, notably

Belgrandia marginata, Vallonia enniensis, *Discus ruderatus* and *Nesovitrea petronella* (Holyoak and Preece 1985), but not *Corbicula fluminalis*, indicating that the interglacial represented is MIS 5e. This is corroborated by interglacial sediments with the biotratigraphically significant mammal *Hippopotamus amphibius* in the lowermost Derwent (e.g., Arnold-Bemrose and Deeley, 1886; Jones and Stanley, 1974), in a deposit equivalent to the Beeston Formation of the main river, and, immediately west of the Lincolnshire Limestone escarpment, in the Fulbeck Sand and Gravel, which represent laterally equivalent sediments attributable to the Witham tributary upstream of its erstwhile confluence with the Trent near Lincoln (Brandon and Sumbler, 1988; Fig. #02(b)). A correlative terrace deposit of the River Bain in the Tattershall area, investigated during the TVPP at Kirkby-on-Bain quarry (TF236604), is confluent with the Tattershall Castle terrace gravels of the main river (Fig. 2). Temperate-climate deposits, interbedded within the sand and gravel of this quarry, have yielded amino-acid data indicative of an Ipswichian (MIS 5e) age (e.g., Penkman, 2007; Penkman and McGrory, 2007; Westaway, 2009a; Penkman et al., 2011). OSL dates of >25 ka for the youngest cold-climate deposits in this quarry and from correlative sites are discussed below. Figure 5(a) illustrates a long-profile projection of the Beeston Terrace and its counterparts after subsequent uplift (estimated in section 4) has been restored, i.e., showing the height at which the uppermost deposits forming this terrace were emplaced.

Figure 5 here: Long profiles with uplift restored

Straw (1979) regarded the Tattershall Castle terrace deposits and their Bain counterparts in the Tattershall area as ‘Fen border gravels’, representing a delta built into a proglacial lake occupying much of the Fen Basin, impounded by the advance of Devensian ice into coastal Lincolnshire and Norfolk. Although this idea was re-stated by Murton and Murton (2012), the TVPP investigations confirmed that these are fluvial terrace deposits (Bridgland et al., 2014; cf. Holyoak and Preece, 1985). Moreover, although Devensian ice indeed surged southward to the Lincolnshire and Norfolk coasts, this event is now dated ~19–17 ka (Clark et al., 2012) or ~21–17 ka (Bateman et al., 2011), a considerable time after these fluvial sediments were emplaced.

Figure 6 here: Comparison of dates for the Beeston and HP terraces

In earlier TVPP publications the end of deposition of the Beeston Sand and Gravel was assigned a nominal age of MIS 4 or ~70 ka (cf. Westaway, 2007; White et al. (2007a). It is, however, apparent that its emplacement continued significantly later. First, Girling (1974, 1977) identified fossiliferous deposits at Tattershall Castle, within the Tattershall Castle Sand and Gravel, that she assigned to the Upton Warren Interstadial, which would place them in MIS 3, a view that the TVPP interpretation now supports (Bridgland et al., 2014). Straw (1979) also listed many radiocarbon dates for organic material from the Tattershall area, the youngest of which, 30,800±360 ^{14}C B.P. (or ~35.0±4.0 ka, calibrated; Fig. 6) was from the now flooded complex of quarries (TF210570) at the Tattershall Castle type locality, later documented by Holyoak and Preece (1985). The results of OSL dating undertaken as part of the TVPP also support younger ages for the Scarle–Tattershall Castle terrace of the Trent (Schwenninger et al., 2007a,b; Bridgland et al., 2014). OSL dating of many sites, of Middle and Late Pleistocene ages, was attempted as part of the TVPP. The dates obtained for Middle Pleistocene sites are in most cases significantly younger than expected on other grounds, indicating the effect of systematic error, and are of little value in this study. However, potentially valuable dates were obtained from the Scarle Formation (Fig. 6) from the South Scarle (SK856639) type locality (62.9±7.5 ka and 45.2±9.3 ka), from Kirkby-on-Bain quarry (TF236604) in the correlative terrace deposit of the River Bain (33.7±5.0 ka and 26.5±4.4 ka), and from Langford quarry (SK810603), north of Newark (30.8±3.3 ka and 27.9±2.8 ka). On the basis of these above dates, we now envisage that emplacement of the Beeston Sand and Gravel and its counterparts continued as late as MIS 2, ending circa 25 ka (Bridgland et al., 2014; Fig. 6).

A third line of evidence for the latest age of the Beeston Formation relates to the correlation of its Upper Trent counterpart, the Second Terrace deposits (Fig. 7), with outwash deposits and glacial overflow channels related to the Devensian glaciation of north-west England, as will be discussed below (section 3.3). This implies a mismatch between the timing of the maximum MIS 2 ice extent on the western and eastern side of Britain, with the earlier Irish Sea–Cheshire ice (at its maximum extent) feeding into the Beeston Formation, whereas the later North Sea ice reached its maximum extent after 21 ka (Bateman et al., 2011; Clark et al., 2012), during the emplacement of the younger Holme Pierrepont Formation (Bridgland et al., 2014; see below).

Figure 7 long profile projection of the Upper and Middle Trent terraces

The lowest terrace of the Trent, and the only one to carry the same name in all but the uppermost reaches of the system, is the formed by the Holme Pierrepont Sand and Gravel, with an upper surface typically 1-2 m above the modern floodplain; this is the 'Floodplain Terrace' of earlier workers (e.g., Clayton, 1953; Posnansky, 1960). In BGS mapping (BGS DigMap) this deposit has been defined from Barton-under-Needwood (Figs 1 and 7) in the lower reaches of the Upper Trent to the Gainsborough area, beside the modern course towards the Humber. The correlative First Terrace of the Upper Trent can be traced farther upstream (Stevenson and Mitchell, 1955), as illustrated in Fig. 7 (see also section 3.3). The Holme Pierrepont Sand and Gravel is typically ~6–7 m thick in the Middle Trent, with a maximum thickness of ~9 m (e.g., Posnansky, 1960; A.S. Howard et al., 2009). The corresponding sediment can also be traced downstream into the Witham valley, as a buried gravel known from boreholes beneath the floodplain alluvium (e.g., Crofts, 1982; Jackson, 1982; Westaway, 2007), indicating that the diversion of the Trent from the Lincoln Gap (Fig. 4(f)) to the Humber occurred while it was flowing at this level. Correlative deposits are also present in many Trent tributaries (see Bridgland et al., 2014, for details; also section 3.2).

TVPP OSL dates for the Holme Pierrepont Sand and Gravel (Schwenninger et al., 2007a, 2007b; Bridgland et al., 2014) include 25.3±3.3 ka at Besthorpe Quarry (SK817626) and 21.0±1.9 ka and 18.9±1.9 ka at Girton Quarry (SK827683), both in the Lower Trent downstream of Newark but upstream of the divergence between the Lincoln and Humber courses (Fig. 6). A date of 24.2±3.5 ka was obtained from Barton-under-Needwood (SK200160) in the Upper Trent. A rather younger date, 10.9±2.9 ka, was obtained from Barrow-on-Trent quarry in the upper Middle Trent (SK344275), and is consistent with Lateglacial (~13–14 ka) calibrated radiocarbon dates for organic deposits underlying this gravel here and at the nearby (SK460303) site of Hemington (Greenwood et al., 2003). This range of numerical ages suggests that much of the Holme Pierrepont Sand and Gravel was emplaced shortly after the downcutting following the end of deposition of the Beeston Sand and Gravel, although other deposition occurred much later.

Posnansky (1960) indeed envisaged a rather complex sequence of aggradation and re-incision during the emplacement of what is now known as the Holme Pierrepont Sand and Gravel. In his view, a first phase of aggradation occurred following the initial downcutting; at this time the river presumably still flowed through the Lincoln Gap. This was followed by the incision of a buried channel, the base of which falls northward from O.D. at Cromwell (~SK800615; ~8 km north of Newark) to ~11 m below O.D. at Marton (~SK840820), south of Gainsborough, in ~22.5 km, a downstream gradient of ~0.5 m km^{-1}, evidently leading towards the Humber. This channel can be inferred to mark the diversion of the river to the Humber. According to Posnansky (1960), its incision was followed by renewed deposition of gravel and the modern floodplain alluvium. Of the dated sites listed above, Besthorpe, Girton and Barton-under-Needwood would presumably relate to the Lincoln course and Barrow-on-Trent to the Humber course. However, given the complex stratigraphy, in the absence of dating evidence it is impossible to say whether any particular bed within the Holme Pierrepont Sand and Gravel at any point upstream of the Trent diversion formed during the initial aggradation, when flow was towards the Wash, or during the subsequent aggradation when it was directed into the Humber. A further complicating factor, recognized for example by Howard (2007), A.S. Howard et al. (2009) and Howard et al. (2011), is the presence close to the modern floodplain of a very low terrace, known as the Hemington Terrace in the Middle Trent and the Quorndon Terrace in the Soar; the former is dated to the Holocene using radiocarbon and the presence of Neolithic and later archaeology.

New insights into the stratigraphy of the Holme Pierrepont Sand and Gravel were provided by Howard et al. (2011), who studied the sediments at its type locality, Holme Pierrepont Quarry (~5 km east of Nottingham; SK625385). They reported that sediments near the base of the local ~5 m succession yielded radiocarbon dates of ~11 ka, indicating deposition during the Loch Lomond Stadial. These deposits evidently fill a buried channel, possibly the one recognized by Posnansky (1960) as having formed when the Trent was diverted to the Humber. However, at Wilford on the southern outskirts of Nottingham, erosion-resistant Triassic bedrock crops out in the bed of the modern river channel (e.g., Lamplugh et al., 1908; A.S. Howard et al., 2009), evidently limiting the magnitude of this incision phase in localities farther upstream. Sediments of equivalent age to those at Holme Pierrepont Quarry are evident at Barrow-on-Trent quarry in the upper Middle Trent (see above) but, farther upstream, deposits at the level of the Holme Pierrepont terrace are significantly older. One example, noted above, is Barton-under-Needwood; sediments of similar disposition in the Tame valley at Coleshill have yielded a radiocarbon date of ~32 ka (Coope and Sands, 1966; Howard et al., 2007), which adjusts after calibration to ~37±2 ka. Furthermore, near Quorn, southeast of Loughborough (at SK 556 181), the basal part of sediments that BGS DigMap designates as the Syston Sand and Gravel, the Soar counterpart of the Holme Pierrepont

Formation, has been radiocarbon dated to ~29 ka (Brown et al., 1994; Howard et al., 2011), which adjusts to ~33 ka after calibration (Fig. 6).

Another instance of relatively old sediments at the expected level of the Holme Pierrepont Sand and Gravel is at Whitemoor Haye Quarry (SK173127) near the Tame–Trent confluence, where sediments beneath the First Terrace of the Upper Trent have yielded palaeoenvironmental data indicative of MIS 3 and well preserved (but not recovered *in situ*) skeletal remains of the woolly rhinoceros *Coelodonta antiquitatis* (Schreve et al., 2013). Schreve et al. (2013) reported calibrated radiocarbon dates from these bones in the range 44.5–47.8 ka as well as OSL dates from the basal Pleistocene sediments (their 'Lower Sands'), possibly from a higher stratigraphic level than that from which the rhinoceros bone was recovered. These 'Lower Sands' yielded two OSL dates that Schreve et al. (2013) considered reliable, which are 24±4 and 22±3 ka when calculated using standard corrections for moisture saturation (and thus comparable with the TVPP OSL dates reported above); the overlying 'Upper Gravels' likewise yielded two more apparently reliable OSL dates, of 15±2 and 11±2 ka. However, Schreve et al. (2013) argued that these sediments had probably been water-saturated since deposition, requiring larger corrections for this effect, increasing the four calculated numerical ages to 39±6, 37±6, 28±4 and 19±3 ka, respectively. Furthermore, the two OSL dates for the 'Upper Gravels' are not in stratigraphic order; combining them gives weighted means (each plus-or-minus twice the standard error) of 13.0±2.8 ka or 22.2±3.4 ka, depending on choice of correction for water-saturation (Fig. 6).

Figure 6 indicates that the available OSL dates support a switch from emplacement of the Beeston Sand and Gravel to incision, then to emplacement of the Holme Pierrepont Sand and Gravel, at ~26–25 ka. However, the limiting radiocarbon dates for the Holme Pierrepont Sand and Gravel in Fig. 6 are significantly older. A potential explanation for this is the 'hard water effect': dissolution in the river water at the time of deposition, or in groundwater subsequently circulating through the sediments, of carbon from ancient sources (e.g., the Carboniferous Limestone or Coal Measures rocks, or the constituents of the Triassic Mercia Mudstone Group, some of which are carbonate-cemented or contain calcite nodules; e.g., A.S. Howard et al., 2009) might affect the carbon isotope ratios in the samples in a manner that mimics the effect of greater age. However, Brown et al. (1994) used ^{13}C isotopic data to exclude this possibility in the case of their date from the Syston Sand and Gravel, although it is noteworthy that three other radiocarbon dates from the same deposit were evidently contaminated by ancient carbon and thus deemed unreliable (i.e., systematically old) by those authors; likewise, Schreve et al. (2013) present ^{13}C isotopic data that exclude this possibility at Whitemoor Haye. Setting the data from Whitemoor Haye aside, the youngest known occurrence of C. *antiquitatis* in Britain has been dated to ~35 ka (calibrated radiocarbon date on bone; Jacobi et al., 2009), in sediments that pre-date the Late Devensian glaciation at Bishopbriggs, near Glasgow (~NS601722), which falls within the age span ~38–32 ka (or Greenland Interstadials - GIs - 8–5), an interval of relative warmth known in the Netherlands as the 'Denekamp Interstadial' (e.g., Whittington and Hall, 2002). Stuart and Lister (2012) placed this occurrence within GI 8 and the subsequent extinction during Greenland Stadial (GS) 8, before the onset of GI 7. Given the stratigraphic complexity and the uncertainties in the available dating it is indeed possible that the fossiliferous sediments at Whitemoor Haye (and, potentially, also at Coleshill and Quorn) might be exhumed remnants of the Beeston Sand and Gravel or its counterparts, potentially also dating from the 'Denekamp Interstadial'; such an age would indeed be consistent with the older alternative age span for the OSL dates from the 'Lower Sands at Whitemoor Haye (Fig. 6), although Schreve et al. (2013) favoured their radiocarbon dates over this interpretation of the OSL dating and thus assigned the C. *antiquitatis* occurrence at Whitemoor Haye earlier, to GI 12 (equivalent to the Hengelo Interstadial), which they placed at ~47 ka using Greenland ice core data. Pending resolution of the precise cause of the evident mismatches between radiocarbon and OSL dates, and any associated uncertainties in the stratigraphy, we adopt nominal ages of 25 ka and 21 ka for the phases of development of the Trent system depicted in Fig. 4(e) and (f).

At ~20 ka (~21–17 ka, Bateman et al., 2011; ~19–17 ka, Clark et al., 2012), the aforementioned North Sea ice lobe blocked the Humber estuary, impounding the proglacial Lake Humber (Fig. #N4(f)). Clark et al. (2004a) estimated that the surface of Lake Humber was ~25 m O.D., based on sedimentary evidence from the immediate vicinity of the Humber Estuary. They thus inferred that this Lake flooded the Trent valley upstream as far as Nottingham, as well as the former course through the Lincoln Gap and into the Fen Basin. However, their estimates are uncorrected for effects of glacio-isostasy, which would have depressed the crust by many metres in the immediate vicinity of this ice lobe; there is certainly no evidence of Lake Humber sediments in the

Newark or Lincoln areas, let alone as far upstream as Nottingham, where the modern Trent floodplain is also below 25 m O.D. On the other hand, Bridgland et al. (2010) suggested that the diversion of the Trent to the Humber accompanied or followed glacio-isostatic adjustment to deglaciation, as is widely observed across northern England. However, they were unable to indicate a precise timing, and it is anyway evident that the surge of the North Sea ice lobe was synchronous with more general deglaciation. Any suggestion that the youngest phase of aggradation of the Holme Pierrepont Sand and Gravel somehow related to Lake Humber (cf. King, 1966) is, however, contradicted by the dating by Howard et al. (2011), which indicates that aggradation continued long after the demise of this proglacial lake. Indeed, as Bridgland et al. (2014) have suggested, the outlet from Lake Humber may have been southward into the Trent west of Lincoln, as tentatively illustrated in Fig. 4(f), as no other outlet has been recognized, the flow direction having later reversed when the Trent switched to its present Humber course. Indeed, initial southward lake drainage would provide an explanation for the breaching of the former Trent–Humber interfluve, with a new channel incised through relatively resistant Triassic strata, an important event that is otherwise difficult to explain (Straw, 2011; Bridgland et al., 2014).

2.2 Relations between Trent terrace deposits and Middle Pleistocene glaciation

The oldest well-established terrace of the Trent, formed by the Eagle Moor Sand and Gravel (Figs.2 and 3), and incorporating a component of glacial outwash (Brandon and Sumbler, 1988, 1991), was generally assigned prior to the TVPP to the Anglian (MIS 12) (e.g., Lewis, 1999; Maddy, 1999). Howard et al. (2007), however, noted the unexplained absence of Trent deposits representing the interval between MIS 12 and 8, the MIS 11 and 9 interglacials being unknown in the Trent, whereas the penultimate interglacial (MIS 7) is well represented in the basal deposits of the next terrace in the sequence, formed by the Balderton Sand and Gravel and its correlatives in the Lower Witham valley (section 2.1; Fig. 3). Connected with this mismatch in the established pre-TVPP chronostratigraphy was the occurrence of low-level glacigenic deposits beneath the Southrey Terrace deposits (Fig. 3). In particular, the Wragby Till of Straw (1966, 1983) forms a widespread sediment body with its base well below modern sea level, a low disposition seemingly at odds with its supposed Anglian age (White et al., 2010; cf. Lewis, 1999). The resulting proposal by White et al. (2010) that this region was glaciated during MIS 8 is one of the key findings of the TVPP. Regionally, the name 'Wragby Glaciation' has been suggested for this glacial event (Bridgland et al., 2014), based on the presence of the Wragby Till in the Lower Witham valley which, from its disposition relative to the valley and the associated staircase of younger fluvial terraces (Fig. 3; see also section 5.5), provides arguably the clearest evidence for post-Anglian, pre-Devensian glaciation of central England.

Decades ago, the prevailing view was, likewise, that evidence of three glaciations is preserved in the English Midlands, as suggested for example by Posnansky (1960) and King (1966); in the opinion of these authors the first such glaciation was (in modern terminology) during the Anglian and the third during the Devensian, placing the middle glaciation, which Rice (1968) referred to as 'Saalian', in the intervening period. Rice (1968, 1981) indeed envisaged that most of the thick succession of glacial deposits covering Leicestershire, in the catchment of the Middle Trent and its principal right bank tributary, the Soar (section 3.2), was emplaced during this 'Saalian' glaciation; this succession consists of chalky till of northeasterly provenance (from an ice-advance over the Yorkshire or Lincolnshire Wolds), known as the Oadby Till, and quartzite rich till, of northwestern provenance (from an ice-advance across the Cheshire Basin, where it picked up clasts of quartzite from the extensive outcrop of Triassic conglomerate), known as the Thrussington Till. In the view of many authors (e.g., Posnansky, 1960; King, 1966; Rice, 1968) the limits of these ice advances coalesced, creating the observed complex glacial stratigraphy of the area. A notable site is at Huncote quarry, SW of Leicester (SK513982; Fig. 8), where the Oadby Till has been well-exposed above sand and gravel emplaced by the pre-Anglian Bytham River (e.g., Rice, 1981; Lewis, 1989a; see also section 6.1). Westaway (2010b) suggested that the MIS 8 glaciation indeed reached this area (Fig. 8). The local stratigraphy does not preclude an Anglian (MIS 12) age for this chalky till, but likewise also does not preclude an age of MIS 8, nor, indeed, the MIS 10 age proposed by others (e.g., Sumbler, 1995, 2001; Hamblin et al., 2000, 2005; Rose, in Clark et al., 2004b) for the Oadby Till of Leicestershire and many of the other glacial deposits across a wider area of Midland England, including the Wragby Till of Lincolnshire. However, suggestions by these authors that Oadby Till, representing the putative MIS 10 glaciation, persists southward into the catchment of the River Thames, were dismissed by Bridgand and Schreve (2009) and Westaway (2011a), who regarded these southern extremities of till as products of the Anglian glaciation; equivalent till facies can evidently develop as a result of similarly-directed ice advances during different climate cycles (Bridgland et al., 2014).

Figure 8 here: Palaeogeography of central England during the Wragby glaciation

Evidence of glaciation during MIS 8 was also recognized in the Peterborough area, up to ~60 km south of the Witham valley, by Langford (2004) (cf. Langford, 2012; Westaway et al., 2012). This evidence was threefold: (1) outwash deposits at Uffington (reaching to ~30 m O.D., ~TF064089), in the Welland valley east of Stamford; (2) a glacial overflow channel at Southorpe, leading southwards through the present interfluve between the Welland and Nene valleys with a col (~TF085030) at ~25 m O.D.; and (3) glacio-lacustrine deposits at Elton in the Nene valley, thought to indicate that this valley was impounded to the north by ice to create a proglacial lake. This had a surface level at ~30 m O.D. relative to the modern landscape (it would restore to ~15 m O.D. after correction for subsequent uplift; Westaway, 2011a), occupying much of what is now the Fen Basin (Fig. 8). This glacial evidence was assigned by Langford (2004) to a preferred age of MIS 8 (an age confirmed by Langford, 2012) by consideration of the subsequent staircase of terraces of the River Nene; Westaway (2011a) supported this age assignment. West (2009) suggested that the outlet from such a 'Wolstonian' (i.e., post-Anglian, pre-Devensian) proglacial lake lay across the modern drainage divide between the rivers Little Ouse and Waveney, and thus into the southern North Sea, but was noncommital regarding the age of this proglacial drainage system.

Another area where evidence for a post-Anglian–pre-Devensian glaciation has received support is in the Nar Valley of north-west Norfolk, at Tottenhill (Gibbard et al., 1991, 1992, 2009; Lewis and Rose, 1991). The evidence there takes the form of progradational gravels, attributed to a glacial outwash delta, overlying interglacial freshwater and estuarine deposits that have been suggested to date from MIS 11 (e.g., Bowen et al., 1989; Westaway, 2009b; cf. Rowe et al., 1997; Scourse et al., 1999). The Tottenhill gravels have been attributed to MIS 6 (Lewis, 1999), an interpretation seemingly corroborated by an OSL date of ~160 ka (Gibbard et al., 2009), although the numerical data supporting this date were not published, making its reliability impossible to assess. Inset into the Tottenhill gravels are the three terraces of the River Nar; in order of decreasing age, these are the Wormegay, Pentney and Marham terraces, the latter ~1 m above the modern floodplain, the vertical separations being 2–3 m (Ventris, 1986). The cold-climate deposits forming the Pentney terrace contain interbedded temperate-climate sediments, the Pentney Priory Beds, which Ventris (1986) assigned to the Ipswichian (MIS 5e) interglacial. The Pentney and Marham terraces are thus Devensian; the emplacement of the upper part of the Wormegay terrace deposit can be assigned to MIS 6, the downcutting to the level at which the Pentney terrace deposit began to accumulate having occurred under cold-climate conditions, also during MIS 6. The phase of incision between the level of the Tottenhill gravels and the base of the Wormegay terrace deposit thus occurred earlier. The Nar valley terrace stratigraphy would therefore seem to indicate a minimum age of MIS 8 for the Tottenhill glacial outwash delta (Westaway, 2010b; White et al., 2010); indeed, Westaway (2010b) suggested that this outwash delta was located in the same proglacial lake as is also evident from Langford's (2004) interpretation of Peterborough area and West's (2009) interpretation of the Little Ouse and Waveney valleys (cf. Fig. 9). However, it should be noted that this assessment of the extent of MIS 8 glaciation in this region is at odds with other recent interpretations (e.g., Gibbard et al., 2012; Hijma et al., 2012; Langford, 2012); Westaway et al. (2012) provided a summary and rebuttal of these arguments.

Figure 9 here: Palaeogeography of East Anglia during the Wragby glaciation

Furthermore, White et al. (2010) envisaged the ice lobe recognized in the Peterborough area by Langford (2004) as a southward continuation of that which crossed the Lower Witham valley and emplaced the Wragby Till in the latter area. This ice lobe was thus envisaged as temporarily diverting the drainage of the Trent catchment farther upstream, which was unglaciated at the time, through the Southorpe palaeovalley and into the contemporaneous Fen Basin proglacial lake, as depicted schematically in Fig. 8. Given the longitudinal gradient of the river, the resulting longer course, relative to that which developed subsquently (Fig. 4), required the Trent to flow at a greater height in localities farther upstream than was subsequently necessary after the removal of the ice-blockage. Furthermore, it is to be expected that during the Wragby glaciation the Earth's crust was glacio-isostatically depressed, so sediments deposited at the time will have rebounded glacio-isostatically shortly afterwards (later within MIS 8) to greater heights than would otherwise be expected. It follows from these considerations that multiple terraces, formed of fluvio-glacial outwash or cold-climate fluvial deposits, can be expected in the Trent catchment, dating from a succession of times during MIS 8. This led to the development of a new terrace scheme for the Trent by Bridgland et al. (2014), with multiple terraces representing MIS 8. The principal significance of this deduction for the present study is that much of the terrace evidence dating from this time is unrelated to long-timescale uplift, modelling of which is the main aim of the present analysis; however, as will become apparent (section 4), the heights of the lowest terraces assigned to MIS 8 in the Bridgland et al. (2014) scheme are consistent with those of the younger terraces for which direct age control is available and provide the ability to infer uplift within the Trent catchment, thus superseding the previous analysis by Westaway (2007). It is interesting to note that multiple Trent terraces resulting from a single glaciation, as implied here, was an interpretation favoured in the mid-20th century (e.g., Pocock, 1954; Posnansky, 1960).

2.3 Key stratigraphic deductions

The TVPP coincided with BGS investigations offshore from the Wash and the Humber estuary that enable former courses of the Trent, prior to the Holocene sea-level rise, to be traced beyond the present coastline (Tappin et al., 2011). The Wash palaeo-drainage can thus be followed for ~100 km (to a point ~40 km ENE of the mouth of the Humber) along a NNE-trending bathymetric deep known as the Inner Silver Pit (Fig. 10). According to Tappin et al. (2011), this submerged landform probably had a polygenetic origin, involving subglacial erosion and tidal scour as well as fluvial processes. There is no equivalent feature offshore from the Humber, although there is a sediment-filled channel extending roughly ESE from here, intersecting the Inner Silver Pit ~30 km offshore (at ~TA770010 or ~53°31'N, ~0°39'E; Harrison, 1992; Fig. 10). The best candidate for the offshore continuation of the latest Pleistocene Humber (and, therefore, the Trent), this is one of many 'Botney Cut channels' (i.e., channels infilled with sediments of the Botney Cut Formation) now recognized in the SW North Sea (Tappin et al., 2011; Fig. 10). These channels probably developed beneath the Devensian ice sheet, before becoming occupied by fluvial channels upon glacial retreat, since they contain subglacial sediments overlain by latest Pleistocene fluvial gravels (e.g., Cameron et al., 1992; Tappin et al., 2011). Farther north, this combined Humber–Wash drainage probably joined the submerged palaeo-river network between North Yorkshire and the Dogger Bank (north of ~54°17'N ~0°8'E or ~TA450800), reaching depths >60 m below O.D., recognized by Dingle (1970).

Figure 10 here: Offshore map

As was discussed in section 2.1, the Holme Pierrepont Sand and Gravel includes sediments deposited along the former Lincoln Trent course and younger sediments emplaced following the diversion to the Humber. The point of diversion was located near Torksey (~SK838788), ~15 km NW of Lincoln, where deposits assigned to the Holme Pierrepont Sand and Gravel diverge (e.g., Price, 1975; Bridgland et al., 2010), the earlier part of the sediment body continuing downstream into the Lincoln Trent as the 'buried terrace' deposits in Fig. 2. The geometry of the offshore continuations of these two courses (Fig. 10) can be used to investigate the effect that this diversion had on the longitudinal profile of the river at sites upstream of the point of diversion. The distance between Torksey and the point where the courses re-join (Fig. 10) is ~165 km via Lincoln and the Wash and ~145 km via the Humber. Taking the representative longitudinal gradient as 0.2 m km^{-1} in both cases (cf. Fig. 5(a)), and neglecting other considerations (e.g., changes in hydrology) this ~20 km of course shortening would be expected to reduce the river level upstream of the point of diversion by ~4 m. During the emplacement of the Scarle Sand and Gravel, the course of the Lower Trent between Newark and the Lincoln Gap was less strongly looped northward and thus ~5 km shorter than during the early stages of emplacement of the Holme Pierrepont Sand and Gravel (compare Fig. 4(e) and (f); see Bridgland et al., 2010, 2014, for more detailed maps). The latest Pleistocene Humber course was thus ~15 km shorter than the Lincoln course during the emplacement of the Scarle Sand and Gravel, accounting for a ~3 m difference in height, given the nominal longitudinal gradient. In detail, the calculation of this height correction is more complex than stated here, as the longitudinal gradients of both the Lincoln and Humber courses have evidently varied in the downstream direction rather than being 0.2 m km^{-1} throughout; for example, for some distance downstream of the point of diversion the latter course developed with a steeper longitudinal gradient, of ~0.5 m km^{-1} (see section 2.1); allowance for this would increase the magnitude of the correction for downstream course shortening somewhat.

In much of the Middle Trent the Beeston (=Scarle) and Holme Pierrepont terraces differ in height by ~5 m; for example, their heights are 25 m and 30 m O.D. in the Wilford area of Nottingham (Fig. 2). The above calculations indicate that most of this height difference is due to the course diversion from the Lower Witham to the Humber. This explanation for the difference in height between these fluvial terraces is one of the more significant deductions to emerge from the TVPP; beforehand, it was assumed that this height difference reflected uplift throughout a substantial span of time representing their difference in age (e.g., Westaway, 2007). Indeed, in the new Nottingham BGS memoir, A.S. Howard et al. (2009) suggested, on the basis of this substantial height difference, that the Bassingfield Sand and Gravel east of Nottingham might correlate with the Egginton Common and Balderton Sand and Gravel and thus date from MIS 6. However, it is now evident that only a relatively small component of this height difference, after removal of the effect of course diversion, can be attributed to uplift; this is in keeping with the downward revision of the age difference between the youngest sediments within the Beeston and Holme Pierrepont terraces to ~15,000 years (~25 ka versus ~10 ka; see section 2.1, also section 3.3 below).

A second important realization that has emerged from the TVPP investigations is that the Trent catchment remained essentially unaltered between MIS 8, represented by the Etwall Sand and Gravel and its counterparts (Fig. 4(c)), and early MIS 2, represented by the youngest deposits of the Beeston Sand and Gravel and their counterparts (Fig. 4(e)). Other workers (e.g., Gibbard et al., 2012, 2013; Hijma et al., 2012) have proposed that much of the English Midlands, including large parts of the Trent catchment, were glaciated during MIS 6. However, the youngest glacigenic deposits in the interior of this catchment are overlain by cold-climate fluvial deposits now assigned to MIS 8 and by temperate-climate deposits securely dated to MIS 7 (see Fig. 3, also section 2.1; see Bridgland et al., 2014, for more details), indicating that MIS 8 is the youngest plausible age for any such glaciation. As will become clear (section 5), this MIS 8 Wragby glaciation had a dramatic effect on the regional drainage.

A third important realization has concerned the stratigraphic complexity of the fluvioglacial outwash deposits and associated oldest terrace deposits of the Trent in the period immediately after the Wragby glaciation. In particular, it has been established that the previously defined Eagle Moor–Martin Terrace of the Lower Trent has two facets (Fig. 3). Upstream of Lincoln, the height difference between the upper and lower facets of the Eagle Moor Terrace is ~10 m, between the type locality (SK88916825), where the upper facet is at ~33 m O.D.) and a nearby flat in the lower facet (at Monson's Farm, Skellingthorpe: SK92317032). The height difference between facets of the Martin Terrace is ~8 m around Potterhanworth, where the upper facet is mapped (Crofts, 1982; Fig. 3) at up to ~26 m O.D. (~TF067679) and the more widespread lower facet is at ~18 m O.D. (Fig. 2) but seems to be less, ~5 m, around Tattershall, where previous mapping (e.g., Jackson and Issias, 1982; Power and Wild, 1982; BGS, 1995) reveals the two facets at ~20 m and ~15 m O.D. In each of these cases the height difference is rather less than that apparent in the Middle Trent, where separately named terraces are thought to equate with these facets of the Eagle Moor–Martin Formation; thus, for example, at Willington the corresponding Sandiacre and Etwall terraces are ~17 m apart, at 80 m and 63 m O.D., respectively. It is inferred (see Bridgland et al., 2014, for more detailed discussion) that in all of these cases the lowest level represents the 'expected' terrace development late in MIS 8, after any effect of glacio-isostasy had been eradicated from the system, whereas the higher terraces/facets represent conditions shortly after Wragby deglaciation, when the system was glacio-isostatically perturbed. Recognition of the dual-facetted morphology of the Eagle Moor–Martin Terrace means that the Middle Trent and Lower (Lincoln) Trent have the same number of terraces, rather than there being an additional terrace in the former, as was previously thought.

More fragmentary glaciofluvial deposits have also been mapped at higher levels than the Sandiacre–Eagle Moor–Martin (upper) terrace, evidently representing conditions earlier during deglaciation of the Wragby ice sheet. In the Middle Trent south of Derby (at SK387296) the top of the Chellaston gravel, previously mapped as Anglian glaciofluvial deposits (BGS, 2001), reaches ~78 m O.D. (~42 m above the modern river and ~22 m above the late MIS 8 Etwall terrace; Fig. 11). Farther upstream, in the Dove, gravel observed by Stevenson and Mitchell (1955) at Somersal Herbert (SK145355; ~130 m O.D.; ~65 m above the modern river; Fig. #N11) was investigated during the TVPP and found to be relatively rich in flint and Carboniferous sandstone, although otherwise similar to other Trent gravels; its greater height above the modern river again possibly reflects a greater glacio-isostatic effect when the deposit was emplaced. In the Lincoln Trent, downstream of Lincoln, fluvio-glacial gravels are mapped over a wider area and reach up to ~50 m above the modern river (e.g., Crofts, 1982; Jackson and Issias, 1982; Power and Wild, 1982; Fig. 2), although in the more open landscape of this area there is less reason to suppose that they represent former valley-floor situations. Indeed, as Fig. 8 indicates, the Wragby Till ice lobe would have filled the low ground (including the axis of the palaeo-Trent valley, forcing drainage to divert westwards to flow around the ice). Given detailed knowledge of the asynchrony of advances of different lobes of the late Devensian (MIS 2) glaciation of northern England (Clark et al., 2012) there is no reason to suppose that the MIS 8 ice advances in the Middle Trent and Lincolnshire were precisely contemporaneous. The complexity of this proglacial drainage and the anticipated lateral variation in glacio-isostasy preclude the reliable long-distance projection of any longitudinal gradient, making it impossible to determine whether any of the fluvio-glacial gravels higher than the Sandiacre–Eagle Moor–Martin terrace were emplaced as part of a through-flowing Trent that escaped southwards through the Southorpe palaeovalley.

Figure 11 here: Long profile diagram for the Derwent and Middle Trent

The high-level gravel capping Wilford Hill, on the southern outskirts of Nottingham (SK582352), was also studied during the TVPP. This deposit was interpreted by Clayton (1953) as the oldest terrace of the Trent sequence and regarded by him as outwash from what he called the 'Catuvellaunian glaciation', which is equivalent in modern terminology to the Anglian. There was later documentation of ~3 m of red–brown cryoturbated pebbly sand, with pockets of clay and northward-inclined foreset bedding (Charsley, 1989; A.S. Howard *et al*., 2009), its top ~91 m O.D. or almost 70 m above the modern river. Clast-lithological analyses during the TVPP (Bridgland et al., 2014) revealed predominantly quartzose material derived from the Triassic of the Midlands, together with minor exotic constituents, including flint (0.5%) and granitic igneous (0.5%) and single clasts of a banded gneissose rock and *Rhaxella* chert, indicative of a glacial input, an interpretation in agreement with Clayton (1953). It is thus envisaged that the Wilford Hill gravel was emplaced shortly after the Anglian ice sheet retreated from the immediate area.

Further important new evidence arsing from the TVPP relates to high-level fluvial gravels in the vicinity of the Ancaster Gap at Rauceby (White et al., 2007a; Bridgland et al., 2014), previously interpreted as glacial by the BGS (Berridge et al., 1999). The Ancaster Gap is an abandoned valley through the erosion-resistant Lincolnshire Limestone escarpment some 25 km south of Lincoln (Fig. 8). It was regarded by several 20th century authors (e.g., Swinnerton, 1937; Posnansky, 1960; King, 1966) as a former course of the Trent, although prior to the TVPP direct confirmation of this from sedimentary evidence was lacking. The floor of the Ancaster Gap is covered by fluvial gravel of local provenance, known as the Belton Gravel (Berridge et al., 1999) and thought to have been emplaced by former headwaters of the River Slea, to the west of the escarpment, before the upper part of that system was captured by the Lincoln Trent (Bridgland et al., 2007b, 2014; cf. Fig. 4(c) and(d)). However, BGS maps also show patches of higher-level gravel that potentially relate to earlier drainage through the escarpment (their locations are documented in Fig. 54 of Bridgland et al., 2007b). Jukes-Brown (1885) first noted these gravel outliers, and that pits had been opened in several of them, exposing 'gravel of a peculiar character'. On the 1972 geological map (IGS, 1972b), these were mapped as 'Glacial Sand and Gravel'; in the 1996 revision (BGS, 1996) they appear as 'glaciofluvial undifferentiated' and the shape and extent of their outcrops are often markedly different in comparison with the earlier mapping. In the memoir accompanying this revision, Berridge et al. (1999) regarded all such gravel as associated with the Anglian glaciation. The most extensive of these outliers caps a ridge that runs west–east to the west of North Rauceby, at ~79 m O.D. (e.g. at TF009462). There are other outliers near Gelston, some 9 km farther west (SK 913453, SK 917455, and SK 927456; Bridgland et al., 2007b, 2014), with a surface height of ~88 m O.D. (at SK 917456) and other gravels emplaced by a more steeply graded southeast-flowing tributary, which joined the main Ancaster Trent river near Rauceby.

Samples from several of these localities were analysed during the TVPP (Bridgland et al., 2014). For example, at Sentinel Wood, west of North Rauceby (SK997461; see Fig. 55 of Bridgland et al., 2007b, for details), the gravel is dominated by fluvially abraded limestone clasts from the local Jurassic escarpment (~75%), with subordinate orthoquartzite (7%), flint (8%), ironstone (6%), other quartzose material (1%) and Carboniferous chert (1%), the ratio of quartz plus quartzite to Carboniferous chert (QC) being 7. West of the Jurassic limestone escarpment, at Gelston (SK917455), the proportion of local limestone is much less (~1%) and those of orthoquartzite (~73%), other quartzose material (~13%) and Carboniferous chert (~2%) are correspondingly greater, the QC ratio (for many more clasts of these lithologies counted) being 36. There is nothing of unequivocal glacial affinity in these gravels; indeed, clast types that might be expected in deposits derived from east-coast ice lobes, such as *Rhaxella* chert (cf. Bridgland, 1986; 1988), igneous rocks from the Whin Sill, or porphyry from the Cheviots, are missing. The gravels instead contain a mixture of material from the immediate bedrock and the typical components of Trent gravel: quartzose rocks from the Permo-Triassic pebble beds and Carboniferous chert from the southern Pennines. This composition led Bridgland et al. (2007b, 2014) to conclude that an early river, perhaps the originator of the Ancaster Gap, was represented by what they called the 'Rauceby Gravel' (see also section 6.2).

3. Evidence from beyond the TVPP study area

The sequences in the principal tributaries that join the Trent in the Derby–Nottingham area, the Derwent and the Soar, have proved extremely informative. Neither formed part of the core TVPP study area but the evidence they have provided has been key to understanding the evolution of the Trent (Bridgland et al., 2014; see below). The Upper Trent, also beyond the TVPP project area, is likewise important for understanding the development of the system in relation to the Late Devensian glaciation.

3.1 The River Derwent

Depositional terraces are sparse in the Derwent except in its lowest reaches, although a system of mainly rock-cut terraces was described by Waters and Johnson (1958) and, coupled with evidence from karstic sequences in the Peak District (Westaway, 2009c; Bridgland et al., 2014), has formed the basis for reinterpretation. This left-bank tributary has been shown to have existed in pre-Anglian times, having been an important feeder of the Bytham system (e.g., Rice, 1991; Brandon, 1996; Rose et al., 2002; Fig. 4(a)).

The Allenton Terrace of the Derwent, named after a type locality in the southeastern outskirts of Derby and reaching ~8 m above the modern river, has long been recognized; downstream projection establishes it as equivalent to the Beeston Terrace of the main river (Fig. 11). Likewise, the established (e.g., Jones and Stanley, 1974) Borrowash and Ockbrook terraces of the lower Derwent project downstream to the Egginton Common and Etwall terraces of the Trent (Fig. 11). As already noted (section 2.1), fossiliferous sediments, including *Hippopotamus amphibius*, occur within the Allenton terrace deposit (e.g., Arnold-Bemrose and Deeley, 1886; Jones and Stanley, 1974), which have thus been assigned to Ipswichian (MIS 5e) Interglacial (e.g., Bridgland et al., 2007c, 2014), consistent with the age span of the Beeston Sand and Gravel (section 2.1). Upstream of Derby, Gibson and Wedd (1913) were the first to recognize a Derwent terrace deposit at Belper (~SK345475), subsequently classified (e.g., IGS, 1972a; Frost and Smart, 1979) as part of the First Terrace of this river. This reaches ~66 m O.D., or ~6 m above the modern river, and was likewise correlated by these authors with the Beeston Terrace of the Middle Trent (Fig. 11). Bridgland et al. (2014) have proposed that the rock-cut Ambergate Terrace of the middle Derwent, recognized by Waters and Johnson (1958), also correlates with the Allenton Terrace of the lower Derwent and, therefore, with the Beeston Terrace of the Middle Trent (Fig. 11).

Farther upstream of Derby, within the Peak District, Waters and Johnson (1958) also recognized, in order of decreasing height, the Hathersage, Hope and Great Rowsley terraces, as well as the aforementioned Ambergate terrace. However, Bridgland et al. (2014) considered that the highest three of these Derwent terraces had not been correlated on a sound basis; this was in part because they were projected at much lower downstream gradients than the modern river, such that each converges with the modern valley floor at some point along the Derwent. This would imply that rates of fluvial incision in the upper reaches of the Derwent have been negligible during the Pleistocene, which is at odds with the disposition of karstic levels beneath the Carboniferous Limestone uplands flanking this valley; these are well documented (e.g., Ford et al., 1983; Waltham et al., 1997) and, as noted previously (Westaway, 2009c, 2012; Bridgland et al., 2014), indicate valley floor lowering (i.e., fluvial entrenchment) at rates of ~0.1 mm a^{-1} or more in reaches of the Derwent where Waters and Johnson (1958) predict no measureable fluvial incision (cf. Banks et al., 2012). A second reason for calling into question the Waters and Johnson (1958) interpretation is that some of it is based on heights of glacial, rather than fluvial, sediments. For example, an outcrop at ~75 m O.D. in the northern outskirts of Derby (~SK347375) was depicted by Waters and Johnson (1958) as a Derwent counterpart of the 'Hilton Terrace' of the Trent, but is now mapped (BGS DigMap) as till, and therefore cannot be part of any Derwent terrace.

Bridgland et al. (2014) have thus tentatively proposed a new fluvial terrace scheme for the Derwent, reinterpreting the evidence from Waters and Johnson (1958) in terms of upstream counterparts of the established Ockbrook and Borrowash terraces and three newly-designated higher terraces, which they called the High Tor, Matlock and Little Eaton terraces. The highest and oldest, the High Tor Terrace, is thought to be of Anglian age, and will therefore be documented later (section 5.4); the other two are thought to mark times when the river system was glacially or glacio-isostatically perturbed during MIS 8. Supporting sedimentary evidence is found at the southern edge of Derby city centre, where an extensive outcrop of fluvial gravel (~1200 m NW–SE by ~300 m NE–SW) was mapped by Waters and Johnson (1958) and Jones and Stanley (1974) as a Derwent counterpart of the 'Hilton terrace' of the Trent, and is now depicted (BGS DigMap) as 'Glaciofluvial Deposits, Mid Pleistocene'. At its highest point (~SK352353; ~1 km south of the city centre) this gravel reaches 71 m O.D. or ~26 m above the modern river and is thus correlated with the Little Eaton Terrace of the Derwent, defined by Bridgland et al. (2014) on the basis of sites farther upstream and regarded as the Derwent counterpart of the Sandiacre Terrace of the Trent (Fig. 11). Similar projection suggests that the Matlock Terrace is the Derwent counterpart of the Chellaston deposits of the main Trent (Fig. 11).

3.2 The River Soar

The Soar terrace sequence was established by Rice (1968), who noted that it is superimposed onto what he termed the Thurmaston Sand and Gravel, now recognized as part of the Baginton Sand and Gravel of the pre-Anglian Bytham River system (Rice, 1968, 1981, 1991; Rose, 1994; Lee et al., 2004; cf. Gibbard et al., 2013). Rice (1968) thus recognized, in order of increasing height, the Quorndon, Syston, Wanlip, Birstall and Knighton terraces of the Soar (Fig. 12). His nomenclature was retained in output from the TVPP, although a tripartite division of his Birstall Terrace into Upper, Middle and Lower facets is now proposed (Bridgland et al., 2014). The correlation of these Soar terraces with the main Trent is illustrated in Fig. 12, a key tie point being the projection of the Middle Birstall Terrace to the height of the Etwall Terrace of the Trent (Figs 11 and 12). The higher Upper Birstall and Knighton terraces thus mark the perturbed state of the river system during the MIS 8 glaciation, their equivalents in the main Trent being the Sandiacre Terrace and Chellaston deposits (Bridgland et al., 2014; Fig. 12); Rice (1968) indeed inferred that the Knighton terrace formed immediately after his 'Saalian' ice sheet had retreated from the Leicester area. Also within the Soar catchment downstream of Leicester is the East Leake gravel, which yielded Palaeolithic artefacts during an early phase of TVPP field work (White et al., 2007c, 2008; White and White, 2007). From its height and from the fluvio-glacial interpretation arising from earlier BGS mapping, this deposit was thought to date from around the time of the Anglian glaciation, with the implication that the artefacts were pre-Anglian (White et al., 2007c, 2008; White and White, 2007) but on the basis of long-profile projection it is now regarded as part of the Knighton Terrace of the Soar (Fig. 12).

Figure 12 here: Long profile of the Soar terraces

3.3 The Upper Trent

Stevenson and Mitchell (1955), who were responsible for the most recent mapping in the Upper Trent upstream of the Dove confluence area, recognized only two terraces (Figs 7 and 13). In subsequent syntheses (e.g., Posnansky, 1960; King, 1966) the higher of these (the Second Terrace of Stevenson and Mitchell, 1955) has been regarded as the upstream counterpart of the Beeston Terrace, the correlation between the Middle and Upper Trent being illustrated in Fig. 7 (cf. Bridgland et al., 2014).

The maximum Devensian advance of the Irish Sea ice lobe impinged upon the uppermost Trent in the Cannock Chase–Stafford–Stoke-on-Trent area (Figs 1 and 4(e)); ice reached the WSW part of the area depicted in Fig. #N7 and covered much of that in Fig. 13. As Stevenson and Mitchell (1955) showed, around Rugeley and east of Stafford, contemporaneous glacial meltwater channels and spreads of outwash deposits led from the glaciated area into the uppermost Trent at the level of its Second Terrace, which projects downstream into the Beeston Terrace (see above; also Figs 7 and 13). Furthermore, Jowett and Charlesworth (1929) established that, farther north, the same maximum ice advance resulted in the discharge of outwash into the upper reaches of the River Churnet, which thus entered the River Dove near Uttoxeter and the Trent at Willington (Fig. 4(e)). However, it is not possible to demonstrate synchrony between this ice margin and the Dove counterpart of the Beeston Terrace, because the intervening gorge reach of the Churnet lacks fluvial terraces.

In addition, several authors (e.g., Boulton and Worsley, 1965; Rees and Wilson, 1998; Worsley, 2005) have drawn attention to a NE–SW-trending group of moraines in the Kidsgrove–Madeley–Woore area at the SE margin of the Cheshire Plain, which continue westward as the Bar Hill–Whitchurch–Wrexham moraine (Fig. 1). Clark et al. (2012) recognized this as one of several positions where the retreat of the Devensian Irish Sea ice lobe experienced a temporary pause. This moraine forms a ridge locally ~50 m high, making it a far more prominent glacigenic landform than in any locality farther southeast; the glacigenic sediments father northwest also appear significantly better preserved and less weathered (e.g., Boulton and Worsley, 1965; Rees and

Wilson, 1998). This moraine corresponds closely with the drainage divide at the uppermost extremity of the Trent catchment; NE of Woore (Fig. 1) the Trent abuts the Cheshire Basin (i.e., River Weaver/Mersey) drainage, whereas farther southwest it adjoins the catchment of the Tern, which flows SW into the Severn. The evidence provided by this extensive moraine suggests that after surging farther southeast the Late Devensian ice margin retreated before stabilizing for some time at the Bar Hill–Whitchurch–Wrexham moraine. As Rees and Wilson (1998) have discussed, while the ice margin lay along this moraine it was adjoined on its SE side by several proglacial lakes, which fed meltwater channels that led into the uppermost reaches of the Trent system; this transient drainage network was illustrated in their Fig. 40. Thus, for example, at Kidsgrove meltwater channels led into Fowlea Brook, which flows along Cliffe Vale (the Etruria Valley), joining the Trent at Stoke-on-Trent, and around Madeley other meltwater channels interconnected several proglacial lakes, leading ultimately through the Whitmore Gap into Meece Brook and thence into the River Sow. Substantial thicknesses (~20 m) of glaciofluvial sediments are found beneath the present-day land surface along the meltwater channels that lead into the Trent (Rees and Wilson, 1998), with similar deposits up to ~40 m thick in the Sow valley around Stafford (Fig. 13).

Figure 13 here: uppermost Trent terraces

During its retreat to the line of the Bar Hill–Whitchurch–Wrexham moraine, the ice margin temporarily stabilized near Newport, Shropshire (Fig.1), forming a proglacial lake with an outlet across the modern Severn–Trent drainage divide, via a ~99 m O.D. col near Gnosall (SJ806228), through which meltwater was directed via the rivers Penk and Sow into the upper Trent (Whitehead et al., 1927). Whitehead et al. (1927) envisaged that this proglacial Lake Newport was enclosed to the southwest (because ice still abutted the Wrekin Hill) and was contemporaneous with the adjacent Lake Buildwas (Wills, 1924) upstream of Ironbridge Gorge in the Severn valley. However, at this time the outlet from Lake Buildwas was lower, ~90 m O.D.; Whitehead et al. (1927) thus inferred that when the ice subsequently retreated from the Wrekin Hill, lakes Newport and Buildwas coalesced to form the larger Lake Lapworth, with the outlet remaining via Ironbridge Gorge. In this scenario, Lake Lapworth persisted, with this outlet, while the ice margin subsequently stabilized at the line of the Bar Hill–Whitchurch–Wrexham moraine, the resultant prolonged meltwater flow causing the entrenchment of the Ironbridge Gorge to its modern depth. As a result, the ancestral upper Severn was permanently diverted into the lower Severn, the former having previously flowed into the Irish Sea via the Dee or Mersey estuary (e.g., Wills, 1924). However, Whitehead et al. (1927) noted that had the outlet of Lake Newport been lower than that of Lake Buildwas, rather than a few metres higher, the upper Severn might well now form the headwaters of the Trent. The realization, during subsequent decades, that subglacial drainage occurs and can significantly affect the landscape has meant the abandonment of the Lake Lapworth paradigm; Evans et al. (2005) and Worsley (2005) synthesized the evidence from this region in terms of modern glacial concepts, although with only limited comment on implications for the Trent. It is nonetheless evident that there was little difference in height between the sub-glacial and pro-glacial drainage directed towards the Trent and Lower Severn, so a slightly different set of conditions could well have indeed resulted in the permanent diversion of the Upper Severn into the former rather than the latter.

As already noted, when this ice margin was farthest southeast (e.g., in the Rugeley area) outwash entered the Trent at the level of its Second Terrace, equivalent to the Beeston terrace farther downstream (Stevenson and Mitchell, 1955; Fig. 7). However, Fig. 13 indicates that terraces at a similar height persist farther upstream in both the Sow and the uppermost Trent, as Whitehead et al. (1927) and Stevenson and Mitchell (1955) also noted. Although these terraces cannot be traced to the headwaters of any tributary, an outwash deposit at Swynnerton, within ~10 km of the Bar Hill–Whitchurch–Wrexham moraine at Madeley, leads southeastward into the Sow at the level of this upstream continuation of the Beeston terrace (Fig. 13). This evidence suggests that emplacement of the Beeston terrace deposits continued while the margin of the Irish Sea ice lobe was retreating towards the line of the Bar Hill–Whitchurch–Wrexham moraine. In contrast, outwash from the latest stage of the glaciation, when the ice margin presumably lay along this moraine, entered the Trent system during the emplacement of its First or Holme Pierrepont terrace deposit (Stevenson and Mitchell, 1955; see also below, section 3.4). Thus, gravels, described by Stevenson and Mitchell (1955) as fluvio-glacial, occur in the lower reaches of the Sow <2 m above the modern river, forming the First or Holme Pierrepont terrace (Fig. 13). However, Whitehead et al. (1927) noted that in the Sow and Trent valleys it is unclear whether this low terrace is an erosional flat cut into the older 'fluvio-glacial gravels' (i.e., cut into the Upper Trent counterpart of the

Beeston Sand and Gravel; section 3.3) or a younger fluvial or fluvio-glacial terrace deposit. Nonetheless, Stevenson and Mitchell (1955) also reported that gravels forming this low terrace of the Upper Trent contain significant glacial erratics, including igneous and metamorphic rocks from the Lake District and SW Scotland; this was evident, for example, at Shirleywich near Weston upon Trent (~SJ585261) and at Manor Park Quarry near King's Bromley (~SK112168) (Figs 7 and 13). TVPP stone counts (Bridgland et al., 2014) from the Middle and Lower Trent revealed little erratic material of this type; this is unsurprising, since much of it would be less durable than the ubiquitous quartzose rocks and cherts of the Trent gravel, the local derivation of which would have progressively diluted the more exotic material with distance downstream. The incision by the Trent that followed the emplacement of the Beeston Sand and Gravel and its counterparts was dramatic; for example, according to Posnansky (1960), at Holme Pierrepont in the Middle Trent the river incised at this time by ~18 m below the surface of the Bassingfield Terrace, reaching ~9 m below its modern floodplain, with incision of a similar magnitude also evident in the Upper Trent (Fig. 7). This phase of incision may reflect changing hydrological conditions, for example high discharge during the melting of the Irish Sea ice lobe as it retreated to the line of the Bar Hill–Whitchurch–Wrexham moraine, or it might be a consequence of glacio-isostatic rebound accompanying this retreat of the ice margin.

Regarding the chronology, Clark et al. (2012) have proposed that the Irish Sea ice lobe advanced to its maximum extent in the upper Trent catchment after 27±2 ka (although this is based on an unpublished date from an undisclosed location) and remained in roughly the same location until ~23–21 ka, before retreating to leave the Cheshire Plain ice-free by ~19 ka. Morgan (1973) presented a radiocarbon date of ~30.5 ka (which would adjust to ~34.6 ka after calibration) from organic sediment beneath the glacial succession at Four Ashes, Staffordshire, near the SW limit of the ice advance (Fig. 1). Farther afield, for sites in northwest England Telfer et al. (2009) reported OSL dates of 27.8±2.6 and 27.2±2.6 ka that pre-date the advance of the Irish Sea ice lobe and dates of 19.3±2.6 and 16.5±1.7 ka that post-date its retreat. Nonetheless, Chiverrell and Thomas (2010) have emphasized the uncertainty in the chronology of this ice lobe in the Cheshire-Staffordshire area.

Another potential method for constraining the chronology of these changes is possible, by relating them to the GI/GS sequence. That sequence includes an interstadial, GI 3, at about the time of the retreat of the Irish Sea ice lobe to the Bar Hill–Whitchurch–Wrexham moraine; GI 3 is dated to ~28 ka (e.g., centred at 27.95±0.01 ka; Fleitmann et al., 2009). One may thus tentatively suggest that the maximum advance of this ice lobe occurred at ~29 ka, during GS 4, and that emplacement of the cold-climate Holme Pierrepont Sand and Gravel began during GS 3, with the ice margin at the Bar Hill–Whitchurch–Wrexham moraine. Such a scenario might explain the presence of boreal vegetation, representing interstadial conditions, at the base of the Holme Pierrepont Sand and Gravel, as is observed at several sites, such as Coleshill in the Tame (Coope and Sands, 1966) and Quorn in the Soar (Brown et al., 1994) (section 2.1).

4. Uplift modelling

To shed light on the crustal processes occurring in the study region, the terrace staircases of the Lincoln Trent and its tributaries have been modelled at representative localities (Willington, Nottingham, Tattershall, Leicester and Derby). The modelling technique, following Westaway (2001; Westaway et al. 2002), incorporates coupling between surface processes, induced flow in the lower continental crust, and the resulting non-steady-state perturbations to the thermal state of the crust. It calculates the isostatic response to phases of lower-crustal flow forcing (LCFF) induced by cyclic loading of the Earth's surface, caused by growth and decay of local ice sheets or fluctuations in water-loading due to eustatic sea-level variations. The characteristic timescale of the uplift response during a phase of LCFF depends on the parameter $W_i=z_i-z_b$, where z_i is the depth at which the lower-crustal flow is concentrated and z_b is the depth of the base of the brittle upper-crust (e.g. Westaway, 2001), which is taken at the temperature threshold T_b=350 °C (cf. Sibson, 1983). For the parameterization used for the mobile lower-crust, with a temperature-dependent linear viscous rheology and an assumed uniform steady-state geothermal gradient, z_i is ~9/10 of the way between z_b and the depth z_m of the base of the mobile layer, which is typically the Moho (Westaway, 1998).

An alternative technique Westaway (2002a) calculates the non-steady-state isostatic uplift-response in a region that experienced an increase in erosion-rates (or a switch from sedimentation to erosion), assuming that the region adjoins a depocentre where the eroded material is re-deposited. The geothermal gradient is thus perturbed in a predictable manner in both regions, causing the base of the brittle layer to advect (relative to the level of the eroding land surface) upward beneath the sediment source and downward beneath the depocentre. The resulting lateral variation in its depth will create a lateral pressure gradient, which will drive mobile lower crust from beneath the depocentre to beneath the sediment source. The technique is more difficult to apply than the alternative as it requires more model parameters to be constrained; also, it can only represent a single phase of LCFF.

Like in other studies (e.g. Westaway et al. 2004, 2006a), the first technique is applied here as an approximation, it being clear that the uplift modelled is probably due at least in part to erosion and not cyclic surface-loading (Westaway et al. 2002). However, tests by Westaway (2002b) showed that uplift histories predicted by both methods can be very similar to each other, justifying this approximation. As well as requiring fewer model parameters, this technique also allows solutions involving multiple phases of LCFF, each characterized by a different timing t_o and magnitude of the forcing, ΔT_e.

Although ΔT_e for any phase of LCFF is a free parameter, estimated by the fit to data, W_i can be estimated independently from surface heat flow and crustal thickness, as previously noted (e.g. Westaway et al. 2006b). The London Platform (or Midlands Microcraton of BGS workers) extends northward into the present study region and is underlain by (?) Late Proterozoic crust (e.g., Pharaoh et al., 2011) that is ~34 km thick (Chadwick and Pharaoh, 1998) and has low heat flow (~50 mW m^{-2}; e.g., Downing and Gray, 1986; Rollin, 1995; Jackson, 2004), with seismicity persisting to ~25 km depth and thus indicating z_b (e.g. Westaway et al., 2006b). However, gravity studies indicate that much of central England is underlain by a layer of mafic rock up to ~3 km thick, at the base of the crust (Al-Kindi et al., 2003), possibly emplaced during the Palaeogene British Tertiary Igneous Province magmatism. As previously noted (e.g. Westaway, 2001; Bridgland and Westaway 2008a, 2008b; Westaway et al., 2009), after cooling to the ambient temperature, such mafic material will not flow and will thus restrict the vertical extent of the overlying mobile layer. It is thus to be expected that in the present study region W_i is no greater than ~6–7 km. The modelling assumes a standard value for κ, the thermal diffusivity in the lower crust, of 1.2 mm^2 s^{-1}, and appropriate values for u, the geothermal gradient in the lower crust. In the present study, like for other modelling of adjoining regions (Westaway, 2011a; cf. Westaway, 2009b), the only phase of LCFF that will be modelled will be that following the Mid-Pleistocene Revolution, the switch to ~100 ka climate cyclicity, which resulted in increased severity of the cold-climate stages, there being no evidence on which to base any modelling of earlier phases of vertical crustal motion.

The idea that river terraces increase in age with height above the modern floodplain is axiomatic to the understanding of these features as fluvial landforms. More equivocal are the notions (1) that terrace height is a measure of the uplift that has occurred in the catchment since what is now the terrace formed the valley floor and (2) that uplift is, indeed, key to the formation of terraces (Antoine, 1994; Van den Berg, 1996; Maddy, 1997; Antoine et al., 2000; Bridgland, 2000; Maddy et al., 2000; Westaway et al., 2002, 2006b, 2009; cf. Kiden and Törnqvist, 1998). A contrary view has also been expressed (e.g., Hancock and Anderson, 2002; Gibbard and Lewin, 2008; Leeder, 2008; Lewin and Gibbard, 2010; Duller et al., 2012), in which fluvial terrace staircases are seen as potentially forming in response to systematic changes, during the Quaternary, in the hydrology of rivers, such that they develop a tendency to flow with increasingly gentle downstream gradients and, as a result, to create successive valley floors at progressively lower levels in their middle reaches, where fluvial terrace staircases tend to be observed. However, given that river terrace staircases are a worldwide phenomenon (e.g., Bridgland and Westaway, 2008a, 2008b; Westaway et al., 2009), for this to be viable as a general explanation it would require conditions to have changed systematically over the Quaternary, for example with progressively wetter climates (and thus progressively greater tendencies for fluvial incision) during successive climate cycles. No-one has ever presented evidence of any such effect; on the contrary, an abundance of evidence indicates that similar conditions have repeated during corresponding parts of successive climate cycles (e.g., Bridgland, 2000; Bridgland and Westaway, 2008a). Since rivers thus maintain similar hydrological conditions during corresponding parts of successive climate cycles, differences in height of fluvial terrace surfaces, representing successive valley floor levels, must thus in general be unrelated to systematic changes in fluvial hydrology and therefore represent amounts of uplift on the intervening timescale, as previously inferred (e.g., Westaway et al., 2002).

The first requirement for fluvial incision below the levels of river terraces to provide a proxy for uplift is that the terraces are subparallel; this can be seen to be so for the Trent from Fig. 2. The second requirement is that the river has maintained similar hydrological and sediment transport regimes during different phases of fluvial aggradation. This can be readily inferred for the Etwall and Egginton Common terrace deposits and their counterparts, since the tops of these represent equivalent periods within late MIS 8 and late MIS 6 when the Trent catchment is thought to have been unglaciated. Comparison with the height of the Beeston Terrace is, however, more problematic as this evidently marks a time when water and sediment were being fed into the Trent by outwash from the Irish Sea ice lobe (e.g., King, 1966; Fig. 4(e); section 3.3). The validity of any estimate of uplift derived from heights of fluvial terraces also depends on whether or not there have been systematic changes within any particular river system. Previous experience indicates that four critical factors that require consideration in this regard are (a) downstream changes in channel length, (b) channel diversions, (c) effects of glacio-isostasy, and (d) changes in catchment size. In the Trent, issues (a) and (b) are interrelated, as (from section 2.3) the Humber route provides a shorter overall course into the central North Sea than does a course via the Lincoln Gap. Issue (c) mainly affects the interpretation of the multiplicity of fluvial terraces that are inferred to have formed during or immediately after the Wragby glaciation, which (as discussed above; section 2.3) are considered to be affected by glacio-isostasy and other disruptive effects of the Wragby glaciation (Fig. 8). Regarding issue (d), other factors (such as bedrock lithology) being equal, one expects an inverse correlation between catchment area and the longitudinal gradient of a river. As already noted (section 2.2), there is no evidence for any changes in the size of the Trent catchment between MIS 8 and MIS 2.

Regarding changes in downstream length, in section 2.3 it was estimated that the diversion of the Trent from the Lower Witham to the Humber caused a ~15 km shortening of the river relative to the course indicated by the Scarle Sand and Gravel, which would result in a ~3 m height correction for the youngest deposits of the Holme Pierrepont terrace relative to those of the Beeston/Scarle terrace. It would be possible, in principle, to include the Holme Pierrepont terrace in the present uplift modelling, after correction for this effect, but given its small difference in age relative to the Beeston/Scarle terrace it is simpler to omit this Humber Trent terrace from the analysis.

Minor changes in downstream length of the Lincoln Trent are also apparent. In particular, during emplacement of the Eagle Moor and Balderton terrace deposits the Trent followed a more-or-less a straight course between Newark and Lincoln (Fig. 4(c)), the straight-line distance between these localities being ~24 km. In contrast, while the Scarle Sand and Gravel was being deposited the river had adopted a longer course to the north of its former valley (Fig. 4(e)). This added ~8 km to the course length, which for a longitudinal gradient of ~0.2 m km^{-1} (section 2.1) placed the river ~1.6 m higher in upstream localities than it would have been had it maintained the same course between Newark and Lincoln. A 2 m correction (to the nearest metre) should thus be subtracted in upstream localities from the height of the Beeston / Scarle terrace and its tributary counterparts, to account for this effect.

Modelling of each study locality was therefore carried out by fitting solutions of the form generated by the Westaway (2001) method through data points representing the Etwall, Egginton Common, and Beeston terraces of the Trent and their counterparts. Key parameters describing each of these solutions are listed in Table 2. Other evidence will later be compared with the longer-timescale extrapolations of these modelling results, to shed light on changes to the geometry of the Trent course and the limits of its catchment before the MIS 8 Wragby glaciation. It should be noted that this approach to modelling differs considerably from that adopted in the earlier TVPP publication by Westaway (2007). The principal reasons for these differences are, first, the realisation that the diversion to the Humber means that the Holme Pierrepont Terrace is not a good datum from which to measure uplift (see above). Second, is the realisation that the age of the Beeston Terrace (i.e., the time since the incision at the end of the emplacement of the Beeston Sand and Gravel) is much less than the MIS 4 or ~70 ka value adopted by Westaway (2007). By underestimating this age difference relative to the Egginton Common Terrace and its counterparts, Westaway (2007) inferred uplift rates that were too high. Establishing the age of the Beeston Terrace is, indeed, critical to constraining this modelling; hence the detailed discussion of this issue in sections 2.1 and 2.3 (see, also, Fig. 6). For modelling purposes this terrace is thus assigned a nominal age of 25 ka, or early in MIS 2, its youngest feasible age. Its actual age might be slightly older, perhaps 26 ka (Fig. 6) or 28 ka (section 2.3), but such small differences will have little effect on the modelling. Third, Westaway (2007) inferred that the youngest facet of the Etwall/Martin terrace dated from MIS 7b, whereas in the present modelling an age of MIS 8 is preferred. Fourth, in contrast with Westaway (2007), the Sandiacre terrace deposits and their downstream counterparts are no longer considered representative of post-MIS 8 uplift;

as discussed in section 2.2, their height is instead attributed to glacio-isostatic perturbation of the Trent system during the Wragby glaciation. The overall effect of these changes is to significantly reduce the estimated amounts and rates of uplift compared with those suggested by Westaway (2007).

Table 1 here: summary of uplift modelling solutions

4.1 Willington

Willington, at the Dove–Trent confluence, marks the upstream limit of the Middle Trent and of clear resolution of the the Egginton Common and Etwall terraces (Figs 2 and 11). The uplift modelling solution for this location (Fig. 14) is consistent with the dating evidence discussed above, including the MIS 8 age of the Etwall Terrace. It predicts 16 m of uplift since MIS 8 (i.e., since 240 ka), rather less than the ~23 m difference in height between the Etwall Terrace and the modern river, this mismatch being primarily due to the downstream shortening of the river as a result of its latest Pleistocene diversion to the Humber, as discussed above. This solution also indicates 34 m of post-Anglian (post-425 ka) uplift.

Figure 14 here: Willington uplift model

The Sandiacre Terrace locally reaches ~40 m above the modern river, some 17 m above the Etwall Terrace, despite being inferred to be only marginally older; this mismatch being interpreted as an effect of glacio-isostasy during the Wragby glaciation. The Somersal Herbert–Chellaston deposits project to the Willington area at ~92 m O.D. (~52 m above the modern river), some 12 m above the Sandiacre Terrace and 29 m above the Etwall terrace.

4.2 Nottingham

The modelling solution for the Nottingham area (Fig. 15) is similar to that for Willington, but reflects slightly faster uplift. Thus, for example, 19 m of post-MIS 8 uplift is predicted at Nottingham, consistent with the height of the Etwall Terrace to the west of the city, compared with 16 m at Willington. Figure 15 also shows the Sandiacre Terrace and Chellaston deposits projected to the Nottingham area, to demonstrate that their heights exceed post-MIS 8 predictions, for the same reasons as at Willington.

Figure 15 here: Nottingham uplift model

Figure 15 also indicates 40 m of post-Anglian uplift in the Nottingham area. The projected height of the East Leake gravel is shown, to indicate that it roughly corresponds with this estimate of post-Anglian uplift, in accordance with the age originally suggested for this deposit by Westaway (2007) and White et al. (2007b). However, this projected height is based on reconstruction of the sediments as deposited on a Trent floodplain running west–east through East Leake, ~12 km south of Nottingham. Now that it is apparent that these deposits were emplaced by the River Soar rather than the Trent (Fig. 12), it follows that their projection to the Nottingham area should include a correction for the longitudinal gradient of this river over the 12 km distance. If the palaeo-course between East Leake and Nottingham involved 5 km distance at a downstream gradient of ~1 m km^{-1}, characteristic of the lower Soar, then 7 km at the ~0.4 m km^{-1} characteristic of the Middle Trent, the resulting ~8 m correction would bring the surface of this deposit into line with the projection of the Chellaston deposits of the main Trent, suggesting that both represent the same point in time when the Trent valley was glacially perturbed during MIS 8 (Fig. 11).

4.3 Leicester

The solution for Leicester (Fig. 16) indicates 9 m of post-MIS 8 uplift and 19 m of post-MIS 12 uplift, both roughly half that observed at Nottingham (Fig. 15). This southward tapering in uplift is consistent with the upstream convergence of the Soar terraces (Fig. 12) and indicates greater Quaternary crustal stability in the Leicester area than around Nottingham. As already noted, the Upper Birstall and Knighton terraces are correlated with the Sandiacre Terrace and Chellaston deposits of the Middle Trent and inferred to represent perturbation of the river system by the MIS 8 Wragby Glaciation (cf. section 4.1).

Figure 16 here: Leicester uplift model

Westaway (2011a) identified a region of relative crustal stability in the southern part of the East Midlands, around Milton Keynes and Northampton. The Leicester area evidently marks the northern limit of this relatively stable region. As he also noted, Lower Palaeozoic ‘basement’ is present in the shallow subsurface in the Milton Keynes–Northampton area, having escaped deep burial by subsequent sedimentation, providing a further indication of crustal stability. Near Leicester, Precambrian basement crops out in the Charnwood Forest inlier, conforming to a similar pattern and again indicating crustal stability. The same pattern of crustal stability can be detected in earlier geological records hereabouts; in the Early Carboniferous, deposition rates of the Carboniferous Limestone tapered both northward and southward towards an east–west-trending zone, several tens of kilometres wide, extending north–south between Leicester and Northampton, known as the 'Midland Barrier' (e.g., Hains and Horton, 1969). The northern margin of this region of relative stability during the Carboniferous is known in the modern literature as the 'Charnwood–Sproxton High' (Pharaoh et al., 2011). It is thus evident that the observed stability has been a characteristic of this region over geological timescales and that it has influenced both the long-term development of geological structure and much more recent Quaternary landscape evolution. The issue of crustal stability will discussed further, below (section 7), including quantitative estimates of the magnitude of variations in crustal properties across the study region.

4.4 Tattershall

The solution for Tattershall (Fig. 17) has been fitted through the heights of the Martin, Southrey and Tattershall Castle terraces of the Lincoln Trent. The uplift thus indicated since MIS 8 has been 11 m, compared with 19 m at Nottingham (Fig. 15. This gradual eastward decrease in uplift is reflected in the gentle downstream convergence of these terraces (Fig. 2). As Figs 2 and 17 show, at Tattershall the upper facet of the Martin Terrace is ~5 m higher than the lower (main) level of this terrace, suggesting a smaller effect of glacio-isostasy than in localities farther upstream.

Figure 17 here: Tattershall uplift model

4.5 Derby

Figure 18 shows a solution for Derby, fitted to the heights of the Allenton, Borrowash and Ockbrook terraces, the Derwent counterparts of the Beeston, Egginton Common and Etwall terraces of the Middle Trent (Figs 2 and 11). This solution indicates amounts of uplift that are similar to Willington, with 15 m of post-MIS 8 uplift and 32 m of post-Anglian uplift. This is to be expected, given the proximity of Derby and Willington, which are only ~10 km apart (Fig. 1).

Figure 18 here: Derby uplift model

5. Reconstruction of the Trent drainage between MIS 12 and MIS 8

In contrast with the abundant evidence for the course of the River Trent since the MIS 8 Wragby Glaciation (Fig. 4(c) to (f)), only very limited evidence exists for the presence of this river system at earlier times. In this section five key forms of evidence will be used to reconstruct the drainage between MIS 12 and MIS 8, as depicted in Fig. 4(b); the pre-Anglian drainage will then be discussed in section 6. The evidence relates to the disposition of the Hathern Gravel near Loughborough (section 5.1), the depth of the Trent Trench (section 5.2), the Wilford Hill Gravel on the southern outskirts of Nottingham (section 5.3), the oldest (High Tor) terrace of the Derwent upstream of Derby (section 5.4), and the form of the pre-Wragby-glaciation land surface beneath the MIS 8 glacial deposits in localities around Lincoln (section 5.5).

5.1 The Hathern Gravel

The Hathern Gravel (Brandon, 1995; Maddy, 1999) underlies till some 4 km northwest of Loughborough (at SK503214) in the modern Soar catchment. It consists of ~3 m of gravel composed largely of clasts of Carboniferous limestone and chert, the latter including angular cobbles. It overlies bedrock at ~53 m O.D. and is overlain by glacial deposits that have been assigned to the Thrussington Member of the Wolston Formation and, therefore, have been inferred to be Anglian (Maddy, 1999). Lithologically, the Hathern Gravel resembles Derwent terrace gravels in the Peak District (cf. Waters and Johnson, 1958); indeed, Maddy (1999), who classified it as the Hathern Member of the Wolston Formation, suggested emplacement as outwash in the 'proto-Derwent' valley, shortly before the region was overridden by the Anglian ice sheet (cf. Rice, 1991).

Figure 12 shows the disposition of the Hathern Gravel relative to the terraces of the modern River Soar and the pre-Anglian Bytham deposits in the Leicester area (section 3.2). The Hathern Gravel is clearly too low in the landscape for a pre-Anglian age to be tenable, which implies that the overlying till must represent the MIS 8 Wragby glaciation, not the Anglian. This carries the further implication that the associated southward-extended course of the ‘proto-Derwent’ must represent the drainage geometry in this part of the Midlands before the MIS 8 glaciation. Evidently, the modern west–east course of the upper Middle Trent, between Willington and its confluences with the Derwent and Soar to the southeast of Derby (Fig. 1), did not exist at this time; it can instead be envisaged that the proto-Derwent and Soar were confluent a short distance to southeast of Hathern, as schematically depicted in Fig. 4(b). As Fig. 8 shows, the modern course of the upper Middle Trent follows the line of the Elvaston and Swarkestone tunnel valleys, which have hitherto been attributed to the Anglian glaciation, although Carney (2007) discussed the possibility that subglacial erosion in both MIS 12 and 10 might be represented. It now seems likely that these tunnel valleys were formed beneath the edges of an MIS 8 ice lobe and that the modern alignment of the upper Middle Trent was initiated only at that stage (Westaway, 2010b; see Bridgland et al., 2014, for more details).

An additional inference from the Carboniferous-limestone-dominated lithology of the Hathern Gravel is that no equivalent of the modern Upper Trent catchment, which is mostly in areas of Triassic outcrop, can have drained eastward into the pre-Wragby-glaciation river system in the East Midlands; had there been a significant input from the Upper Trent area, the Carboniferous limestone in the Hathern Gravel would have been diluted by quartzose clasts derived from the Triassic. The farthest upstream that the Trent catchment might have extended in the pre-Wragby-glaciation period is thus the River Dove; the Dove drains mainly Carboniferous limestone outcrop and so would have been expected to transport gravel of similar lithology to the Derwent. The Dove is, therefore, tentatively included as part of the pre-Wragby-Glaciation Trent catchment (Fig. 4(b)).

As Fig. 12 also shows, the top of the Hathern Gravel is ~4 m above the Middle Birstall Terrace of the Soar, which represents the valley floor in latest MIS 8 after the glacio-isostatic adjustment following the Wragby Glaciation had ended, at a present-day height of ~52 m O.D. There is, therefore, only a ~6 m fall from this terrace level to the projected ~46 m O.D. height of the Etwall Terrace in the Wilford Hill area of Nottingham, ~15 km away, via the post-Wragby-glaciation drainage. In contrast, for the Hathern Gravel to project to the Nottingham area at the height appropriate to represent the uplift since early MIS 8, it would be required to fall by ~9 m to ~47 m O.D. (Fig. 15). This suggests at first sight that the combined Derwent–Soar river that linked the Hathern and Nottingham areas prior to the Wragby glaciation had a longitudinal gradient of ~0.6 m km^{-1} (i.e., ~9 m / ~15 km), somewhat steeper than the ~0.4 m km^{-1} gradient of the modern Middle Trent (section 4). However, it will be recalled (see above) that uplift at Nottingham (Fig. 15) has been significantly greater than at Leicester (Fig. 16); the estimates of uplift since 250 ka, of 20 m at Nottingham and 10 m at Leicester, can be interpolated to indicate ~16 m at Hathern. Calculation of the longitudinal gradient of the pre-Wragby-glaciation river between Hathern and Nottingham thus requires correction for this ~4 m difference in uplift, which means increasing the estimated fall from ~9 to ~13 m. The resulting gradient, of ~0.9 m km^{-1} (i.e., ~13 m / ~15 km; Fig. 5(b)), therefore exceeds the estimate when no correction is made for differential uplift. This gradient, more than double that of the modern river, would imply a smaller catchment size (see above), consistent with the earlier inference, from the lithology of the Hathern Gravel, that prior to the Wragby Glaciation there was no headwaters system equivalent to the modern Upper Trent.

According to Potter (1966), the modern Trent upstream of Colwick Weir, Nottingham (SK616393), has a catchment area of 7483 km^2. Of this, 1385, 1188, 1019, and 208 km^2 are made up, respectively, by the the Soar, Derwent, Dove and Erewash tributaries of the Middle Trent. Upstream of Shardlow (i.e., upstream of the Derwent, Soar and Erewash confluences) the total area of the Trent catchment was estimated by Potter (1966) as 4412 km^2. Subtracting from this the 1019 km^2 catchment of the Dove and the ~200 km^2 catchment area of local tributaries of the uppermost Middle Trent gives ~3200 km^2 as the area of the modern Upper Trent catchment. If this area drained elsewhere before the Wragby glaciation, then the former catchment upstream of Nottingham was ~40% smaller than at present, in rough agreement with the above estimate that the pre-Wragby-glaciation gradient in the Nottingham area was roughly double its modern value.

The Hathern Gravel can also be projected upstream for the purpose of comparison with the younger Derwent terraces at Derby (Fig. 18). The modern Lower Derwent falls by 16 m, from 46 m to 30 m O.D., in the 12 km between Derby and the Trent, at ~1.3 m km^{-1}. Hathern is located ~20 km downstream of Derby, so, at a gradient of 1.3 m km^{-1}, the top of the Hathern Gravel projects upstream (in a straight line) to Derby at ~82 m O.D. (as indicated by a dashed line in Fig. 11). That is ~16 m above the Ockbrook Terrace of the Derwent, the ‘expected’ fluvial terrace level from MIS 8, post-dating the glacio-isostatic adjustment following the Wragby glaciation. Perhaps ~1 m of this difference is attributable to the small difference in age (probably no more than

15,000 years) between the deposits, which are separated in time by the duration of the Wragby Glaciation. The remaining ~15 m difference therefore indicates the correction required to convert incision to uplift in this reach of the Derwent for the span of time since the emplacement of the Hathern Gravel. However, it should also be noted that, if the palaeo-Dove joined the palaeo-Derwent somewhere SE of Derby but upstream of the Derwent–Soar palaeo-confluence in the period prior to the MIS 8 glaciation, the longitudinal gradient of the combined Derwent–Dove downstream of their confluence would have been somewhat less than 1.3 m km^{-1}, and so the predicted height of the palaeo-Derwent at Derby would have been less than was calculated above. For example, if the gradient of the combined Derwent–Dove was 1.1 m km^{-1} and the Dove-Derwent palaeo-confluence was ~7 km SE of Derby (and, thus, ~13 km NW of Hathern), then the fall between Derby and Hathern can be estimated as ~23 m, indicating that the Hathern Gravel would project to ~79 m O.D. in the former locality (as illustrated by a solid line in Fig. 11). This would imply a height difference relative to the Ockbrook Terrace of ~13 m (Fig. 18), of which ~12 m is attributable to post-Wragby-Glaciation 'rejuvenation' of the river, i.e., incision brought about by the resultant drainage diversions (compare Fig. 4(b) and (c)). This predicted longitudinal profile for the period immediately preceding the Wragby glaciation, with its ~23 m fall between Derby and Hathern, is also depicted in Fig. 5(b); this is illustrated after correction for the ~16 m of subsequent uplift (Fig. 18(a)), so that the contemporaneous height of the river is illustrated, rather than the present-day altitude of its deposits illustrated in Fig. 11.

It is thus inferred that, regardless of the prior geometry of the lowermost Derwent relative to the ancestral Dove, the Wragby Glaciation marked a significant rejuvenation of the Derwent in the Derby area. Two main factors can be invoked in explanation of this rejuvenation event. First, the modern west–east course of the upper Middle Trent was established for the first time, with its separate confluences with the Dove, Derwent and Soar (Fig. 4(c)). The downstream distance between Derby and the newly-established Derwent–Trent confluence, at the relatively steep longitudinal gradient of the Derwent, was much less than the former downstream distance to the pre-Wragby-glaciation Derwent–Soar confluence near Hathern (Fig. 4(b)). Second, the integration of the modern Upper Trent with the Middle Trent resulted in a larger river than had existed previously, with a significantly lower longitudinal gradient.

In contrast with the ~12 m rejuvenation in the Derby area, there is none evident at Nottingham; here, by chance, the river was at almost exactly the same height before the Wragby glaciation as afterward (Fig. 15). Before the glaciation, the height of the river in the Wilford area of Nottingham is estimated as the projected 47 m O.D. height of the Hathern Gravel minus the 20 m of subsequent uplift, or ~27 m O.D. (Fig. N5(b)); as a comparative indication of the height of the post-Wragby-glaciation river in this area, the Beeston Terrace is locally at ~30 m O.D. and indicates a contemporaneous river level of ~28 m O.D., given the ~2 m of uplift estimated since its emplacement (Fig. N5(a)). For comparison, following the downstream course shortening that accompanied the latest Pleistocene diversion of the Trent to the Humber (section 2.3), the modern river in the Wilford area is ~22 m O.D., the typical water level being 20.7 m O.D. at Trent Bridge (SK581382) according to Potter (1966).

It follows from the above analysis, given the steeper pre-Wragby-glaciation longitudinal gradient of the river, that in localities downstream of Nottingham the Trent was lower in the landscape before the Wragby glaciation than it has been since. For example, at East Stoke (~SK755500), at the NE end of the Trent Trench ~25 km downstream of Wilford, the estimated fall of ~23 m would have placed the river, in the absence of subsequent uplift, at circa 4 m O.D. (Fig. 5(b)), almost 10 m below its present level. Downstream of the Trent Trench, the pre-Wragby-glaciation Trent likewise flowed at a lower height than the modern river.

5.2 The depth of the Trent Trench

As already noted, the Trent Trench is most clearly developed between Radcliffe-on-Trent (~SK645397) and East Stoke (~SK755500); the cliffs bounding its southeast flank are typically the most prominent, due to the presence of relatively erosion-resistant Triassic lithologies. Northeast of Radcliffe-on-Trent, the land surface rises to the summit of Gibbet Hill (SK652407) at 67 m O.D. or ~50 m above the modern river. Farther downstream, the land surface abutting the southeast flank of the trench rises even higher, reaching 77 m O.D. at the summit of Old Hill, East Bridgford (~SK702452), and 76 m O.D. at the summit of Toot Hill, Kneeton (~SK704458), these points being ~60 m above the modern river (Fig. 2). The depth of the Trent Trench therefore greatly exceeds the uplift estimated in this area since the MIS 8 Wragby glaciation (Fig. 15) and it can thus be inferred that it formed earlier, presumably as a result of the Anglian glaciation (as Posnansky, 1960, amongst others, previously suggested; see, also, Bridgland and White, 2007; A.S. Howard et al., 2009). Given that this is a confined gorge reach it is clear that the Trent has occupied it since incision began (except, presumably, while the valley was inundated by ice during MIS 8). Moreover, it can also be presumed that the overriding of the region by the Anglian ice sheet resulted in the development of a low-relief landscape formed on glacial deposits, across which the incipient River Trent flowed immediately after the ice receded; the ~60 m maximum depth of the Trent Trench thus provides a lower bound to the fluvial incision since the Anglian. Conversely, the uplift estimated in the Nottingham area since the Anglian is no more than ~40 m (Fig. 15); the mismatch between incision and uplift on this timescale will be explained by Anglian glacio-isostasy (section 5.7).

5.3 The Wilford Hill Gravel

The Wilford Hill Gravel, reaching ~91 m O.D. in the southern outskirts of Nottingham, was documented in section 2.3. The 40 m of post-Anglian uplift at Nottingham (Fig. 15) is rather less than the ~69 m of overall fluvial incision that has occurred below this gravel. This places the river in the vicinity of Wilford Hill at a much greater height than at present after correction for the subsequent uplift; it was ~51 m O.D. (Fig. 5(c)) rather than the present ~22 m O.D. or the immediately pre-Wragby-glaciation ~27 m O.D. (Fig. 5(b)). Nonetheless, projecting the Wilford Hill Gravel downstream at the ~0.9 m km^{-1} gradient estimated above for this reach of the pre-Wragby-Glaciation Trent over the ~16 km to the East Bridgford–Kneeton area would place its surface at ~77 m O.D., in close agreement with the height of the modern land surface in this area (Fig. 2; section 5.2). This comparison indicates that the Wilford Hill Gravel marks the height of the River Trent immediately after the Anglian ice sheet retreated from the Nottingham area and immediately before the entrenchment of the Trent Trench began (Bridgland et al., 2014; section 5.2).

5.4 The High Tor terrace of the Derwent

A rock flat at ~160 m O.D. or ~70 m above the modern River Derwent near Matlock (~SK300590) has long been thought to mark the river level around the time of the Anglian glaciation (e.g., Gibson and Wedd, 1913; Smith et al., 1967). Bridgland et al. (2014) called this the High Tor Terrace, inferring, from its disposition in the landscape, that it dates from immediately after that glaciation. Taking the age of this terrace as MIS 12, a subsequent time-averaged incision rate of ~0.16 mm a^{-1} would be indicated. The Derwent falls by ~44 m, between ~90 and ~46 m O.D., in the ~26 km between Matlock and Derby, at a mean downstream gradient of 1.7 m km^{-1}. High Tor is ~1.5 km south of Matlock, so projection of the High Tor Terrace parallel to the modern river places it at ~119 m O.D. at Derby. This crude projection process neglects any variation in post-Anglian uplift between the Matlock and Derby areas but it allows the High Tor terrace to be tentatively projected into the present study region (Fig. 11), for comparison with the supposedly contemporaneous Wilford Hill Gravel; it also enables plotting of this terrace ~70 m above the uplift-modelling datum used in Fig. 18, for comparison with the other Derwent terraces in the Derby area.

In detail, given the above projection, the High Tor Terrace projects to Derby 68 m above the 51 m O.D. uplift-modelling datum. In contrast, Fig. 18 indicates 32 m of post-Anglian uplift at Derby. The post-Anglian incision at Derby has thus exceeded the contemporaneous uplift by some 36 m, indicating that the river flowed at ~87 m O.D. (i.e., 51 m + ~36 m O.D.) in the immediately post-Anglian period (Fig. N5(c)), well above its present ~46 m O.D. height. Like the Wilford Hill Gravel and the height of the bedrock flanking the Trent Trench, the High Tor Terrace of the Derwent thus provides an indication of the level at which drainage developed immediately after the Anglian glaciation.

5.5 The pre-Wragby-glaciation bedrock surface

It has previously been argued (White et al., 2010) that, given the uplift occurring in the study region, the low present-day height of the top of the bedrock and the base of the overlying glacigenic deposits across much of the region favours a relatively young age for the latter, consistent with a MIS 8 age for the Wragby glaciation. The height of this bedrock–till contact surface also bears upon drainage reconstruction. Straw (2011) has noted that near Sturton-by-Stow (~12 km NW of Lincoln; ~SK890805; Fig. 19) the bedrock–till contact, which is inferred to date from the Wragby Glaciation, occurs at ~8–10 m O.D., indicating, even after allowance for the possibility of subglacial erosion, that the Wragby ice sheet advanced here into an area where the land surface was already relatively low. Given the uplift-modelling solutions presented earlier for Nottingham (Fig. 15) and Tattershall (Fig. 17), some 15 m of uplift can be expected locally since the Wragby glaciation, so in the absence of subglacial erosion the bedrock surface at Sturton-by-Stow would restore to ~-5 m O.D. at the time of this glaciation.

Figure 19 here: Map of the pre-Wragby glaciation landscape

Even lower heights of the pre-Wragby-glaciation land surface are evident in the Witham valley downstream of Lincoln (Fig. 19). For example, MAR boreholes place this surface at a present-day height of ~O.D. near Potterhanworth (~TF065680; Crofts, 1982) and ~-3 m O.D. near Southrey (~TF145662; Jackson, 1982), some 8 km farther ESE. This bedrock–till contact can be taken as approximating the pre-Wragby-glaciation valley floor; indeed, the MAR borehole data map out the floor of a palaeo-valley, several kilometres wide, in a similar location to the modern Lower Witham. Moreover, the longitudinal gradient indicated by this ~3 m fall in ~8 km is ~0.4 m km^{-1}, a low value indicative of a relatively large river system. Furthermore, approximating the uplift since the Wragby glaciation at both Potterhanworth and Southrey as 12 m, the same as at Tattershall, farther southeast (Fig. 17), the valley floor levels restore to ~-12 m and -15 m O.D. at these localities (Fig. N5(b)), rather lower than the bedrock surface at Sturton-by-Stow. The low bedrock surface beneath the Wragby Till in this area was first recognized by Straw (1958) (Fig. 19), although he attributed the palaeo-valley thus revealed to an ancestral River Ancholme, with a catchment restricted to the area east of the Lincolnshire Limestone escarpment (Fig. 19). It is now suggested that it marks the pre-Wragby-glaciation Trent, already established as a river flowing through the Lincoln Gap, an idea that is pursued further in section 5.6.

5.6 Drainage reconstruction immediately before the Wragby Glaciation

In principle, during the span of time immediately before the MIS 8 Wragby glaciation the downstream continuation of the Trent from the northeast end of the Trent Trench might have led into the North Sea via the Humber or via the Lincoln Gap (Fig. 1), there being no other plausible routes for it through the Lincolnshire Limestone escarpment. A route through the Ancaster Gap at this late stage can be excluded, given the variations in uplift across the region, this col being too high for such a route to be plausible (see also section 6.2). As discussed above (section 5.5; Fig. 19), although there is evidence of a low-level base of the Wragby glacial deposits (~-5 m O.D.), which might indicate a pre-Wragby-glaciation col between the Trent Trench and the Humber in the Sturton-by-Stow area, there is no evidence of a contemporaneous river valley leading through this col. The possibility thus exists that this col was created by subglacial erosion beneath the Wragby ice sheet, as Straw (2011) indeed suggested, rather than by a pre-Wragby-glaciation river. On the other hand, there is strong evidence from rockhead data (Fig. 19) that a pre-Wragby-glaciation valley underlies the modern Lower Witham, the low longitudinal gradient of this palaeo-valley requiring a substantial river, which must thus have had an upstream catchment west of the Lincoln Gap.

Figures 11 and 5(b) illustrate the inferred long profile of the Trent immediately prior to the Wragby glaciation. This is consistent with the upstream and downstream projections of the Hathern Gravel discussed in section 5.1 (Fig. 5(b)); it has been reconstructed from the confluence of the ancestral Soar and Derwent near Hathern through the Trent Trench with a gradient of ~0.9 m km^{-1}, decreasing thereafter to 0.4 m km^{-1}. At the latter gradient, the river is predicted to have fallen by some 16 m in the ~40 km distance via the Lincoln Gap between the downstream end of the Trent Trench at East Stoke and Potterhanworth, consistent with its predicted difference in height betwen these localities (+4 m versus -12 m O.D.; Fig. 19).

In the onshore area downstream of Tattershall there is no direct evidence to constrain this pre-Wragby-glaciation drainage. A borehole at Boston (at TF327441; BGS borehole designation TF34SW26) revealed a great thickness of Chalky/Jurassic till (i.e., containing clasts of chalk and flint with a matrix derived from Jurassic clay) overlying Jurassic clay bedrock at >50 m below O.D. (e.g., Whitaker and Jukes-Browne, 1889; Woodward, 1904), although the borehole record (searchable online using the borehole ID, at http://mapapps.bgs.ac.uk/GeoRecords/GeoRecords.html) indicates uncertainty as to the precise depth of the contact between till and underlying Jurassic clay. However, this places the top of the bedrock >30 m below the long-profile projection in Fig. #N5(b), after allowance for subsequent uplift (cf. Fig. 17), indicating significant subglacial erosion (presumably, during the Wragby Glaciation) in this locality. Farther downstream, the initial development of the Wash is thought to have been a consequence of Anglian subglacial erosion; ice flowing southwestward through the area now occupied by this embayment breached what had previously been a ridge in the Chalk, forming a southeastward continuation of the Lincolnshire Wolds and the NE valley-side of the Bytham system (e.g., Perrin et al., 1979; Clayton, 2000). The pre-Wragby-glaciation Trent would have entered the North Sea through this breach; if the ~0.4 m km^{-1} longitudinal gradient persisted downstream, this river is predicted to have flowed at a level of ~-40 m O.D. at the present mouth of the Wash in the period immediately before the Wragby glaciation (Fig. 5(b)). Gallois (1979) recognized evidence of a former drainage system beneath the present-day Wash, his 'Wash River', which indeed entered the North Sea through this breach. From Wingfield et al. (1978), the top of the Chalk bedrock is as low as -48 m O.D. along the axis of the breach (~TF650530), which is ~1.5 km wide at the level of -30 m (this would restore to ~-40 m O.D. after correction for post-Wragby-glaciation uplift; cf. Fig. 17). Furthermore, parts of the 'Wash River' channel system recognized by Gallois (1979) are infilled by Chalky/Jurassic till; like at Boston, these can now be assigned to the Wragby glaciation. The recognition by Gallois (1979) of palaeo-drainage tens of metres below O.D. in and around the Wash is thus consistent with the reconstruction of the long profile of the ancestral Trent in Fig. 5(b).

Interpolating between the predictions in Figs. 15 and 17, some 15 m of uplift can be expected since MIS 8 in the Lower Trent between Newark and Lincoln, so the predicted height of the pre-Wragby-glaciation long profile in Fig. 5(b) of, say, -3 m O.D. at South Scarle equates to a present-day height of ~12 m O.D., similar to the height of the much younger Scarle Sand and Gravel (Fig. 5(a)). It is therefore to be expected that any evidence of the pre-Wragby-glaciation drainage in this reach has been obliterated by later fluvial erosion, in contrast with the reach downstream of Lincoln where the river flowed lower within the landscape before MIS 8 than at any time since. Nonetheless, Fig. 5(b) implies that during pre-Wragby-glaciation interglacials (i.e., during MIS 9, and possibly also during MIS 11), when the sea surface was near its present level, the ancestral Trent would have formed a large estuary extending inland through the Lincoln Gap (Fig. 20), similar to the modern Humber (Fig. 1). It is indeed predicted from the long profile in Fig. 5(b) that this ancestral river passed through sea level maybe 4 km northeast of the modern location of Newark (Fig. 1), which thus adjoined the head of the temperate-stage pre-Wragby-glaciation estuary of the ancestral Trent. The Woodston Beds and Nar Valley Clay evidently marked offshoots that existed during MIS 11, when Fen Basin rivers flowed into the same main estuary, as illustrated schematically in Fig. 20.

Figure 20 here: pre-Wragby glaciation interglacial palaeogeography

This pre-Wragby-glaciation reconstruction can also be used to infer the changes to the Trent caused by the MIS 8 glaciation. First, the pre-Wragby-glaciation valley both upstream and downstream of Lincoln was infilled with glacigenic deposits. Downstream of Lincoln, it was filled with till (see above), whereas upstream of the Jurassic escarpent, in the vicinity of Whisby and Doddington (~SK910680 to SK910700), widespread laminated clays record the presence of the Skellingthorpe proglacial lake (e.g., Jackson, 1977; White et al., 2010). These clays reach ~24 m O.D. and are overlain by ~4–5 m of the outwash-charged Eagle Moor Sand and Gravel (Fig. 3). As White et al. (2010) discussed, a probable explanation for the presence of this proglacial lake is that during part of the Wragby Glaciation the Lincoln Gap was blocked by ice on its eastern side.

The second major change in the Trent drainage arising from the Wragby glaciation was the integration of the modern Upper Trent with the rest of the system (compare Fig. 4(b) and (c)), significantly increasing the catchment size. Being a much larger river than its pre-Wragby counterpart, the post-MIS 8 Trent had a much lower downstream gradient (see above). As previously discussed, the level of the river was little changed in the Nottingham area, but farther downstream the gentler post-MIS 8 longitudinal gradient would have raised it above its pre-Wragby-glaciation level, an effect that was no doubt facilitated by the aforementioned plugging of the pre-existing valley with Wragby Till. Assuming longitudinal gradients of 0.4 m km^{-1} within and above the Trent Trench and ~0.2 m km^{-1} farther downstream, the post-Wragby-glaciation Trent is thus predicted to have flowed at heights of ~4 m O.D. at Tattershall and -7 m O.D. at the mouth of the Wash (Fig. 5(a)), compared with -20 m and -40 m O.D. for its pre-Wragby-glaciation counterpart (Fig. 5(b)). The post-Wragby-glaciation Trent is thus predicted to have passed through sea level a few kilometres downstream of Boston, not far from the modern coastline; the estuary extending far inland, predicted to have developed at times of high sea level during pre-MIS 8 interglacials (Fig. 20), therefore did not reform during later interglacials.

5.7 Immediately post-Anglian drainage reconstruction
The regional drainage can also be reconstructed for the time immediately after the Anglian glaciation. The evidence (from the heights of the High Tor Terrace of the Derwent, the Wilford Hill Gravel, and the bedrock flanking the Trent Trench) indicates (after correction for subsequent uplift) that the drainage was 20–30 m higher at the end of the Anglian glaciation compared with immediately before the Wragby Till ice advance (Fig. 5(c)). Such an effect might have resulted from glacio-isostasy, with heights exaggerated because, when the rivers flowed at the levels indicated, the adjacent crust was still depressed by the Anglian ice sheet; the fluvial sediments and landforms were thus raised to higher levels a short time afterwards, after the effects of the ice load were fully removed, by glacio-isostatic rebound

Earlier discussion established that at Anglian deglaciation the Trent at Wilford Hill flowed at a level of ~51 m O.D., some 24 m above its estimated level immediately prior to the Wragby glaciation. Assuming that the same ~0.9 m km^{-1} downstream gradient was maintained through the Trent Trench at both times, the immediately post-Anglian river level at East Stoke would have been ~28 m O.D. (Fig. 5(c)). Farther upstream, the immediately post-Anglian High Tor Terrace, at a projected height of ~119 m O.D. at Derby (section 5.4), restores to ~87 m O.D. (Fig. 5(c)) after correction for 32 m of uplift (Fig. 18). A ~36 m fall can thus be estimated between this and the Wilford Hill Gravel at Nottingham (Fig. 5(c)), equal to the ~36 m fall deduced from the projections of the Hathern Gravel (Fig. 5(b)) but greater than the 23 m fall of the modern river between these localities. Notwithstanding the uncertainties involved, which allow the possibility of modest changes to the drainage in this region between the immediately post-Anglian and immediately pre-Wragby-glaciation intervals, these projections suggest that the geometry of the rivers in this region remained stable throughout this span of time, with the Derwent–Soar palaeoconfluence a short distance east or southeast of Hathern (Fig. 4(b)).

6. Pre-Anglian drainage
The pre-Anglian drainage of the study region (Fig. 4(a)) will be reconstructed here, using the configuration of the Thurmaston Sand and Gravel and other Bytham River deposits, and of the Rauceby Gravel in the vicinity of the Ancaster Gap. Other evidence for the pre-Anglian drainage is provided by the disposition of karstic levels and bedrock flats in the Peak District, which indicates that the entrenched valleys of the rivers Derwent and Dove within the Carboniferous Limestone uplands of the Peak District are of considerable antiquity, probably dating back to the Mid-Pliocene (e.g., Westaway, 2009; Bridgland et al. (2014).

6.1 The Thurmaston Sand and Gravel
Quartzose sands and gravels have long been recognized in the Leicester area, and were regarded by Shotton (1953) as evidence for his 'Proto-Soar', which he envisaged as a river with headwaters well to the south and west of the modern Soar catchment but with lower reaches following the modern valley towards Nottingham. These deposits were subsequently studied in detail by Rice (1968, 1981, 1991), who applied the name Thurmaston Sand and Gravel, after a type locality in a now disused quarry at Thurmaston (SK615101) that revealed ~2 m of gravel overlain by ~6 m of sand (its top at ~68 m O.D.), overlain in turn by glacigenic sediments. Rice (1991) also described a nearby site at Wanlip Hill (~SK590110) showing ~2 m of gravel overlain by ~6 m of sand (its top at ~72 m O.D.), again beneath glacigenic sediments. Equivalent deposits farther upstream occur the complex of quarries at Huncote (SK 513982), where ~5 m of gravel is overlain by ~6 m of sand, its top ~78 m O.D., then by glacigenic deposits (e.g., Rice, 1981; Lewis, 1989a; Bridgland et al., 2014). These sites are plotted in Fig. 12; the ~6 m fall in ~16 km between Huncote and Wanlip Hill indicates a gradient of ~0.38 m km^{-1}, much less than the longitudinal gradient of the modern River Soar (Fig. 12), indicating emplacement by a much larger river.

There is a complex history of stratigraphic nomenclature for these deposits, as Bridgland et al. (2014) have discussed. Most recently, Maddy (1999) assigned them to the Baginton Formation: the sand as the Brandon Member (type locality near Coventry: SP384763) and the gravel as the Thurmaston Member. The Baginton Formation also includes temperate-climate fossiliferous deposits known as the Waverley Wood Member (type locality SP365715), underlying the Thurmaston Member. Amino-acid dating places the Waverley Wood Member in the latest Cromerian, probably in MIS 13a (e.g., Parfitt et al., 2005; Westaway, 2009, 2010c; Penkman et al., 2011), suggesting that the Thurmaston and Brandon members were emplaced in MIS 12, shortly before the region was overridden by the Anglian ice sheet.

Following Rose (1987, 1989), Shotton's (1953) 'Proto-Soar' concept has been superseded by the hypothesis that the Baginton Formation represents the upper reaches of the Bytham River, which flowed eastward from the Leicester area along the Wreake Valley, in the opposite direction to this modern Soar tributary, through the Lincolnshire/Leicestershire Wolds at Castle Bytham, across the Fen Basin to Shouldham, then southward to Mildenhall and eastward across East Anglia to the North Sea coast near Lowestoft, as illustrated in Fig. 4(a). This is consistent with the evidence, from its long profile, that the Baginton Formation was emplaced by a much larger river than exists in the Leicester area at present.

Gibbard et al. (2013) have proposed that Rose's (1989) interpretation of the Bytham River is incorrect and that the Thurmaston Member was emplaced by a late Middle Pleistocene river that flowed northeastward from Castle Bytham across the Fen Basin near Spalding and then along the Inner Silver Pit (Fig. 1). In the view of these authors, this river system was disrupted by a hypothetical 'Wolstonian' glaciation, during MIS 6. This proposal is contradicted by an abundance of evidence, as discussed earlier in the present account. Furthermore, the Thurmaston Sand and Gravel is much too high in the landscape to have any plausible connection with the regional drainage prior to the MIS 8 Wragby Glaciation. For example, the Gibbard et al. (2013) interpretation projects these deposits to heights of ~55 m O.D. at Spalding and ~50 m O.D. in the SW part of the Wash, east of Boston, where it would have been confluent with the pre-Wragby-glaciation Trent (Figs 4(b) and 5(b)). However, the evidence from the Lower Witham valley (Figs 5(b) and 19; sections 5.5 and 5.6) indicates that the pre-Wragby-glaciation river level in the Boston area was ~20 m below O.D. and would restore to ~10 m below O.D. (not ~50 m above O.D.) after subsequent uplift. Such considerations mean that any suggestion that the Thurmaston Sand and Gravel was emplaced by drainage through what is now the Wash can be discounted; the combination of the high elevation and low longitudinal gradient of these sediments means that they can only have been emplaced by a river flowing eastward across East Anglia, as Rose (1989) proposed.

The surface of the Thurmaston Sand and Gravel projects upstream from 72 m O.D. at Wanlip Hill (Rice, 1991) to ~74 m O.D. at the Leicester uplift-modelling locality, or ~20 m above the modern River Soar (Fig. 12). The 21 m of post-450 ka uplift deduced locally (Fig. 16) thus roughly matches the height of this deposit above the modern river, indicating that the Bytham River flowed locally at a height of ~53 m O.D., similar to that of the modern River Soar at Leicester. However, this agreement is fortuitous, having arisen because the combination of lower longitudinal gradient and greater downstream length of the Bytham River leads it to intersect with the shorter but steeper modern Trent and Soar levels.

Downstream variations in lithological composition are important for determining the overall dimensions of the Bytham catchment; new stone counts determined as part of the TVPP have contributed to this (Bridgland et al., 2014). In particular, numerous authors have measured the QC ratio, albeit for different size fractions. At Pools Farm Pit, Brandon (~SP388762) this ratio is ≥21 (Maddy, 1989), whereas at Waverley Wood it is 26 (Bridgland et al., 2014), likewise reflecting the very low concentrations of Carboniferous chert in the gravels. There is no outcrop of Carboniferous Limestone (the original source rock for the chert) in this vicinity; however, beds of conglomerate rich in Carboniferous chert have long been recognized within the Upper Carboniferous succession of the Warwickshire coalfield (e.g., Shotton, 1927, 1933). In contrast, at Huncote the QC ratio has been determined as 10, by both Lewis (1989a) and Bridgland et al. (2014). This implies an input of additional chert, presumably by a previously documented left-bank tributary, the Hinckley River, which flowed south of Charnwood Forest, as schematically illustrated in Fig. 4(a). The buried valley of the Hinckley River can be traced northwestward across SW Leicestershire for ~30 km from its palaeo-confluence with the Bytham (e.g., Douglas, 1980; Worssam and Old, 1988; Bridge et al., 1998; Bridgland et al., 2014), joining the latter upstream of Huncote (Fig. 4(a)). It can thus be inferred that the Hinckley River was a pre-Anglian downstream continuation of the ancestral River Dove, the Carboniferous chert at Huncote having been transported from the Peak District. A further change in composition occurs to the northeast of Leicester, starting at the aforementioned Wanlip Hill site; although some localities northeast of here have QC ratios as high as ~10 (i.e., similar to Huncote), at others it is as low as 3. As Rice (1991) first noted, this increase in the concentration of Carboniferous chert is due to mixing of Bytham River gravels with chert-rich input from the ancestral River Derwent (the 'proto-Derwent' or 'Derby River' of A.S. Howard et al., 2009), which drains the eastern part of the Peak District (Fig. 4(a)). Farther downstream relatively low QC ratios are maintained, for example at Red Barn pit, Castle Bytham (~SK980198), where they are in the range ~4–12 (Rose, 1989) and in coarser fractions can be as low as 2 (Bridgland et al., 2014). The characteristic Bytham clast-lithological signature, relatively rich in Carboniferous chert, can also be readily recognized in East Anglia and is an important provenance indicator (e.g., Lewis and Bridgland, 1991; Rose et al., 2001, 2002; Westaway, 2011b).

The early Middle Pleistocene gravel filling a palaeovalley beneath Anglian till in the eastern Leicestershire Wolds, on the route envisaged for the Bytham River (*sensu* Rose, 1994; Lee et al., 2004), reaches 66 m O.D. at Castle Bytham (~SK998184; Rose, 1989) and ~61 m O.D. at Witham-on-the-Hill (~TF030177; Lewis, 1989b), ~3.3 km farther downstream. The local present-day eastward gradient of the surface of these sediments is thus quite steep, being ~5 m / ~3.3 km or ~1.5 m km^{-1}, which is readily apparent from long-profile reconstructions, such as that by Lee et al. (2004). This local steep gradient is presumably a consequence of constriction of the valley through the erosion-resistant Lincolnshire Limestone, the valley having much gentler gradients elsewhere (see above). The Castle Bytham–Witham-on-the-Hill area is roughly halfway between Rauceby, where the uplift since ~450 ka can be calculated as 33 m (see below), and Peterborough, where modelling by Westaway (2011a) indicates 35 m of uplift during this interval. By interpolation, one may estimate ~34 m of post-450 ka uplift in the Castle Bytham–Witham-on-the-Hill area, such that the upper surface of the fluvial sediment was emplaced at a height, after correction for subsequent uplift, of ~27–32 m O.D.

Traced upstream along the Wreake valley towards Leicester, the Bytham deposits pass laterally into the Thurmaston Sand and Gravel, although there is a gentle decline of the sediment body in this direction (Fig. 11), as noticed by Rice (1991). Conversely, eastward palaeoflow is confirmed by the orientation of cross bedding and by consideration of the clast content of the gravel (e.g., Lewis, 1989c; Rice, 1991). It is now possible to explain the eastward (downstream) rise in the Thurmaston–Bytham Gravel as a result of back-tilting caused by differential uplift between the relatively stable crustal region around Leicester and the more rapidly uplifting area to the east. At the time of emplacement, the immediately pre-Anglian Bytham valley-floor deposits were at an estimated (reconstructed) height of ~53 m O.D. at Leicester and ~32 m O.D. at Castle Bytham, ~50 km farther downstream (see above). A reconstructed fall of ~21 m is thus indicated along this reach, indicating a mean downstream gradient of ~0.42 m km^{-1}, in rough agreement with other estimates for the Leicester area. At Mildenhall, ~100 km downstream of Castle Bytham, the Bytham valley floor is reconstructed (with subsequent uplift removed) as close to O.D. (Westaway, 2009b), indicating an average downstream gradient between Witham-on-the-Hill and Mildenhall of ~0.28 m km^{-1}.

Farther upstream, the difference between the ~34 m of post-450 ka uplift estimated at Castle Bytham and the 21 m estimated at Leicester (Fig. 16) indicates overall net post-Anglian westward back-tilting of the landscape by ~13 m / ~50 km or ~0.26 m km^{-1}. Moreover, as noted above, Rice (1968, 1991) reported the upper surface of the Thurmaston Sand and Gravel in the vicinity of the Wreake-Soar confluence at 68 m O.D. at Thurmaston and at up to 72 m O.D. at nearby Wanlip Hill. In contrast, some 15 km farther ENE upstream along the modern Wreake valley around Great Dalby (near Melton Mowbray), Rice (1991) reported that the same deposits reach ~75 m O.D. The present-day decline of these deposits between Great Dalby and Castle Bytham, by ~9 m in ~25 km or ~0.36 m km^{-1}, roughly matches the reconstructed downstream gradient of the Bytham River, indicating that the Great Dalby area has uplifted by roughly the same distance since the Anglian as Castle Bytham. Had the same uplift history extended farther west, one would expect the top of the deposits in the Thurmaston–Wanlip area to be ~6 m higher (i.e., ~0.4 m km^{-1} × ~15 km) than at Great Dalby rather than the observed ~3 m lower. It can thus be estimated that the Thurmaston–Wanlip area has experienced post-450 ka uplift by ~9 m less than Great Dalby and Castle Bytham, or by ~25 m. Linear interpolation of the estimates of post-450 ka uplift between Leicester (21 m; Fig. 16) and Nottingham (44 m; Fig. 15) would predict ~23 m of uplift since ~450 ka in the Thurmaston–Wanlip area, in good agreement. The back-tilting evident between this area and Great Dalby, by ~9 m in ~15 km distance, or by ~0.6 m km^{-1}, has more than cancelled out the original ~0.4 m km^{-1} downstream gradient of the Bytham River, resulting in the present-day westward tilt of the sediments by ~3 m in ~15 km, or ~0.2 m km^{-1}, that perplexed Rice (1991). Conversely, as already noted, upstream of the Thurmaston–Wanlip area, the surface of the Thurmaston Sand and Gravel rises upstream at almost 0.4 m km^{-1}. Thus, in this reach, the deposits are not back-tilted; they have evidently experienced a uniform, low, amount of post-Anglian uplift, reflecting the crustal stability of the Leicester area.

Finally, the disposition of the Thurmaston Sand and Gravel can be used to infer the predicted height of the valley floor of the River Derwent in the immediately pre-Anglian period, enabling the idealised long profile illustrated in Fig. 11 to be constructed. It is evident (section 4.3) that the Leicester area has uplifted relatively slowly, by 21 m since ~450 ka. The Bytham River flowed at an estimated restored height (i.e., with the effects of subsequent uplift deducted) of ~53 m O.D. at Leicester and, taking account of its ~0.4 m km^{-1} longitudinal gradient, ~51 m O.D. at the contemporaneous Derwent confluence in the vicinity of Wanlip Hill, where the top of the pre-Anglian fluvial sand is now 72 m O.D. (see above). Assuming, as before, a longitudinal gradient of 1.3 m km^{-1} and neglecting for the time being any variation in uplift, the pre-Anglian Derwent projects to Derby, which is ~36 km upstream of the Bytham confluence along the inferred palaeo-Derwent course (Fig. 4(a)), at a contemporaneous height of ~98 m O.D. The uplift modelling (Table 2 6.2. The Rauceby Gravel) indicates that since 450 ka the Derby area has uplifted by 35 m; adding this gives the height at Derby as ~133 m O.D., at which any immediately pre-Anglian deposits of the Derwent would be anticipated at the present day (Fig. 11). Likewise, at Hathern, ~16 m upstream of the Derwent–Bytham palaeoconfluence, immediately pre-Anglian sediments are expected at a height of ~72 m + 16 km × 1.3 m km^{-1} + 35 m - 21 m or ~107 m O.D., if it is assumed (as in section 5.1) that this locality has experienced the same uplift history as Derby (Fig. 11). At both Hathern and Derby, these predicted heights are well above the highest points in the present-day landscape; therefore, unlike at Leicester, there is no possibility of observing in situ pre-Anglian fluvial sediment at either locality. Nonetheless, this predicted ~133 m O.D. present-day height of the immediately pre-Anglian River Derwent at Derby is some 14 m above the 119 m O.D. present-day height expected for the immediately post-Anglian river, based on downstream projection of the High Tor terrace (section 5.4) and some 54 m above the projected height at Derby of the Hathern Gravel (section 5.1). Allowing for the predictions of uplift by ~3 m between 450 and 425 ka and by ~19 m between 450 and 250 ka (Table 2), it can be seen that the Derwent at Derby experienced incision in excess of uplift by ~11 m between ~450 and ~425 ka and by ~35 m between ~450 and ~250 ka. The overall ~35 m rejuvenation was evidently the result of the major downstream course change that followed the Anglian glaciation (cf. Figs 4(a) and (b)); however, as was discussed in section 5.7, ~24 m of this (represented by the difference between the long profiles in Figs. 5(b) and (c)) was temporarily 'masked' at ~425 ka by the transient effect of glacio-isostasy immediately after the Anglian deglaciation. These effects were distinct from the later 12 m rejuvenation of the Derwent at Derby noted in section 5.1, as a result of the various changes that followed the MIS 8 Wragby Glaciation (cf. Figs 4(b) and (c)).

6.2 The Rauceby Gravel

As discussed in section 2.3, the newly-defined Rauceby Gravel is believed to have been emplaced by the river that initiated the Ancaster Gap, rather than the glacial interpretation of these deposits advocated by the BGS (Bridgland et al., 2007b, 2014; contra Berridge et al., 1999). Earlier discussion (section 5.1) established that before the Anglian glaciation the River Derwent extended southeastward of its modern Trent confluence and flowed into the Bytham River near the modern Wreake-Soar confluence to the northeast of Leicester (Fig. 4(a)). Any contemporaneous 'Ancaster Trent' river could therefore not have had headwaters farther west than the Nottingham area, although farther north its catchment might have reached the eastern flank of the Peak District. The catchment upstream of the Ancaster Gap might well have been substantial, therefore, possibly as much as ~3000 km^2 (approximating a rectangle ~50 km east–west by ~60 km north–south). A catchment of this order would exceed those of modern Trent tributaries such as the Soar (area 1385 km^2; Potter, 1966) and would indeed be comparable with the modern Upper Trent catchment (~3200 km^2; section 5.1).

To reconstruct this pre-Anglian river system, the heights of deposits must be corrected for post-Anglian uplift. The uplift since ~450 ka, the approximate time when this former river system was disrupted by the advancing Anglian ice sheet, has previously been estimated as 26 m at Tattershall (Fig. 17) and 44 m at Nottingham (Fig. 15). The Ancaster Gap is roughly 30% of the way between these localities, so by interpolation some 31 m of post-Anglian uplift may be estimated. Restoring this uplift places the deposits in the Rauceby area, which are now ~80 m O.D., at ~49 m O.D. around the start of the Anglian glaciation. For comparison, the deposits of the Bytham River at Castle Bytham (~25 km south of the Rauceby area; section 6.1), now at 66 m O.D., restore to a pre-Anglian height of ~32 m O.D. Furthermore, the 18 m of tapering in post-450 ka uplift thus estimated, across the ~60 km straight-line distance between Nottingham and Tattershall, indicates eastward tilting by ~0.3 m km^{-1}. The ~1 m km^{-1} present-day slope of the Rauceby Gravel thus requires correction by this amount, to give an estimate of the contemporaneous longitudinal gradient of the Ancaster Trent river, of ~0.7 m km^{-1}. This value, smaller than the estimated downstream gradient of the post-Anglian Trent in the Nottingham area (section 2), is nonetheless indicative of a substantial upstream catchment.

Rose et al. (2001. 2002) proposed that the drainage through the Ancaster Gap prior to the Anglian glaciation was directed eastward, into the Cromer area of North Norfolk, ~120 km away. The Cromer area has not uplifted or subsided significantly since the early Middle Pleistocene (Westaway, 2009b), so the level of the 'Ancaster River' would have declined over this distance by ~49 m, indicating a mean downstream gradient of ~0.4 m km^{-1}. This is rather lower than would be expected for the anticipated size of river, thus calling the interpretation into question. An alternative interpretation for the pre-Anglian 'Ancaster Trent' is for it to have flowed southeastwards to join the Bytham in what is now the NE part of the Fen Basin. For example, a confluence point near Wisbech (Figs 1 and 4(a)) would have been ~50 km downstream from Castle Bytham and ~60 km downstream of Rauceby. As was discussed above, the Bytham River had a steep longitudinal gradient for ~3 km between Castle Bytham and Witham-on-the-Hill, where it was ~27 m O.D. in the immediate pre-Anglian period. Assuming a uniform 0.28 m km^{-1} longitudinal gradient of the Bytham River across the Fen Basin (section 6.1), is height is predicted to have been ~14 m O.D. at Wisbech, making the fall of the Ancaster Trent below Rauceby ~49 m - 14 m or ~35 m and its mean longitudinal gradient ~35 m / ~60 km or ~0.58 m km^{-1}. More realistically, the gradient of the Bytham River would probably have decreased downstream of a confluence with so substantial a tributary. If the decrease was, say, from ~0.33 to ~0.23 m km^{-1}, then the height of the Bytham at Wisbech would have been ~11 m O.D., making the mean gradient of the Ancaster Trent ~38 m / ~60 km or ~0.63 m km^{-1}, in approximate agreement with the earlier calculation for the reach in the vicinity of the Ancaster Gap. On the basis of consideration of the pre-Anglian palaeogeography and the heights of fluvial deposits and associated downstream gradients of rivers, this seems the most plausible reconstruction of the pre-Anglian drainage in this area.

Others (e.g., Briant et al., 1999; Lee, 2009) have noted that the lithology of pre-Anglian fluvial gravels and nearshore fluviallyderived marine gravels at sites in North Norfolk (e.g., Sidestrand, Trimingham and West Runton) is distinct from that of the Bytham River gravels and have attributed this to emplacement by the aforementioned hypothetical river flowing eastward from the Ancaster Gap. These gravels contain significant quantities of Pennine lithologies, including Carboniferous chert, 'Millstone Grit' sandstone, and white or colourless quartzose clasts that are probably also derived from 'Millstone Grit', as well as *Rhaxella* chert from North Yorkshire. A mix of fluvial, glacial and even longshore drift inputs, potentially at different times, can thus have given rise to the composition of these gravels, in which the durable components might have been multiply recycled, there being no clear evidence from which to attribute them to a river flowing through the Ancaster Gap. If a purely fluvial origin is sought then, instead of the 'Ancaster River', we suggest that these lithologies might have been transported to North Norfolk by a pre-Anglian river system with affluents farther north in the Pennines and in North Yorkshire. The lower reaches of such a river might have been oriented southeastward, along the axis of the gentle syncline in the now offshore part of the East Midlands Shelf, across which the dip direction in the Chalk changes from gently NE in the Lincolnshire Wolds to more steeply SW along the SW flank of Sole Pit High, some 80 km offshore of the modern Lincolnshire coastline (cf. Van Hoorn, 1987; Japsen, 2000; Hillis et al., 2008). Low proportions of clasts of Spilsby Sandstone have also been reported in the shallow marine (but fluvially derived) Wroxham Crag of North Norfolk (e.g., Rose et al., 2001). With the above-envisaged fluvial geometry, these could not have been fluvially derived direct from the outcrop around Spilsby (cf. Rose et al., 2001; Lee, 2009); they might instead have been transported by the Bytham River before being worked into North Norfolk by marine processes, or might even have been derived from equivalent rocks exposed in the sea floor to the north of Norfolk (cf. Van Hoorn, 1987; Bradshaw et al., 1992; Hillis et al., 2008). Conversely, small proportions of Spilsby Sandstone have been reported in deposits of the Bytham River (e.g., Rose et al., 1999), suggesting that its type outcrop around Spilsby lay within the Bytham catchment, which we envisage was due to the Ancaster Trent being a left-bank tributary as depicted schematically in Fig. 4(a). The Lincolnshire Wolds thus formed the northeastern limit of the Bytham catchment in the early Middle Pleistocene.

The downstream variations in QC ratios in the Bytham River deposits also support the interpretation that the Ancaster Trent was a tributary of the Bytham. The Ingham River gravels in East Anglia have QC ratios that are intermediate between the low values (2–3) from the Wreake valley (palaeo-Derwent confluence to Castle Bytham; section 6.1) and the much higher values from the Ancaster Trent (e.g., 36 at Gelston; section 2.3), a pattern that implies mixing of inputs of different compositions. For example, Bridgland and Lewis (1991) reported QC=5 at Timworth (TL853692) and Lewis and Bridgland (1991) reported QC=6 at Ingham

(TL855715), for a sample of 2220 clasts of 11.2–16 mm size. Neglecting input from other Bytham tributaries, such as the Northamptonshire Brigstock River (e.g., Westaway, 2011a), these values (QC=2, 5 and 36) are consistent with mixing of Bytham and Ancaster Trent deposits in the ratio of ~10:1, indicating that the Bytham was by far the most important arm of the river flowing into East Anglia and that the Ancaster Trent was not its sole affluent (contra Gibbard et al., 2013).

The uplift modelling also allows inferences to be made regarding the landscape in the vicinity of the Ancaster Gap in the period immediately after the Anglian glaciation. Interpolating as before between the estimates of post-425 ka uplift of 24 m at Tattershall (Fig. 17) and 40 m at Nottingham (Fig. 15), taking account of the location of the Ancaster Gap roughly 30% of the way between the former locality and the latter, one obtains 29 m for the Ancaster Gap. Neglecting any effect of glacio-isostasy, the present-day col at ~45 m O.D. thus restores to ~16 m O.D., so if the present morphology of the Ancaster Gap developed as a result of incision by (sub)glacial meltwater during the Anglian (cf. Pocock, 1954; Berridge et al., 1999; A.S. Howard et al., 2009) it would have formed a notable topographic low, albeit somewhat above the contemporaneous reconstructed height of the Lincoln Gap, which we estimate was -8 m O.D. (Fig. 5(b)). However, we favour the alternative interpretation that the present morphology of the Ancaster Gap developed later, its final phase of evolution having resulted from downcutting by the extended River Slea following the MIS 8 Wragby glaciation (Fig. 4(c)).

7. Discussion; Lateral variations in uplift rates and crustal properties

As has been noted previously (e.g., Westaway et al., 2006b; Westaway, 2009b, 2010a, 2011a), the results of uplift modelling of the type presented in section 4 can be compared with independent information on crustal properties from geophysical observables. Along with the uplift modelling results, variations of three such observables across the study region are compiled in Table 2: Moho depth, from seismic reflection profiling; thickness of mafic underplating at the base of the crust, from gravity studies; surface heat flow, from geothermics. The uplift modelling indicates the thickness of the mobile lower-crustal layer and thus, given the other observables, the depth of the base of the brittle upper crust. The surface heat flow at each locality can be partitioned into contributions from basal heat flow and from radioactive heat production in the upper crust. Table 2 shows that, at each locality, the surface heat flow can be so partitioned as to cause a temperature distribution at depth that places the 350 °C geotherm (taken as the temperature of the base of the brittle layer; cf. Sibson, 1983) at the depth where the uplift modelling requires the base of the brittle layer to be. The only independent variables in this comparison are the rate of heat production by radioactive decay in the upper crust and the depth distribution of this radioactive heating. Subject to the assumption of a standard depth distribution, a rate of heat production can be determined at all localities, such that the uplift modelling is consistent with the independent geophysical observations. No observational data currently exist, however, from which the resulting estimates of radioactive heat production in the study area can be verified; nonetheless, this quantity could be independently measured in the future and would thus provide a test of the solutions presented here.

Notwithstanding local complexities, the surface heat flow shows a regional increase from south to north across the study region (e.g., Rollin, 1995; Jackson, 2004), from <50 mW m^{-2} in the interior of the London Platform of southeast England to the much higher values (up to 115 mW m^{-2}; Manning et al., 2007) characteristic of northern England, where the crust is pervasively intruded by highly radioactive Palaeozoic granite. The crustal thickness shows a corresponding northward decrease, from 34–35 km in the London Platform to a typical value of ~31 km beneath Yorkshire (Chadwick and Pharaoh, 1998). The study region is also crossed from southeast to northwest by a zone of mafic underplating, ~2–3 km thick, at the base of the crust. The distribution of this layer, revealed by gravity data (e.g., Jackson, 2004), has been mapped out by al-Kindi et al. (2003).

The lowest uplift rates in and near the study region are evident in the south, at sites within the London Platform: at Peterborough (Westaway, 2011a) and Leicester (Fig. 16). As noted above (section 4.3), the Leicester area forms the northern part of a high-stability 'core' of the London Platform, which extends southward to the Northampton–Milton Keynes area (cf. Westaway, 2011a). Analysis of the uplift modelling in this study confirms previous conclusions (Westaway et al., 2002; Westaway, 2009b, 2011a) that the relative stability of the London Platform is a consequence of constriction of the mobile lower-crustal layer as a result of the low (≤50 mW m^{-2}) surface heat-flow. The latter is inferred to be a consequence of relatively low radioactive heat production in the uppermost 10 km of the crust, which is estimated as only ~0.8 μW m^{-3} at Leicester, a low value for continental crust by global standards. Low radioactive heat production is also indicated by the analyses for Peterborough and Mildenhall in Table 2, and also persists eastward across northern East Anglia (Westaway, 2009b). Estimates of the temperature at the base of the mobile lower-crustal layer, at ~430–440 °C, and of the lithospherric thickness, at ~100 km (Table 2), are close to previous estimates for sites in the London Platform (e.g., Westaway, 2009b). The general increase in uplift rates northward from Leicester is primarily due to the regional-scale northward increase in surface heat flow. The analysis indicates that this effect is a consequence of an increase in the radioactive heat production in the upper crust from <1 μW m^{-3} to ~2 μW m^{-3}. The resulting subtle variations in Moho temperature and thickness of the mobile lower-crustal layer are comparable with those estimated for central–southern England by Westaway et al. (2006b) and much less than the overall variation relative to northern England, where radioactive heat production at up to ~5 μW m^{-3} contributes to Moho temperatures of >600 °C (Westaway, 2009c).

The variations in uplift histories in the central part of the study region, between Derby and adjoining modelling sites, correlate with variations in crustal proprties (Table 2), notably with the slight variations in thickness of the mafic underplating that are indicated by the gravity data. Likewise, the westward decrease in uplift rates evident across the westernmost part of the study region (cf. Bridgland et al., 2014) is primarily a consequence of a westward increase in thickness of this mafic layer. Similar trends are evident elsewhere in the British Isles: for example, in the south, mafic underplating seems to be absent beneath central–southern England, but thickens westward to ~6 km beneath Cornwall and ~10 km beneath southern Ireland; farther north, however, the underplating is ~4–5 km thick beneath the North Pennines and the Lake District, increasing westward to ~10 km beneath Northern Ireland (e.g., Barton, 1992; al-Kindi et al., 2003; Westaway, 2010a; Green et al., 2012). These trends thus culminate in the extreme Late Cenozoic landscape stability evident in Ireland, where the mafic underplating reaches thicknesses of ~10 km (Westaway, 2010a). This regional westward increase in mafic underplating is a consequence of the distribution of magmatic activity associated with the Early Cenozoic British Tertiary Igneous Province, as Al-Kindi et al. (2003) pointed out.

Al-Kindi et al. (2003) reported no measurable thickness of mafic underplating beneath the northeast part of the present study region around Tattershall. Thus, although the crust is thinner here (estimated as ~31.5 km thick: Table 2) than in any other locality analysed, the mobile layer extends to the Moho and is thus thick enough to sustain rates of vertical crustal motion that are faster than around Leicester, albeit slower than in other parts of the TVPP project area; the 24 m of post-Anglian uplift inferred from the modelling implies a time-averaged rate of ~0.06 mm a^{-1}. Although uplift rates decrease northward from Peterborough to Tattershall, they may increase north of the latter locality. For example, the estuarine and marine beach deposits at Kirmington (documented by Bridgland et al., 2014) indicate a contemporaneous sea-level 18–25 m above that at the present day. Previously reported as Hoxnian (Watts, 1959; cf. Thomas, 2001), this site could date from MIS 11, MIS 9 or even MIS 7e. Time-averaged uplift rates of ~0.04–0.06 mm a^{-1}, ~0.05–0.08 mm a^{-1}, and ~0.08–0.11 mm a^{-1}, respectively, would be indicated for each of these possible age interpretations, assuming (as before) that during each of these interglacials the eustatic sea level was the same as at present (cf. Siddall et al., 2003). At Bielsbeck (SE 861 378), ~70 km north of Lincoln and ~15 km north of the inner Humber Estuary), fossiliferous freshwater marl reaches ~12 m O.D. and is assigned to late MIS 7 (MIS 7a; ~190 ka) on the basis of its mammalian biostratigraphy (Schreve, 1999). This deposit is likely to represent a fragment of the pre-Devensian Yorkshire Ouse drainage system that has survived the MIS 2 glaciation; it was not overridden by ice but instead protrudes from near the edge of Lake Humber sediments, in the modern valley of the Foulness, a north-bank Humber tributary (Halkon, 1999). A post-MIS 7a uplift rate no greater than ~0.06 mm a^{-1} (~12 m / ~190 ka) is indicated here. At Sewerby (TA 199 686), on the East Yorkshire coast ~100 km north of Lincoln, a raised-beach gravel that reaches ~4 m O.D. is securely dated to the Ipswichian Interglacial (MIS 5e) from the occurrence in it of *Hippopotamus amphibious* and a luminescence date of 121±12 ka from overlying aeolian sand (Bateman and Catt, 1996). A local uplift rate of ~0.03 mm a^{-1} (~4 m / 125 ka) is indicated. Corresponding northward tapering in uplift rates is also apparent further inland. Thus, at Creswell Crags, ~35 km north of Nottingham, where a minor left-bank tributary of the Trent has incised a 30 m deep gorge through the Permian Magnesian Limestone, U-series dating (Rowe et al., 1989; Jacobi et al., 1998) of karstic levels indicates a rate of incision/uplift of 0.06 mm a^{-1}, less than the ~0.08–0.09 mm a^{-1} time-averaged post-Anglian uplift rates evident at Derby and Nottingham.

The part of East Yorkshire with relatively low uplift rates, represented by Bielsbeck and Sewerby, is bounded to the north by the west–east-trending Flamborough Head Fault Zone, which separates regions with significantly different structural development indicative of different crustal properties (e.g., Donato and Megson, 1990; Donato, 1993). North of this fault zone, the altitudes of the (? MIS 7) Speeton Shell Bed (e.g., Wilson, 1991) and the aforementioned MIS 7 Easington raised beach (e.g., Davies et al., 2009) are indicative of faster uplift (Westaway, 2009c). Like other examples already discussed, here, too, a crustal discontinuity that has had a strong effect on past geological evolution also influences the lateral variation in Quaternary landscape development.

Like Tattershall, the above-mentioned localities lie beyond (i.e., northeast of) the region where Al-Kindi et al. (2003) identified mafic underplating at the base of the cust; the main factor determining their low uplift rates is the relative thinness of the crust (typically ≤31 km; Chadwick and Pharaoh, 1998). Thus, although the mobile layer extends to the Moho beneath each of these localities, it is nonetheless constricted by the relatively shallow Moho depth. Uplift rates thus increase northward across Leicestershire and south Nottinghamshire, then decrease further north to lower values across north Nottinghamshire, north Lincolnshire and east Yorkshire, before faster uplift resumes in North Yorkshire and County Durham. Such complexity in patterns of vertical crustal motion was unforeseen prior to the TVPP; for instance, Westaway (2009c) envisaged a monotonic increase in uplift rates from low values in the London Platform, Lincolnshire and East Yorkshire to higher values in North Yorkshire and County Durham.

The zone of relative Quaternary landscape stability in East Yorkshire, revealed by Bielsbeck and neighbouring sites, corresponds with the Mesozoic ‘Market Weighton Axis’, marked by non-deposition in the Middle and Late Jurassic and Early Cretaceous (such that, uniquely for England, the Chalk directly overlies the Early Jurassic ‘Lower Lias’ or Scunthorpe Mudstone Formation). Thus, while localities both to the north and south were subsiding and acting as depocentres for marine sediments, this region of East Yorkshire remained virtually stable; the Jurassic and Early Cretaceous successions thin towards it from both north and south. This locality thus provides another example of a correlation between rates of Quaternary landscape development and evidence of crustal stability inherited from the ancient geological past (see also Westaway, 2011a).

The axis of most rapid uplift, recognized in the present study, extends northwest from the Nottingham area across the Peak District, where it is represented by the modelling localities of Lathkill Dale and Castleton (Fig. #01) documented by Bridgland et al. (2014). Late Middle Pleistocene and Late Pleistocene uplift rates in these localities are up to 0.15 mm a^{-1}; within the Peak District, this uplift is revealed primarily by the disposition of karstic levels. Bridgland et al. (2014) have thus inferred up to 160 m of uplift in this region since the Early Pleistocene, with an estimated maximum of >400 m since the Mid-Pliocene. Uplift at similar rates (~0.15–0.2 mm a^{-1}) is also evident in the Yorkshire Dales karst (i.e., in the Central Pennines) and persists farther north into the North Pennines and North Sea coastal region of County Durham (Westaway, 2009c, 2012; Waltham and Long, 2011; Waltham, 2012), where a corresponding rate is indicated by the height of the MIS 7 Easington raised beach (Davies et al., 2009). It can be presumed that similar uplift rates also pertain in the South Pennines, including the Dark Peak uplands of the northern Peak District, although the absence of any suitable evidence (this region has neither karst nor long-timescale fluvial sequences) means that this is not at present firmly established. The axis of maximum Quaternary uplift, running along the Pennines, is thus enhancing the overall anticlinal form of these uplands.

In summary, this analysis indicates that most of the variation across the study region in the vertical crustal motions and the thermal state of the crust arises as a result of lateral variations in radioactive heat production. The substantial variations in surface heat flow, from ~50 to ~65 mW m^{-2}, thus include a very minor variation, by ±2 mW m^{-2} about 40 mW m^{-2}, in the basal component (Table 2). The lithosphere thickness (assuming that the base of the lithosphere marks a uniform temperature of 1400 °C) is also near constant, at ~102±5 km. For comparison, the North Pennines are underlain by granite with radioactive heat production of ~5 μW m^{-3} (e.g., Richardson and Oxburgh, 1978) and a vertical extent of ~10 km (e.g., Bott, 1967; Kimbell et al., 2010), over which the surface heat flow is typically ~100 mW m^{-2} (cf. Downing and Gray, 1986; Rollin, 1995; Manning et al., 2007). The Moho depth is ~32 km, with the deepest ~5 km of crust composed of mafic underplating (e.g., Barton, 1992; Chadwick and Pharaoh, 1998). The same procedure as is used to calculate the parameters in

Table 2 would indicate that beneath the North Pennines the base of the brittle upper crust (at ~350 °C) is ~15 km deep, so the mobile lower crust is ~12 km thick (compared with ~7 km beneath the Trent catchment). Representative temperatures beneath the North Pennines of ~540 °C at the base of this mobile layer and ~630 °C at the Moho can thus be estimated, making a representative lithospheric thickness ~80 km. Given the much greater mobility of the lower crust beneath the North Pennines, expected from the higher temperatures at depth, Westaway's (2009c) estimate of up to ~900 m of post-Mid Pliocene uplift (in comparison with the upper limit of ~400 m in the Peak District, from Bridgland et al., 2014) is not unreasonable. It is thus evident that, like in other parts of Britain, most of the present relief in these regions has developed since the Mid-Pliocene. Finally, until recently, the literature on landscape evolution of Britain was dominated by the view that uplift has originated in the mantle, as a result of mantle-plume activity (see Green at al., 2012, and references cited therein). The present dataset indicates that, on the contrary, the process has arisen as a result of the effects of surface processes (driven by climate) on induced flow in the lower continental crust, as is evident in many other regions worldwide (e.g., Westaway et al., 2009).

8. Conclusions

Study, during the TVPP, of fluvial deposits within the greater Trent system, building upon previous knowledge, has indicated major changes in catchment geometry, each associated with glaciation. Prior to the Anglian glaciation, much of the modern Trent catchment was drained by the Bytham River, which flowed eastward from the Leicester area across the Fen Basin and East Anglia to the southern North Sea, as illustrated in Fig. 4(a). The lower reach of the modern Middle Trent, east and northeast of Nottingham, drained eastward across the Lincolnshire Limestone outcrop, upon which it emplaced the Rauceby Gravel. The river in question, termed the 'Ancaster Trent', appears to have been a left-bank tributary of the much larger Bytham, joining the latter river in the area now occupied by the Fen Basin (Fig. 4(a)).

Following the Anglian glaciation, a new drainage pattern arose, in which the newly formed Soar joined the Derwent north of Loughborough, the combined river flowing northward to Nottingham then northeastward along the primordial Trent Trench, ultimately reaching the newly formed Fen Basin and the North Sea (Fig. 4(b)). This river had a much steeper longitudinal gradient than the modern Trent (Fig. 5(b)), due to its smaller catchment, which excluded the modern Upper Trent; as a result its reconstructed valley floor passed below O.D. near Newark-on-Trent and, during the MIS 11 and MIS 9 interglacial marine highstands, a large estuary would have extended inland to this point, as illustrated in Fig. 20.

Much of the Trent system was again overridden by ice during the MIS 8 Wragby glaciation, following which the catchment attained its present dimensions and the river developed its present longitudinal gradient, although it continued to flow via the Lincoln Gap into the Fen Basin (Fig. 4(c)). Most of the extant Trent terrace staircase developed while this course was maintained. However, many of the earliest terrace deposits were emplaced during deglaciation, so their heights are glacially or glacio-isostatically perturbed. In the Middle Trent, the height of the Etwall terrace is representative of conditions in late MIS 8 after the glacio-isostatic effect had been removed from the system, that of the Egginton Common terrace records the valley floor during MIS 6, and that of the Beeston terrace during MIS 2, although in the latter case the uppermost Trent was again affected by glaciation, at the southeastern extremity of the Late Devensian Irish Sea ice lobe (Fig. 4(e)). The height of the youngest Pleistocene terrace, the Holme Pierrepont Terrace, relates to the modern Humber course rather than the former Lincoln Gap drainage, despite the fact that the river initially cut down from the Beeston level while still flowing through the Lincoln Gap.

The Etwall, Egginton Common and Beeston terraces of the Trent and their tributary counterparts, all associated with the former course through the Lincoln Gap (Figs. 4(c) to (e)), are, therefore, the only ones for which height differences are representative of uplift. Modelling of these height differences reveals significant lateral variations in uplift across the catchment, and also allows heights of older deposits to be restored to their times of emplacement, enabling the palaeo-drainage reconstructions for earlier times (Figs. 4(a),(b),and 5(b),(c)). The most dramatic lateral variation in uplift evident across the Trent catchment is the northward increase along the Soar valley between the Leicester and Nottingham areas, reflected in the downstream divergence of the Soar terraces (Fig. 12) and the associated estimates of 19 m and 40 m of post-Anglian uplift in these localities (Figs 15 and 16; Table 2). Leicester adjoins the northern limit of a region of relative crustal stability, which occupies much of the southern part of the English Midlands. Modelling of these and other uplift histories indicates subtle lateral variations in crustal properties across the study region, the relative stability around Leicester apparently reflecting low radioactive heat production in the upper continental crust (Table 2).

Thus, in addition to the provision of improved documentation and understanding of the Quaternary deposits pertaining to the history of the 'greater Trent' that has arisen from the TVPP (Bridgland et al., 2014), including placing the fluvial record within the context of repeated glaciation, the present study has related the heights of these various deposits to crustal properties, in connection with modelling of the uplift they record.

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Figure Captions

Figure 1. Location map showing the extent of the modern Trent and Witham catchments in relation to Devensian ice margins (Cheshire ice margin after Parker and Chambers, 1997, Clark et al., 2004a, and Worsley, 2005; margin of the North Sea ice lobe in Lincolnshire and Norfolk after Clark et al., 2004a, and Bateman et al., 2008). Note that these ice margins were diachronous; recent analyses (e.g., Bateman et al., 2011; Clark et al., 2012) indicate that the maximum advance of the Cheshire ice lobe occurred circa 25 ka whereas that of the North Sea ice lobe was within the span of time ~21–17 ka. The sites at which uplift histories have been modelled as part of this study are depicted, along with those previously modelled by Westaway (2009b) (Mildenhall) and by Westaway (2011a) (Peterborough). For the four sites in the Peak District (Castleton, Lathkill Dale, Matlock, and Wetton Mill), preliminary uplift modelling solutions have been presented by Bridgland et al. (2014) and definitive analyses will be published elsewhere. Inset shows location.

Figure 2. Long profile projection of the Trent terrace deposits, illustrating the relative positions of the localities discussed in the text and the stratigraphic relations between them. The projection is oriented west–east, with the western end of (a) at SK150450 (near Ashbourne, Derbyshire) and the eastern end of (b) at TF350450 (near Boston, Lincolnshire). Reaches along which the Trent or its tributaries do not flow eastward are thus foreshortened to some extent. The data depicted comprise a mixture of heights of terrace surfaces, measured from local 1:25,000 scale topographic map coverage, and thicknesses of sediment bodies, obtained from MAR publications, with distinctive symbols used for the different terrace deposits. Dashes indicate the present level of the Trent. Terrace surfaces are correlated using tie lines, the symbols and tie lines being less prominent where tentative. Abbreviations (listed here both for this and later Figures) denote individual fluvial terraces and/or fluvio-glacial deposits. Trent terraces are denoted thus: B, Bassingfield; Ba, Balderton; Be, Beeston; BT, 'Buried Terrace'; Eg, Egginton Common; EMl, Eagle Moor (lower facet); EMu, Eagle Moor (upper facet); Et, Etwall; HP, Holme Pierrepont; Ml, Martin (main or lower facet); Mu, Martin (upper facet); S, Scarle; Sa, Sandiacre; So, Southrey; TC, Tattershall Castle; and TT, Tattershall Thorpe. Derwent terraces are denoted thus: Al, Allenton; Am, Ambergate; Bo, Borrowash; LE, Little Eaton; Mt, Matlock; and Oc, Ockbrook. The Sandiacre / Eagle Moor (upper) / Martin (upper) terrace is labelled using different ornament to

signify that it is considered glacio-isostatically perturbed. C denotes the high-level Somersal Herbert–Chellaston fluvio-glacial deposits, showing in faint ornament a tentative tie line to indicate how it is envisaged that these deposits relate to those at East Leake and elsewhere.

Figure 3. Schematic west–east cross-section linking the reaches of the Lincoln Trent west and east of Lincoln, illustrating the proposed terrace correlation and the stratigraphic relations to between cold-climate fluvial sands and gravels, temperate-climate deposits, and glacigenic deposits. Modified after Fig. 2 of White et al. (2010).

Figure 4. Sequence of palaeogeographic maps illustrating changes to the geometry of the Trent catchment since ~500 ka. See text for discussion.

Figure 5. Predicted longitudinal profiles for the Middle and Lower Trent and the lower reaches of the Derwent at key times during the Pleistocene, with subsequent uplift restored. Profiles are labelled using the same ornament styles as in Fig. 11, with more prominent ornament used to indicate sites that contribute significant observational constraints. All profiles illustrate downstream distance measured along the river from zero at Willington; some sites therefore plot at different distances downstream of Willington at different times, as a result of the course changes depicted in Fig. 4. **(a)** The predicted river level at the time of emplacement of the uppermost part of the Beeston Sand and Gravel and its counterparts. **(b)** The predicted river level in the period immediately before the Wragby Glaciation. **(c)** The predicted river level in the immediately post-Anglian period. **(d)** Comparison of (a) to (c). Note that we consider it probable that the drainage at the level depicted in (c) only existed as a result of glacio-isostasy for a short time immediately after the Anglian ice sheet withdrew progressively from the region, including the Derwent Valley - Wilford Hill - Trent Trench area. Localities farther east and south were probably already deglaciated at this time; we thus expect that the glacio-isostatic effect died out downstream, so the long profile of the river at the time of emplacement of the Wilford Hill gravel converged downstream with that in (b).

Figure 6. Comparison of limiting OSL and radiocarbon dates for the Beeston Sand and Gravel and Holme Pierrepont Sand and Gravel. Radiocarbon dates are plotted as solid lines with their calibrated numerical ages in brackets; OSL dates are plotted as dashed lines with numerical ages without brackets. Error bars that lack end ticks denote composites of multiple dates, rather than individual dates. Calibration of these and other radiocarbon dates discussed in the main text (except those from Schreve et al., 2013) utilized version 1.5 of quickcal 2007, available online at www.calpal-online.de; the INTCAL09 calibration curve (Reimer et al., 2009) gave similar results. Dashed horizontal line indicates the possible demarcation between the terraces at 26 ka on the basis of the OSL dating evidence. For the Beeston Sand and Gravel and its tributary counterparts, 1 is the youngest radiocarbon date from Tattershall Castle, of 30800±360 radiocarbon years, from Straw (1979), which adjusts after calibration to 34963±416 yrs, whereas 2 and 3 are the youngest OSL dates from the Kirkby-on-Bain and Langford quarries, from the TVPP (Schwenninger et al., 2007a; Bridgland et al., 2014). For the Holme Pierrepont Sand and Gravel and its tributary counterparts, 4 is the radiocarbon date from Quorn, near Loughborough, in the Syston Sand and Gravel of the Soar, of 28875±205 radiocarbon years, from Brown et al. (1994), which adjusts after calibration to 33387±368 yrs; 5 is the radiocarbon date from Coleshill, in the Tame valley, of 32160+1780/-1450 radiocarbon years, from Coope and Sands (1966), which adjusts after calibration to 36991+2232/-1870 yrs; 6 and 7 represent the ranges of calibrated radiocarbon dates on mammal bone and OSL dates, from the Lower Sands and Upper Gravels at Whitemoor Haye at the Tame-Trent confluence, discussed by Schreve et al. (2013); each set of OSL dates is the weighted mean (plus or minus twice its standard error) calculated for a standard level of water saturation [A] and for a history of water saturation throughout [B]). Finally, 8, 9 and 10 are the oldest OSL dates from Hanson Quarry, Barton-under-Needwood, and Besthorpe and Girton quarries, from the TVPP (Schwenninger et al., 2007a; Bridgland et al., 2014). See text for discussion.

Figure 7. Long-profile projection across the boundary between the Upper and Middle Trent in the Rugeley–Burton-on-Trent–Willington area. Projection is oriented N70°E, with its origin (at 0 km distance) at SJ650050.

Height data are from Stevenson and Mitchell (1955), for the Upper Trent, and from BGS DigMap elsewhere. Note the evident miscorrelation on BGS DigMap between some outcrops of fluvial sand and gravel and named terraces; the correlations of the Egginton Common and Etwall terraces between the Burton-on-Trent and Barton-under-Needwood areas are tentative. Also shown are the ranges of height of the sediments exposed in the Whitemoor Haye (~SK179316), Barton-under-Needwood (Hanson Quarry; ~SK201163), Manor Park (~SK112168), and Rugeley (~SK051176) quarries. These quarries are situated within deposits that have been assigned to the Holme Pierrepont terrace, close to the level of the modern Upper Trent floodplain; the first two are discussed at length by Bridgland et al. (2014), the others by Stevenson and Mitchell (1955).

Figure 8. Map of central England, modified after Fig. 1 of White et al. (2010) and Fig. 2 of Westaway (2010b) to show the localities discussed that are indicative of the inferred MIS 8 Wragby glaciation, in relation to the Trent system and the documented glacial limit during the Devensian (the localities depicted as ice-covered having not necessarily been glaciated simultaneously). Ornament style [1] denotes the limits of the Wragby glaciation inferred by White et al. (2010); style [2] indicates the limits proposed in the present study, where they differ from [1]. These inferred glacial limits in the Birmingham and Peterborough areas may, likewise, not have been glaciated simultaneously; the ice-marginal outwash channel system envisaged by White et al. (2010) as extending southward from the Lincoln area to the vicinity of Peterborough may thus have developed after deglaciation of the region east of Birmingham and Derby. Thus the east-trending tunnel valleys in the area south of Derby may well date from the Wragby glaciation, not the Anglian as hitherto assumed (see also Bridgland et al., 2014). A different geometry of ice-maginal drainage may be envisaged during the Wragby glacial maximum, in the area east of Birmingham, possibly utilising low points in the Jurassic limestone such as the Coston and Ancaster gaps. See text for discussion.

Figure 9. Interpretation of the ice-margin geometry in East Anglia during the Wragby glaciation, proposed as Fig. 1(b) of Westaway (2010b) and consistent with Straw (1965, 1973). The ice margin is placed north of the Nar valley as sediments there, dated to the Hoxnian (e.g., Scourse et al., 1999) and discussed in the main text, show no evidence of subglacial deformation. The eastern extremity of this ice sheet in NW Norfolk is envisaged in the modern Glaven valley, from which outwash was directed southward into the modern Bure catchment, the associated fluvio-glacial deposits being envisaged as inset against older glacigenic sediments forming the Cromer Ridge. Glacigenic sediments attributed to this stage are overlain by deposits of the Morston raised beach (e.g., Gale et al., 1988; Gale and Hoare, 2008); the MIS 7 age recently proposed for these beach deposits (Hoare et al., 2009) makes MIS 8 the youngest feasible age for the underlying glacigenic sediments. However, other radically different interpretations for this area, including ages as young as MIS 6 (e.g., Hamblin et al., 2000, 2005) and as old as MIS 12 (Pawley et al., 2008) for the glacigenic sediments in the Glaven valley have also been proposed. Although this ice sheet is inferred to have abutted relatively high ground in western Norfolk, preventing the development of an outwash channel along its southern margin (in contrast with the configuration in MIS 2, illustrated in Fig. 4(f), where such a channel was present). Outwash from the Tottenhill area was thus directed southward, in accordance with the available evidence (e.g., Gibbard et al., 1992, 2008), and so could only reach the North Sea via the Little Ouse / Waveney overflow channel. Conversely, the extent of development of fluvial terraces in the valleys of the Waveney (Coxon, 1993; Westaway, 2009b) and Little Ouse (Gibbard et al., 2008) would seem to preclude any age for the Wragby glaciation of this area that is later than MIS 8. Straw (2011) has advocated a less conservative ice margin than is depicted here; his interpretation crosses the North Sea coast near Weybourne (i.e., in the same place as is drawn here), then turns southeastward, reaching within a few kilometres of Norwich, before continuing southwestward in the vicinity of Thetford and Cambridge. The maximum extent of any Wragby glaciation ice sheet can, indeed, not extend any significant distance east of Weybourne, as the glacial succession in localities farther east, including the glacigenic gravels capping the Cromer Ridge, is securely dated and is unequivocally Anglian (e.g., Preece et al., 2009; Westaway, 2010c). It is noted in passing that Murton and Murton (2012) have suggested that the Little Ouse / Waveney overflow channel system was active during the Devensian. This is another old idea (see citation of references by Westaway, 2010b), which is nonetheless contradicted by an abundance of evidence, such as the presence of the aforementioned terrace staircases in the Little Ouse and Waveney valleys. Furthermore, it is now apparent that during the maximum Devensian ice advance in Lincolnshire and NW Norfolk, the outlet from the Fenland

proglacial lake lay along the southern margin of the ice sheet, in the immediate vicinity of the Norfolk coast (Moorlock et al., 2002, 2008), as schematically illustrated in Fig. 4(f).

Figure 10. Map of part of the SW North Sea showing the Inner Silver Pit, the offshore (submerged) course of the former 'Wash River', along with the 'Botney Cut channel' that appears to mark the submerged course of the latest Pleistocene River Humber. The Sand Hole, also shown, has been suggested (Tappin et al., 2011) as marking a later course of the Humber (? Early Holocene), formed when the sea surface approached its present level. Also shown are the locations of borehole 81/52A (Kristensen et al., 1998; Scourse et al., 1999) and the nearby sites of Fisher et al. (1969); each of these localities have yielded interglacial marine sediments of probable Hoxnian (MIS 11) age from depths of ~50 m below the sea surface, suggesting that the Inner Silver Pit (before its form was modified by later subglacial processes) probably formed an embayment into which the estuary depicted in Fig. #20 debouched. Modified from part of Fig. 4.2.17 of Tappin et al. (2011).

Figure 11. Long profile projection of the Trent terrace deposits between Burton-on-Trent and the Trent Trench, in relation to those of the Derwent downstream of Belper, illustrating the stratigraphic relations between localities discussed in the text. The projection is oriented N80°W–S80°E, with its origin (downstream distance 0 km) at SK 200 300 (~10 km ESE of Uttoxeter and ~10 km NNW of Burton-on-Trent). Reaches where rivers do not flow parallel to this section line are thus foreshortened to some extent. Place names are provided for location purposes, those in the Derwent being appropriately annotated. The Derwent and its terraces are depicted schematically, modified after Waters and Johnson (1958) and other sources discussed in the text (sections 3.1 to 3.5). Terraces of the modern river system (which first developed during Wragby deglaciation) are labelled using the abbreviations set out in the caption to Fig. 2. Terrace surfaces are indicated using distinctive symbols with tie lines, the symbols being fainter and the tie lines dashed where tentative. Terraces of the River Soar (Fig. 12), projected to the modern Soar-Trent confluence, are indicated thus: UB, Upper Birstall; MB, Middle Birstall; LB, Lower Birstall; W, Wanlip; and Sy, Syston (see Fig. 12 for more details). As illustrated in Fig. 12, it is envisaged that the highest, Knighton, terrace of the Soar in the Leicester area projects downstream into the East Leake gravel and thence into a downstream projection of the Chellaston fluvio-glacial deposits in the Middle Trent valley. Also depicted are schematic representations of the present-day level of the courses of the Derwent–Trent system in the immediately post-Anglian and pre-Wragby-Glaciation periods, and of the Bytham River and its Derwent tributary; these are constructed as explained in the text. Numbered key localities are identified thus: 1, Huncote (SP513978); 2, Wanlip Hill (SK590110); 3, Thurmaston (SK615101); 4, Wilford Hill (SK582352); 5, the highest point on the interfluve flanking the Trent Trench around Kneeton (including Old Hill, SK702452, 77 m O.D., and Toot Hill, SK704458, 76 m O.D.); 6, Hathern (SK503214); 7, East Leake (SK558248); 8, Chellaston (SK387296); 9, Somersal Herbert (SK145355, or ~6 km beyond the western end of the profile, ~130 m O.D.; evidently fluvio-glacial outwash routed into the Middle Trent by the Dove around the time when the Chellaston deposits were emplaced); 10, Sandiacre (SK463345); 11, Derby (SK352353); and 12, the lower and upper gravels (SK477352, ~42 m O.D., and SK478357, ~49 m O.D.) at Springfield Park, Sandiacre, near the Trent–Derwent confluence. The former is the only outcrop of the Egginton Common Sand and Gravel currently recognized in BGS mapping (BGS DigMap) downstream of the Derwent confluence; the latter, hitherto mapped by BGS (BGS DigMap) as 'Terrace 2', is at roughly the expected height of the Etwall Sand and Gravel. Note the much steeper longitudinal gradient determined for the post-Anglian, pre-Wragby-Glaciation Trent compared with the modern river (see section 5.6), and the gentle back-tilting (recognized by Rice, 1991) evident for the Bytham River deposits in the area northeast of Leicester, due to the downstream increase in post-Anglian uplift (see section 6.1).

Figure 12. Long profiles of the Soar terraces, also showing the Hathern Gravel and the Thurmaston Sand and Gravel. Modified from part of Fig. 13 of Rice (1968), with additional information from Lewis (1989a), Rice (1991) and Maddy (1999), also incorporating the 75 m O.D. height of the East Leake gravel from Bridgland et al. (2014). The Birstall Terrace of Rice (1968) is here differentiated into Lower, Middle and Upper facets; these might alternatively be named, respectively, the Loughborough, Sutton Bonington, and Barrow-upon-Soar terraces, after localities where each facet is well-developed. Note the downstream divergence of the Soar terraces, attributed to a northward increase in uplift (see section 4), the low disposition of the Hathern Gravel,

and the fact that the Thurmaston Sand and Gravel is much lower in the landscape than the oldest, Knighton, terrace deposit of the modern River Soar, even though the former body of sediment is much older than the latter. The very low Quorndon Terrace of the Soar (Rice, 1968) appears to correlate with the upper surface of the Hemington Sand and Gravel of the Middle Trent.

Figure 13. Long-profile projection along the Upper Trent, between its headwaters and the Rugeley area. Projection is oriented SE, with its origin (0 km) at SJ750450 (~15 km west of Stoke-on-Trent). Heights are from BGS DigMap, supplemented by data from Stevenson and Mitchell (1955) in the Rugeley area, where (for a few kilometres) this profile overlaps with that in Fig. 7, and by data from Whitehead et al. (1927) and Morgan (1973) in the area around and upstream of Stafford. Note that outwash deposits emplaced when the Irish Sea ice lobe approximated its maximum extent are confluent with the Second Terrace (i.e., Beeston Terrace) of the Upper Trent, indicating equivalence in age (as recognized by Stevenson and Mitchell, 1955), as indicated for example by the disposition of the outwash from the Irish Sea ice lobe at its maximum extent in the Cannock Chase area, which entered the Trent valley from the south at Rugeley (circa SK042183). However, as discussed in the main text, deposits at the height of the Beeston Terrace also persist farther upstream, indicating that their emplacement continued during deglaciation, being confluent (for example) with outwash deposits leading southward from the Swynnerton area (circa SJ860346) into the Sow valley near Norton Bridge. However, it is inferred that the emplacement of the lower-level deposits (at the height of the First Terrace (i.e., Holme Pierrepont Terrace) of the uppermost Trent persisted while the margin of the Irish Sea ice lobe lay along the Bar Hill-Whitchurch-Wrexham moraine with the Trent catchment fed by meltwater through cols such as those at Bathpool (SJ838525), leading into Fowlea Brook and thence into the Trent, and at Whitmore Gap (SJ811415) at the headwaters of Meece Brook, leading into the River Sow.

Figure 14. Model predictions for the uplift history at Willington. The present local level of the Trent is 40 m O.D.; its terraces have been assigned heights (O.D.) of 43 m (Holme Pierrepont), 50 m (Beeston), 56 m (Egginton Common), 63 m (Etwall), 80 m (Sandiacre), and 92 m (Chellaston fluvio-glacial deposits), as in Fig. 2. Uplift has been measured relative to a reference level of 47 m O.D.; the height of the Beeston terrace has been corrected downward by 2 m to account for the longer downstream course at this time (Fig. 4(e)), compared with the Egginton Common and Etwall terraces. This prediction uses W_i=6 km, u=16 °C km^{-1}, κ=1.2 mm s^{-1}, t_o=0.87 Ma, and ΔT_e=-3.4 °C. Solid symbols indicate preferred age assignments for terrace deposits; open symbols indicate possible alternatives. Fainter ornament is used to denote observational evidence associated with downstream channel-lengthening effects and/or glacio-isostasy during the deglaciation phase following the Wragby glaciation (cf. section 2.3; in this case, for the Chellaston deposits and Sandiacre terrace), and thus unrepresentative of regional uplift. Large numbers indicate MIS stages when the youngest deposits forming a given terrace are inferred to have been emplaced, 8(G) denoting terraces at heights that have been glacially (or glacio-isostatically) perturbed. (a) observed and predicted uplift history. (b) predicted history of uplift rate variation.

Figure 15. Model predictions for the uplift history in the Wilford Hill area of Nottingham. The present local level of the Trent is 22 m O.D.; its terraces have been assigned heights (O.D.) of 25 m (Holme Pierrepont), 30 m (Beeston), 37 m (Egginton Common), 46 m (Etwall), 55 m (Sandiacre), and 67 m (Chellaston fluvio-glacial deposits), with the top of the East Leake Gravel at 75 m O.D. and the top of the Wilford Hill Gravel at 91 m O.D. Furthermore, the top of the Hathern Gravel has been projected downstream from its type locality to a height of 47 m O.D., as discussed in section 5.2. Uplift has been measured relative to a reference level of 26 m O.D.; the height of the Beeston Terrace has been corrected downward by 2 m to account for the longer downstream course at this time (Fig. 4(e)), compared with the Egginton Common and Etwall terraces. The prediction uses the same model parameter values and display format as Fig. 14 except ΔT_e=-4.0 °C. Furthermore, the Hathern and Wilford Hill gravels have been labelled using different ornament to signify that they were emplaced by the river system (Fig. 4(b)) that developed following the Anglian glaciation and was disrupted by the Wragby glaciation (see sections 5.2 and 5.3); the preferred age of the Hathern Gravel is thus labelled as 8+ to signify that it was emplaced during MIS 8 but before the maximum extent of the Wragby glaciation. Mismatches between the model prediction and the heights of the Sandiacre terrace and Chellaston

deposits are attributed, respectively, to glacio-isostasy and downstream channel-lengthening effects during the Wragby glaciation (see section 2.3). The agreement between the model prediction and the height of the Hathern Gravel is fortuitous, given the catchment and downstream gradient changes that occurred following the Wragby glaciation (see section 5.6). The ~24 m mismatch between the model prediction and the height of the Wilford Hill Gravel is explained in the main text (section 5.7). (a) observed and predicted uplift history. (b) predicted history of uplift rate variation.

Figure 16. Model predictions for the uplift history at Leicester. The present local level of the Soar is 54 m O.D.; its terraces have been assigned heights (O.D.) of 55 m (Syston), 59 m (Wanlip), 62 m (Lower Birstall / Loughborough), 65 m (Middle Birstall / Sutton Bonington), 70 m (Upper Birstall / Barrow-upon-Soar), and 85-91 m (Knighton). The top of the Thurmaston Sand and Gravel of the Bytham River is assigned a height of 74 m O.D., all these data being consistent with Fig. 12. Uplift has been measured relative to a reference level of 56 m O.D.; the height of the Wanlip terrace has been corrected downward by 2 m to account for the longer downstream course at this time (Fig. 4(e)), compared with the Lower Birstall / Loughborough and Middle Birstall / Sutton Bonington terraces. The prediction uses the same model parameter values and display format as Fig. 14 except ΔT_e=-1.7 °C and u=14 °C km^{-1}. Mismatches between the model prediction and the heights of the Upper Birstall / Barrow-upon-Soar and Knighton terraces are attributed to glacio-isostasy and downstream channel-lengthening effects associated with the Wragby glaciation (cf. section 2.3), these being the Soar counterparts of the Sandiacre terrace and Chellaston deposits of the Middle Trent. The ~3 m mismatch between the model prediction and the height of the Thurmaston Sand and Gravel is a consequence of this sediment having been deposited by a different river system (the Bytham), not the modern Soar. Even so, the height of this deposit (fortuitously) provides a rough indication of the uplift in this area since ~450 ka (see section 6.1). Like in Fig. 15, this deposit is labelled using distinct ornament to signify that it was not emplaced by the modern river system; the number 12+ signifies that it was emplaced before the Anglian glaciation. (a) observed and predicted uplift history. (b) predicted history of uplift rate variation.

Figure 17. Model prediction for the uplift history at Tattershall. The present local level of the Witham is close to O.D.; the terraces of the Lincoln Trent have been assigned heights (O.D.) of -1 m ('Buried Terrace' / Holme Pierrepont), 5 m (Tattershall Castle), 10 m (Southrey / Tattershall Thorpe), 15 m (Martin main facet at Tattershall Airfield; Power and Wild, 1982; BGS, 1995), and 20 m (Martin upper facet at Welsyke Wood MAR borehole (TF22776431; Power and Wild, 1982). Higher-level flats formed in outwash gravel from the Wragby glaciation are assigned the following heights O.D., all from Power and Wild (1982): 27 m (Tower Farm MAR borehole, at TF21476433); 36 m (Thornton MAR borehole, at TF23826681); 43 m (Scrivelsby, ~TF265659); and 53 m (Langton Hill, at TF247690) (see Fig. 2 and section 3.5). Uplift has been measured relative to a reference level of 4 m O.D.; the height of the Tattershall Castle terrace has not been corrected downward by 2 m (unlike for its counterparts elsewhere) because the longer downstream course at this time (Fig. 4(e)), compared with the Southrey and Martin terraces, developed in a reach located farther upstream. The prediction uses the same model parameter values and display format as Fig. 14 except ΔT_e=-2.4 °C. Mismatches between the model prediction and the heights of the upper facet of the Martin Terrace and the higher outwash flats are attributed to glacio-isostasy and/or downstream channel-lengthening effects associated with the Wragby glaciation, as discussed in section 2.3. (a) observed and predicted uplift histories. (b) predicted histories of uplift rate variation.

Figure 18. Model prediction for the uplift history at Derby. The present local level of the Derwent is 46 m O.D.; its terraces have been assigned heights (O.D.) of 54 m (Allenton / Ambergate), 60 m (Borrowash), 66 m (Ockbrook), 72 m (Little Eaton), 91 m (Matlock), and 119 m (High Tor), the latter height being based on a downstream projection from the type locality as discussed in section 5.4 and the height of the Little Eaton terrace being based on an upstream projection of the 'Derby gravel' (section 2.2; Fig. 11) by 1 km. Furthermore, the top of the Hathern Gravel (emplaced by the extended River Derwent that existed prior to the Wragby glaciation) has been projected upstream from its type locality to a height of 79 m O.D., as discussed in section 5.2. Uplift has been measured relative to a reference level of 51 m O.D.; the height of the Allenton–Ambergate terrace has been corrected downward by 2 m to account for the longer downstream course at this

time (Fig. 4(e)), compared with the Borrowash and Ockbrook terraces. The prediction uses the same model parameter values and display format as Fig. 14 except ΔT_e=-3.2 °C. Mismatches between the model prediction and the heights of the Little Eaton and Matlock terraces are attributed to glacio-isostasy and/or downstream channel-lengthening effects associated with the Wragby glaciation, these being the Derwent counterparts of the Sandiacre terrace and Chellaston deposits; cf. section 2.3. The ~12 m mismatch between the model prediction and the height of the Hathern Gravel is explained as a consequence of the rejuvenation of the river system due to the catchment and gradient changes that followed the Wragby glaciation. The ~35 m mismatch between the model prediction and the height of the High Tor terrace is explained in the main text (section 6.1). (a) observed and predicted uplift history. (b) predicted history of uplift rate variation.

Figure 19. Map of Lincolnshire showing the present-day altitude of the land-surface on which glacigenic sediments were deposited during the Wragby glaciation, modified with some simplification from Fig. 3 of Straw (1958). Heights at altitudes of ≤25 ft or ~8 m (1 ft = 0.3048 m) are emphasized, and newer data for the Rampton, Potterhanworth, Southrey, Tattershall and West Fen areas, from MAR publications (Price, 1975; Crofts, 1982; Jackson, 1982; and Jackson and Issaias, 1982), have been added.

Figure 20. Map of the study region showing inferred palaeogeography during the MIS 11 and MIS 9 interglacials. The extent of the very large Trent estuary, extending inland from the Wash to the vicinity of Newark, is based on the reconstructed pre-Saalian downstream gradient of the river in Fig. 5(b), making the assumption (as a generalization) that sea level during these interglacials was the same as at the present day. Sea-level variations of a few metres between interglacials, as might be expected (e.g., Siddall et al., 2003), would affect the predicted positions of the head of each estuary (by distances that can be readily calculated given the specified river gradients) but would not affect the overall conclusions significantly. Labelling is the same as for Fig. 4, except for the ornament to indicate the extent of the pre-Wragby-glaciation interglacial palaeo-estuary. Dimensions of the contemporaneous estuary of the River Nene, in relation to the locations of the Hoxnian Woodston Beds and Nar Valley clays, is from Fig. 6 of Langford and Briant (2004). Note, however, that calculations by Westaway (2011a) indicate that the top of the extant Woodston Beds lay some 10 m below the contemporaneous interglacial sea-level, indicating that these deposits were formerly more extensive and their original upper part has been lost to erosion. Given the ~0.5 m km^{-1} downstream gradient of the River Nene, estuarine conditions can be inferred to have persisted along the Nene valley for ~20 km (~10 m / ~0.5 m km^{-1}) upstream of the limit depicted, making the pre-Wragby-glaciation Nene palaeo-estuary of comparable dimensions to that of the Trent.

Table 1

Summary of theTrent terrace age model and supporting sedimentary and geomorphological evidence used in our drainage reconstructions.

Age and environment	Upper Trent	Middle Trent and tributaries			Lower Trent via Lincoln	
		Derwent	Soar	MiddleTrent	Newark–Lincoln	Lincoln–Boston
'Lincoln Trent'catchment						
Since late MIS2 (≤25 ka)	FirstTerrace	Chad. Sidings	Syston	Holme Pierrept	Holme Pierrept	'Buried terrace'
Latest MIS6 to MIS2	SecondTerrace	Allenton[#,a,b]	Wanlip	Beeston[c]	Scarle[#,d]	Tattershall Castle#
Latest MIS8toMIS6	NP	Borrowash[b]	L. Birstall[e]	Egginton Com.	Balderton*	Southrey*
Late MIS 8	NP	Ockbrook[b]	M. Birstall[e]	Etwall	Eagle M. (l.facet)	Martin(main facet)
Late MIS 8 (glacio-isos.)	NP	LittleEaton[b,f]	U. Birstall[e]	Sandiacre	Eagle M. (u.facet)	Martin(upper facet)
Late MIS8 (lateWrag.glac)	NP	Matlock[b,f]	Knighton	Chellaston	Skel.Clay[g]	Outwash gravels[h]
Wragby Glaciation - fluvial system disrupted						
Mid MIS 8 (Glac.max.)	Glaciated	Glaciated	Glaciated[i]	Glaciated	Glaciatedi	Glaciated; WragbyTill
'Soar-Trent' catchment, draining via Lincoln						
Early MIS 8 (pre-Wrag.Glac.)	NE	HathernGravel[b,j]	Hath. Gr.[j]	Trent Trench[k]	NP	Bedrock low[l]
Latest MIS12 to MIS 9	NE	Karst[b]	NP	TrentTrench[k]	NP	Bedrock low[l]
Late MIS12(glacio-isostasy)	NE	HighTor[b,f]	NP	Wilford Hill[k]	NP	NP
Anglian Glaciation – fluvial system disrupted						
Mid MIS12(glac.lmax.)	Glaciated	Glaciated	Glaciated	Glaciated	Glaciated	Glaciated
Pre-Anglian Bytham River catchment						
Until early MIS12	HinckleyRiver[l,m]	Derby River[b,m]	ThurmastonGr.[m]	NR	RaucebyGravel[n]	Rauceby Gravel[n]

Complications accompanying or following the latest Pleistocene diversion of the Lincoln Trent to the Humber are omitted from this Table (seethemain text). Some of the stated ages are supported by dating evidence (forexample: # and * denote evidence of the Ipswichian and MIS 7 interglacials), others are estimated; see the main text for details. NP denotes 'No Preservation', meaning that we infer that the locality was part of the Trent catchment at the time, but no sedimentary or geomorphological evidence has been preserved. NE denotes 'No Evidence', meaning that we infer from the lack of evidence that the locality was not part of the Trent catchment at the time .NR denotes 'No River', meaning that our drainage reconstructions suggest that no river bearing any resemblance to its modern counterpart in the locality existed at the time.

a This name applies to the Lower Derwent; it is considered here to include the Belper Gravel and to be equivalent to the Ambergate Terrace of the Middle Derwent.

b Karstic levels in the Peak District demonstrate that the RiverDerwent existed during this span ot ime.

c Equivalent to the Bassingfield Terrace in the reach of the Middle Trent between Nottingham and the upstream end of the Trent Trench at Radcliffe-on-Trent.

d Equivalent to the Fulbeck Sand and Gravel of the River Witham, upstream of I ts former Trent confluence, within which the evidence of the Ipswichian interglacial has been documented.

e The Lower, Middle and Upper Birstall terraces of the Soar might also be named as the Loughborough, Sutton Bonington, and Barrow-upon-Soar terraces.

f Terrace defined on the basis of bedrock flats rather than sedimentary evidence.

g Deposits emplaced in a proglacial lake upstream of Lincoln (probably after the maximum extent of the Wragby Glaciation), while the reach downstream of Lincoln remained glaciated.

h Fluvioglacial/glaciofluvial gravels flanking the modern Witham valley at sites such as Tower Farm, Thornton, Scrivelsby, and Langton Hill.

i Locality outside the most conservative estimates of the extent of the Wragby Glaciation but within limits considered plausible in this study.

j The Hathern gravel demonstrates the existence of the River Derwent and indirectly demonstrates the existence of the River Soar, into which the Derwent must have flowed.

k The Trent Trench must already have existed at this time and so contributes to constraining the drainage geometry.

l Bedrock lows recognized beneath the WragbyTill downstream of Lincoln and along the former 'Hinckley River' contribute to constraining the drainage geometry.

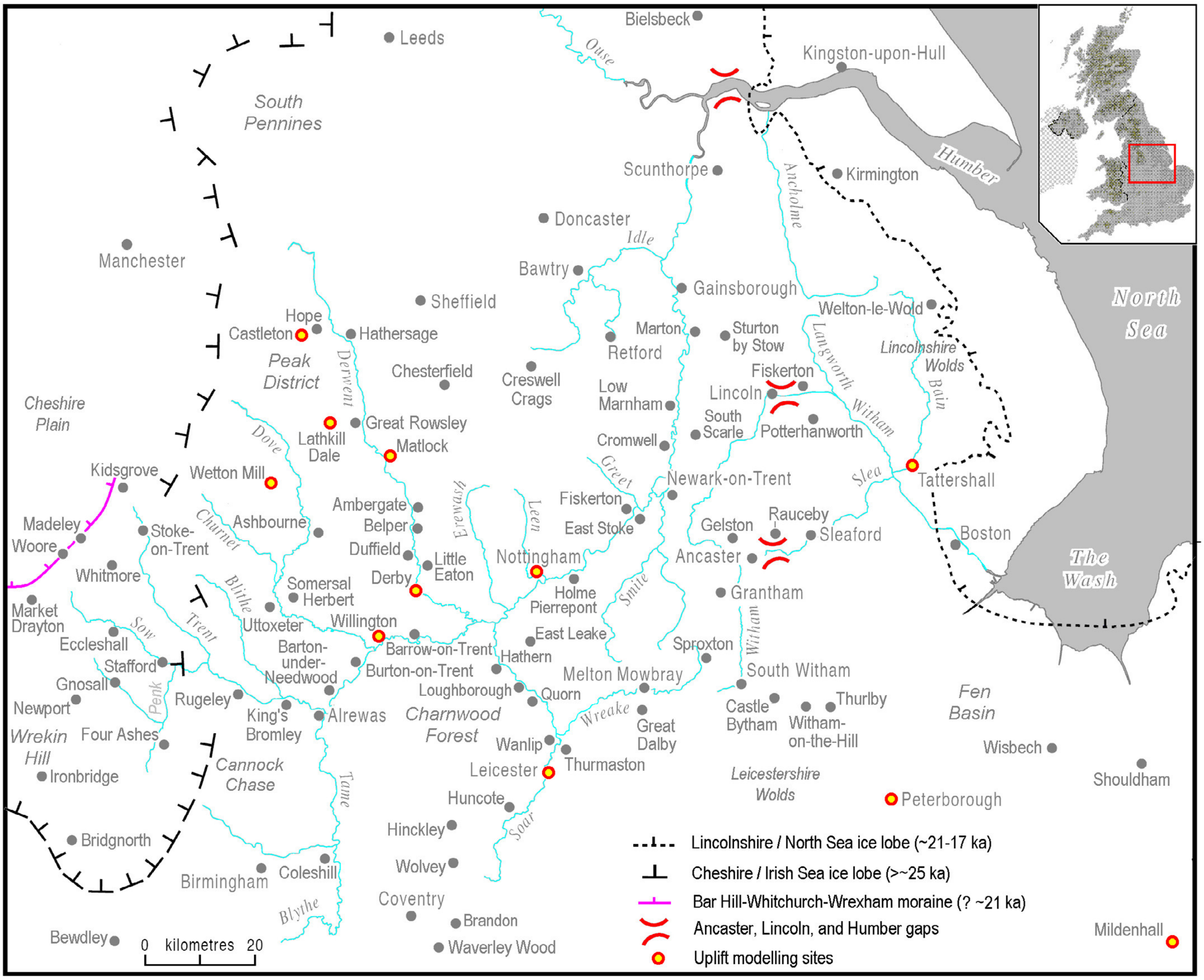
m Variations in ratios of Carboniferous chert to quartzose clasts demonstrate sediment transport into the Bytham River via the 'Hinckley River' and 'Derby River'.

n The Rauceby Gravel crops out west and east of the Lincolnshire Limestone escarpment, thus demonstrating throughgoing drainage, interpreted here as a tributary of the Bytham River.

Table 2
Summary of uplift modelling results.

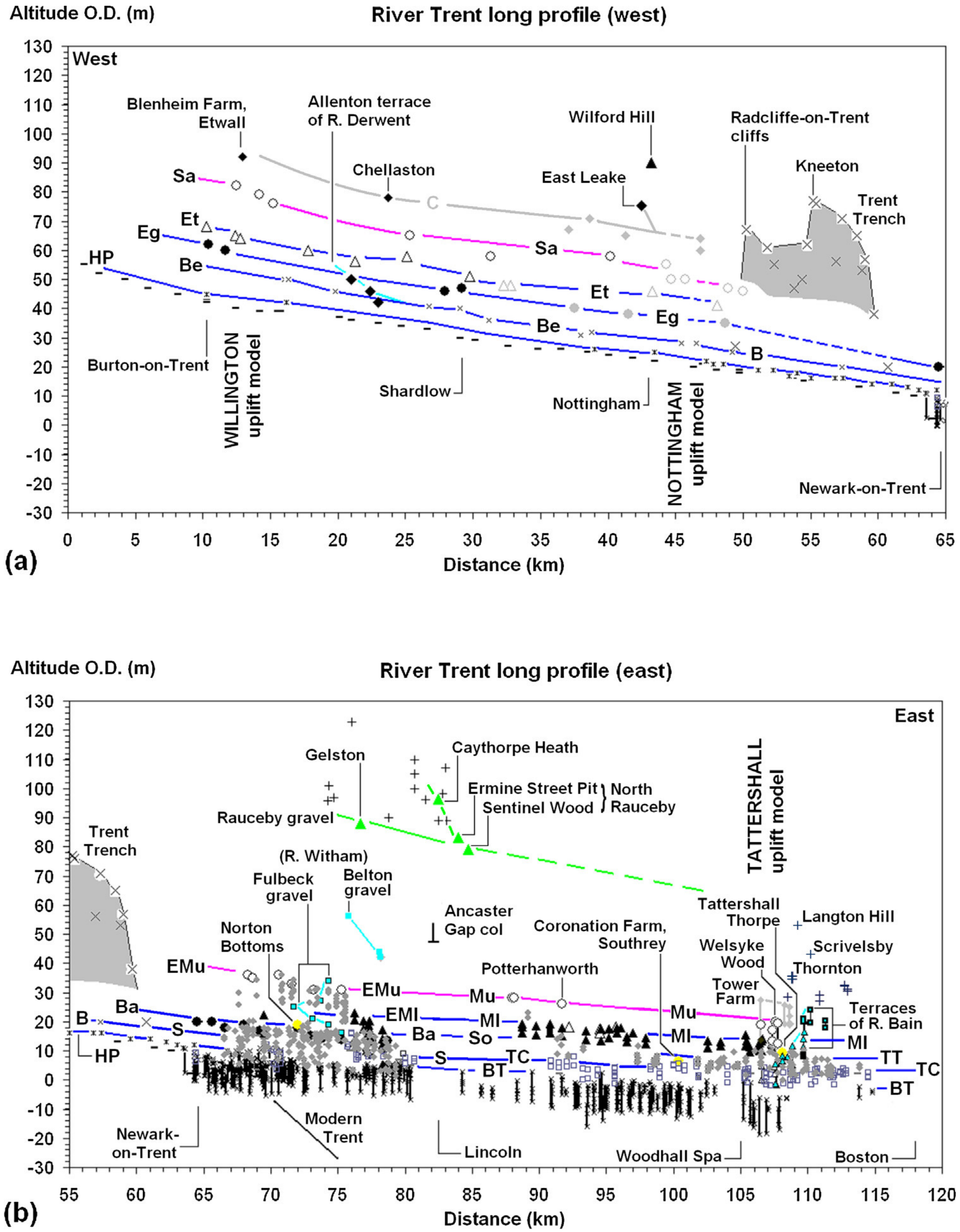
Site	W_i (km)	H_u (km)	$v_{u\,max}$ (mm a^{-1})	$t(v_{u\,max})$ (ka)	$U(250)$ (m)	$U(425)$ (m)	$U(450)$ (m)	$U(875)$ (m)	z_m (km)	z_b (km)	Q_s (mW m^{-2})	Y (μW m^{-3})	Q_o (mW m^{-2})	T_u (°C)	T_m (°C)	z_a (km)
Derby	6.0	2.5	0.196	700	16	32	35	93	32.0	22.8	62	2.22	39.8	439	472	101.9
Willington	6.0	2.0	0.208	700	17	34	37	99	32.5	23.8	62	2.43	37.7	434	459	107.4
Nottingham	6.0	2.0	0.246	700	20	40	44	117	32.5	23.8	60	2.18	38.2	435	461	106.3
Tattershall	6.0	0.0	0.147	700	12	24	26	70	31.5	24.8	55	1.74	37.6	434	434	108.6
Leicester	6.0	2.0	0.119	700	10	19	21	56	32.0	23.3	50	0.80	42.0	443	471	98.3
Peterborough	6.0	2.0	0.196	700	16	32	35	93	32.0	23.3	50	0.80	42.0	443	471	98.3
Mildenhall	6.0	2.0	0.181	700	14	29	32	85	34.0	25.3	42	0.21	39.9	439	465	104.2

Sites are listed in geographical order, from northwest to southeast; the solutions are illustrated, thus: Derby, Fig. 18; Willington, Fig. 14; Nottingham, Fig. 15; Tattershall, Fig. 17; and Leicester, Fig. 16; those for Mildenhall and Peterborough were originally documented by Westaway (2009b, 2011a) but are included here as they lie on the southeastern periphery of the TVPP project area and are used to estimate uplift histories at other sites within this area. The Mildenhall solution is based on solution 1 of Westaway (2009b), recalculated with $k = 3\,W\,m^{-1}\,°C^{-1}$ as the assumed thermal conductivity of the lower crust. W_i is the effective thickness of the lower-crustal layer, defined as $z_i - z_b$ where z_b is the depth of the base of the brittle upper crust and z_i is the depth at which lower-crustal flow is concentrated. W_i is thus roughly equivalent to ~90% of the overall thickness of the mobile lower-crustal layer, between the base of the brittle upper crust and the base of the mobile layer (cf. Westaway, 1998, 2002b). H_u is the estimated thickness of mafic underplating at the base of the crust, from Al-Kindi et al. (2003). $v_{u\,max}$ is the predicted maximum uplift rate, $t(v_{u\,max})$ being its timing. $U(t)$ shows the predicted uplift since time t (in ka) for four significant time scales. z_m is the local crustal thickness (i.e., Moho depth), from Chadwick and Pharaoh (1998). z_b is the associated estimate of the depth of the base of the brittle layer, calculated as $z_m - H_u - W_i/0.9$. Q_s is the local surface heat flow, estimated from Jackson (2004). Q_s is assumed to consist of a basal component Q_o and a component Q_R produced by radioactive heating in the upper crust. The latter component is estimated as $Y \times D$ (so $Q_o = Q_s - Y \times D$) where Y is the heat production rate and D is the thickness of the radiogenic layer. D is taken as 10 km, the top of this layer (at a temperature of 10 °C) being at the Earth's surface. T_u is the temperature at the base of the mobile layer (i.e., at depth $z_u = z_m - H_u$), estimated as $T_b + (Q_o/k) \times (z_u - z_b)$; T_m, the Moho temperature, is estimated as $T_b + (Q_o/k) \times (z_m - z_b)$. In this calculation, $T_b = 350\,°C$ (cf. Sibson, 1983) is the assumed temperature at the base of the brittle upper crust (at depth z_b). The depth of the base of the lithosphere, z_a, is also crudely estimated assuming a uniform geothermal gradient, as $z_b + (T_a - T_b) \times (k/Q_o)$, where T_a, the estimated asthenosphere temperature, is 1400 °C.



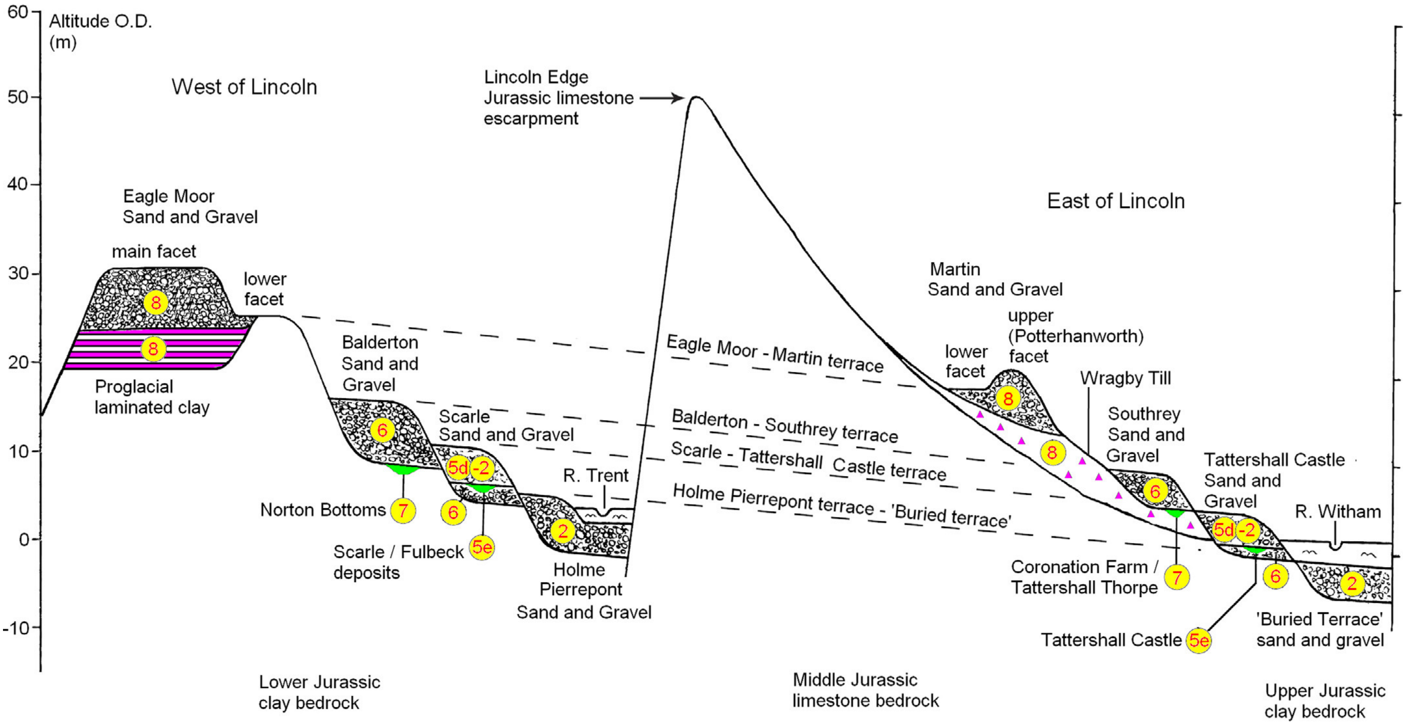
Leeds
Bielsbeck
Kingston-upon-Hull
Ouse
Humber
South Pennines
Scunthorpe
Kirmington
Ancholme
Doncaster
Idle
Manchester
Bawtry
Gainsborough
Sheffield
Hope
Castleton
Hathersage
Marton
Sturton by Stow
Retford
Welton-le-Wold
Langworth
Lincolnshire Wolds
North Sea
Peak District
Derwent
Chesterfield
Creswell Crags
Low Marnham
Fiskerton
Lincoln
Bain
Cheshire Plain
Great Rowsley
Lathkill Dale
Matlock
South Scarle
Potterhanworth
Witham
Dove
Cromwell
Kidsgrove
Wetton Mill
Greet
Newark-on-Trent
Slea
Tattershall
Ambergate
Erewash
Leen
Fiskerton
Madeley
Stoke-on-Trent
Churnet
Ashbourne
Belper
East Stoke
Rauceby
Gelston
Sleaford
Boston
Woore
Duffield
Little Eaton
Nottingham
Ancaster
The Wash
Whitmore
Somersal Herbert
Derby
Holme Pierrepont
Smite
Grantham
Market Drayton
Blithe
Trent
Uttoxeter
Willington
East Leake
Eccleshall
Sow
Barton-under-Needwood
Barrow-on-Trent
Hathern
Sproxton
Stafford
Burton-on-Trent
Melton Mowbray
South Witham
Fen Basin
Gnosall
Penk
Rugeley
Loughborough
Quorn
Castle Bytham
Thurlby
Newport
King's Bromley
Alrewas
Charnwood Forest
Wreake
Great Dalby
Witham-on-the-Hill
Wrekin Hill
Four Ashes
Wanlip
Wisbech
Shouldham
Ironbridge
Cannock Chase
Tame
Leicester
Thurmaston
Leicestershire Wolds
Peterborough
Huncote
Hinckley
Soar
Bridgnorth
Lincolnshire / North Sea ice lobe (~21-17 ka)
Birmingham
Coleshill
Wolvey
Cheshire / Irish Sea ice lobe (>~25 ka)
Coventry
Bar Hill-Whitchurch-Wrexham moraine (? ~21 ka)
Blythe
Brandon
Ancaster, Lincoln, and Humber gaps
Bewdley
0 kilometres 20
Waverley Wood
Uplift modelling sites
Mildenhall

Figure 1



Altitude O.D. (m)
River Trent long profile (west)
West
Blenheim Farm, Etwall
Allenton terrace of R. Derwent
Chellaston
Wilford Hill
East Leake
Radcliffe-on-Trent cliffs
Kneeton
Trent Trench
Sa
Et
Eg
Be
HP
C
B
Burton-on-Trent
WILLINGTON uplift model
Shardlow
Nottingham
NOTTINGHAM uplift model
Newark-on-Trent
Distance (km)
(a)
Altitude O.D. (m)
River Trent long profile (east)
East
Trent Trench
Gelston
Caythorpe Heath
Ermine Street Pit
Sentinel Wood
North Rauceby
Rauceby gravel
(R. Witham)
Fulbeck gravel
Belton gravel
Ancaster Gap col
Norton Bottoms
Coronation Farm, Southrey
Potterhanworth
TATTERSHALL uplift model
Tattershall Thorpe
Langton Hill
Welsyke Wood
Scrivelsby
Thornton
Tower Farm
Terraces of R. Bain
EMu
EMl
Mu
Ml
Ba
So
S
TC
BT
TT
B
HP
Newark-on-Trent
Modern Trent
Lincoln
Woodhall Spa
Boston
Distance (km)
(b)

Figure 2



Altitude O.D. (m)
60
50
40
30
20
10
0
-10
West of Lincoln
Lincoln Edge Jurassic limestone escarpment
East of Lincoln
Eagle Moor Sand and Gravel
main facet
lower facet
8
8
Proglacial laminated clay
Balderton Sand and Gravel
6
Scarle Sand and Gravel
5d
-2
R. Trent
Norton Bottoms
7
6
Scarle / Fulbeck deposits
5e
2
Holme Pierrepont Sand and Gravel
Eagle Moor - Martin terrace
Balderton - Southrey terrace
Scarle - Tattershall Castle terrace
Holme Pierrepont terrace - 'Buried terrace'
Martin Sand and Gravel
lower facet
upper (Potterhanworth) facet
Wragby Till
Southrey Sand and Gravel
Tattershall Castle Sand and Gravel
R. Witham
Coronation Farm / Tattershall Thorpe
Tattershall Castle
'Buried Terrace' sand and gravel
Lower Jurassic clay bedrock
Middle Jurassic limestone bedrock
Upper Jurassic clay bedrock

Figure 3

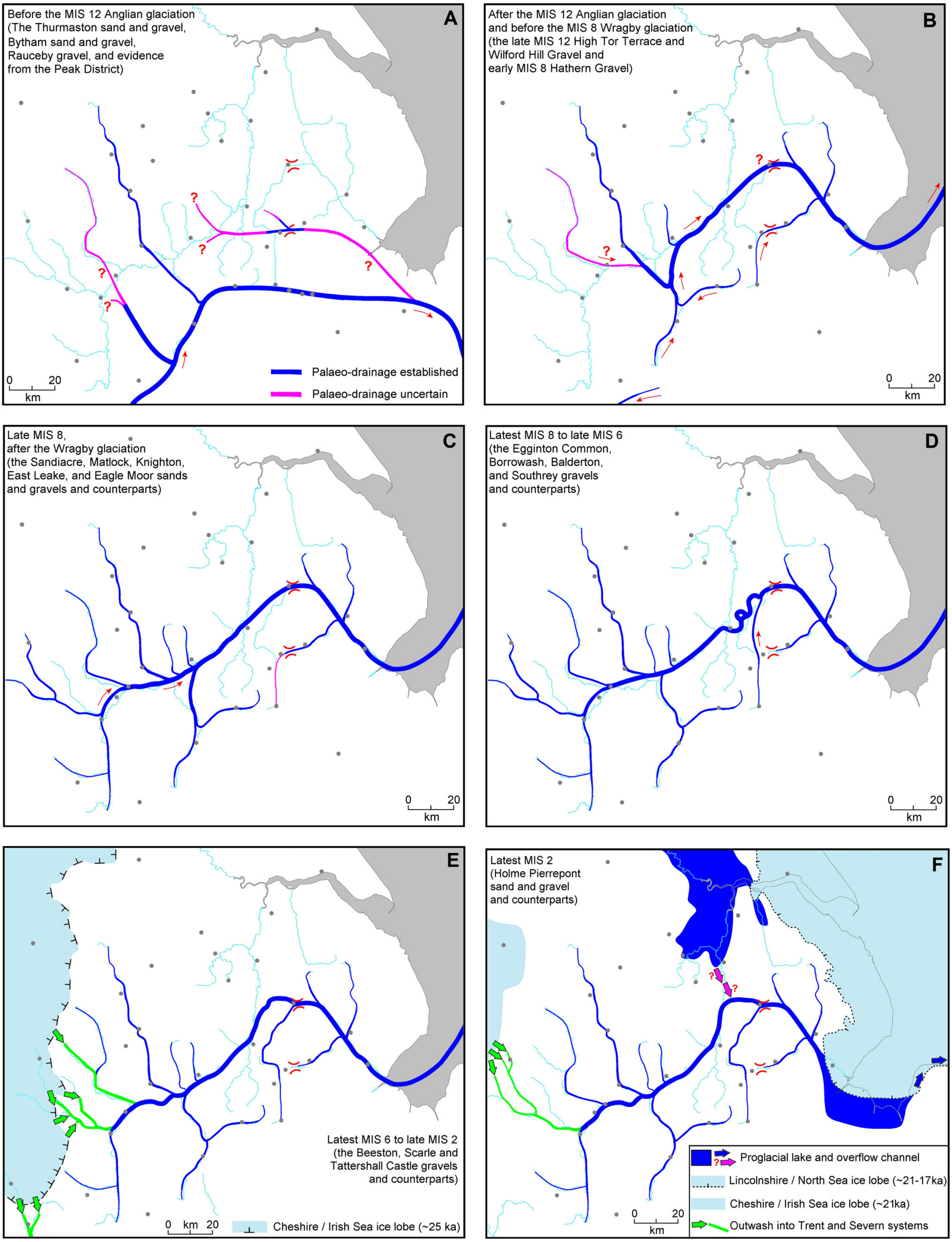

Before the MIS 12 Anglian glaciation
(The Thurmaston sand and gravel,
Bytham sand and gravel,
Rauceby gravel, and evidence
from the Peak District)
A
Palaeo-drainage established
Palaeo-drainage uncertain
0 20 km
After the MIS 12 Anglian glaciation
and before the MIS 8 Wragby glaciation
(the late MIS 12 High Tor Terrace and
Wilford Hill Gravel and
early MIS 8 Hathern Gravel)
B
0 20 km
Late MIS 8,
after the Wragby glaciation
(the Sandiacre, Matlock, Knighton,
East Leake, and Eagle Moor sands
and gravels and counterparts)
C
0 20 km
Latest MIS 8 to late MIS 6
(the Egginton Common,
Borrowash, Balderton,
and Southrey gravels
and counterparts)
D
0 20 km
E
Latest MIS 6 to late MIS 2
(the Beeston, Scarle and
Tattershall Castle gravels
and counterparts)
0 km 20
Cheshire / Irish Sea ice lobe (~25 ka)
Latest MIS 2
(Holme Pierrepont
sand and gravel
and counterparts)
F
Proglacial lake and overflow channel
Lincolnshire / North Sea ice lobe (~21-17ka)
Cheshire / Irish Sea ice lobe (~21ka)
Outwash into Trent and Severn systems
0 20 km

Figure 4

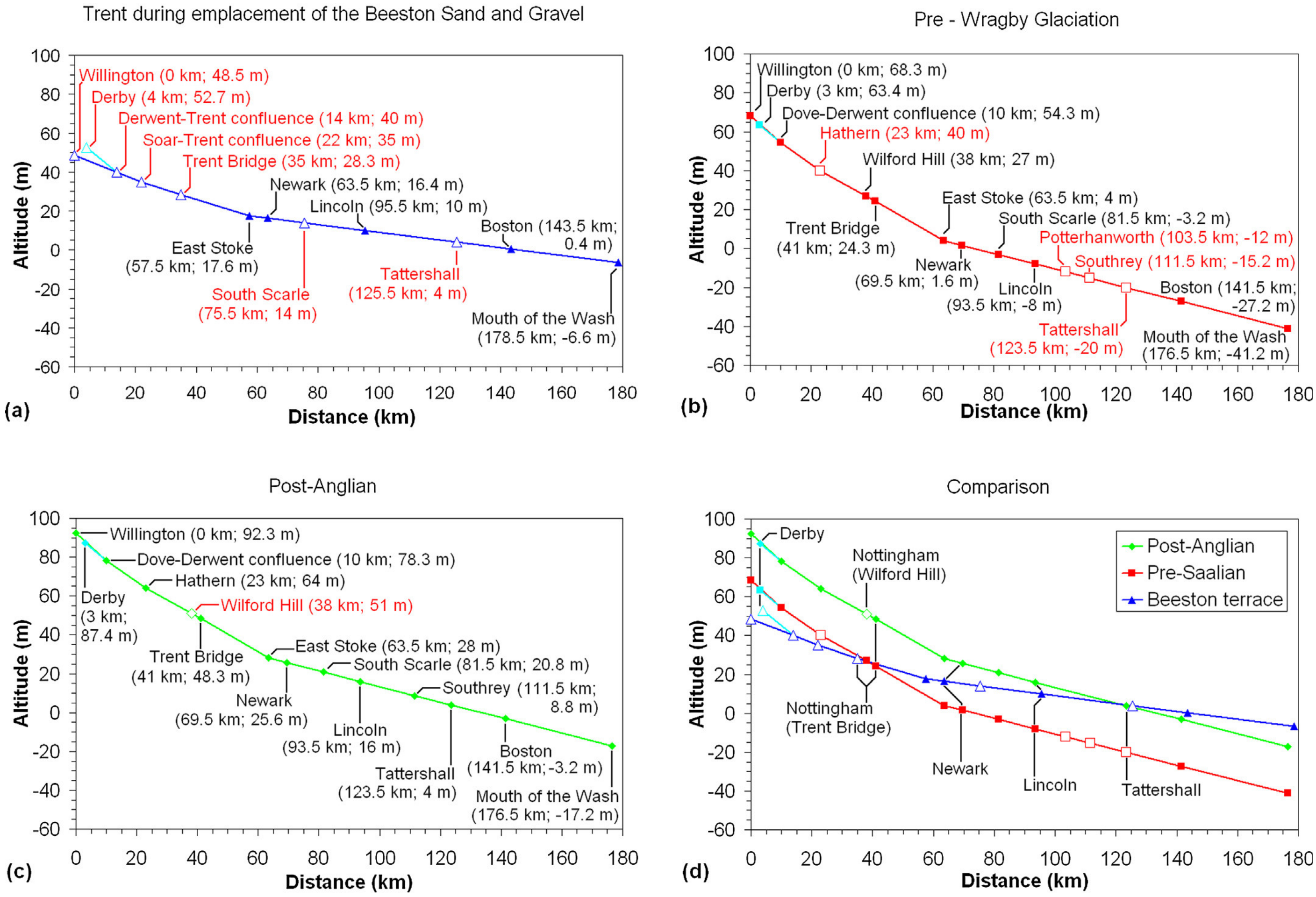

Trent during emplacement of the Beeston Sand and Gravel
Willington (0 km; 48.5 m)
Derby (4 km; 52.7 m)
Derwent-Trent confluence (14 km; 40 m)
Soar-Trent confluence (22 km; 35 m)
Trent Bridge (35 km; 28.3 m)
Newark (63.5 km; 16.4 m)
Lincoln (95.5 km; 10 m)
Boston (143.5 km; 0.4 m)
East Stoke (57.5 km; 17.6 m)
South Scarle (75.5 km; 14 m)
Tattershall (125.5 km; 4 m)
Mouth of the Wash (178.5 km; -6.6 m)
Altitude (m)
Distance (km)
(a)
Pre - Wragby Glaciation
Willington (0 km; 68.3 m)
Derby (3 km; 63.4 m)
Dove-Derwent confluence (10 km; 54.3 m)
Hathern (23 km; 40 m)
Wilford Hill (38 km; 27 m)
Trent Bridge (41 km; 24.3 m)
East Stoke (63.5 km; 4 m)
South Scarle (81.5 km; -3.2 m)
Potterhanworth (103.5 km; -12 m)
Southrey (111.5 km; -15.2 m)
Newark (69.5 km; 1.6 m)
Lincoln (93.5 km; -8 m)
Boston (141.5 km; -27.2 m)
Tattershall (123.5 km; -20 m)
Mouth of the Wash (176.5 km; -41.2 m)
Altitude (m)
Distance (km)
(b)
Post-Anglian
Willington (0 km; 92.3 m)
Dove-Derwent confluence (10 km; 78.3 m)
Hathern (23 km; 64 m)
Derby (3 km; 87.4 m)
Wilford Hill (38 km; 51 m)
Trent Bridge (41 km; 48.3 m)
East Stoke (63.5 km; 28 m)
South Scarle (81.5 km; 20.8 m)
Newark (69.5 km; 25.6 m)
Southrey (111.5 km; 8.8 m)
Lincoln (93.5 km; 16 m)
Boston (141.5 km; -3.2 m)
Tattershall (123.5 km; 4 m)
Mouth of the Wash (176.5 km; -17.2 m)
Altitude (m)
Distance (km)
(c)
Comparison
Derby
Nottingham (Wilford Hill)
Post-Anglian
Pre-Saalian
Beeston terrace
Nottingham (Trent Bridge)
Newark
Lincoln
Tattershall
Altitude (m)
Distance (km)
(d)

Figure 5

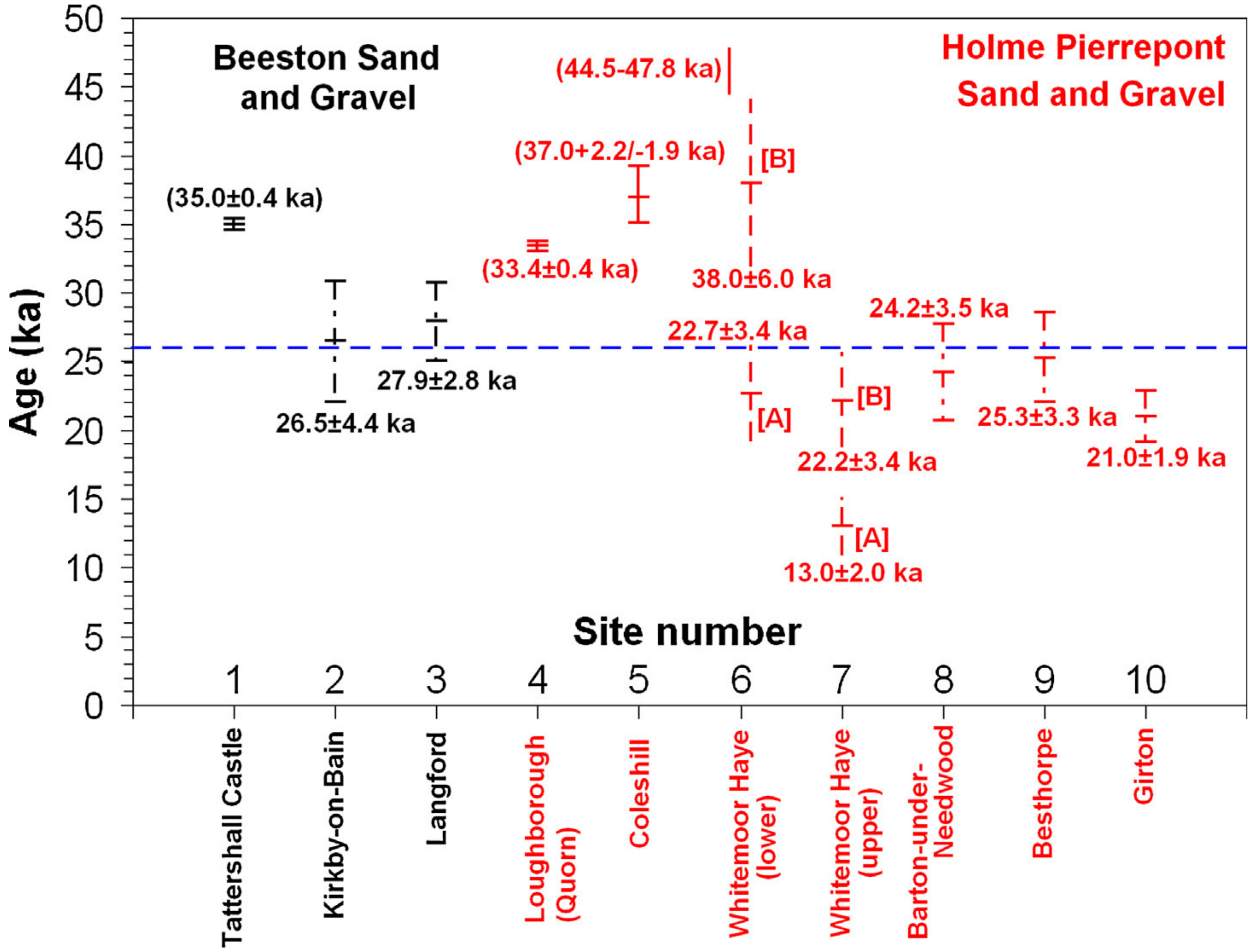

Beeston Sand and Gravel
Holme Pierrepont Sand and Gravel
Age (ka)
Site number
(35.0±0.4 ka)
26.5±4.4 ka
27.9±2.8 ka
(33.4±0.4 ka)
(37.0+2.2/-1.9 ka)
(44.5-47.8 ka)
[B]
38.0±6.0 ka
22.7±3.4 ka
[A]
[B]
22.2±3.4 ka
[A]
13.0±2.0 ka
24.2±3.5 ka
25.3±3.3 ka
21.0±1.9 ka
Tattershall Castle
Kirkby-on-Bain
Langford
Loughborough (Quorn)
Coleshill
Whitemoor Haye (lower)
Whitemoor Haye (upper)
Barton-under-Needwood
Besthorpe
Girton

Figure 6

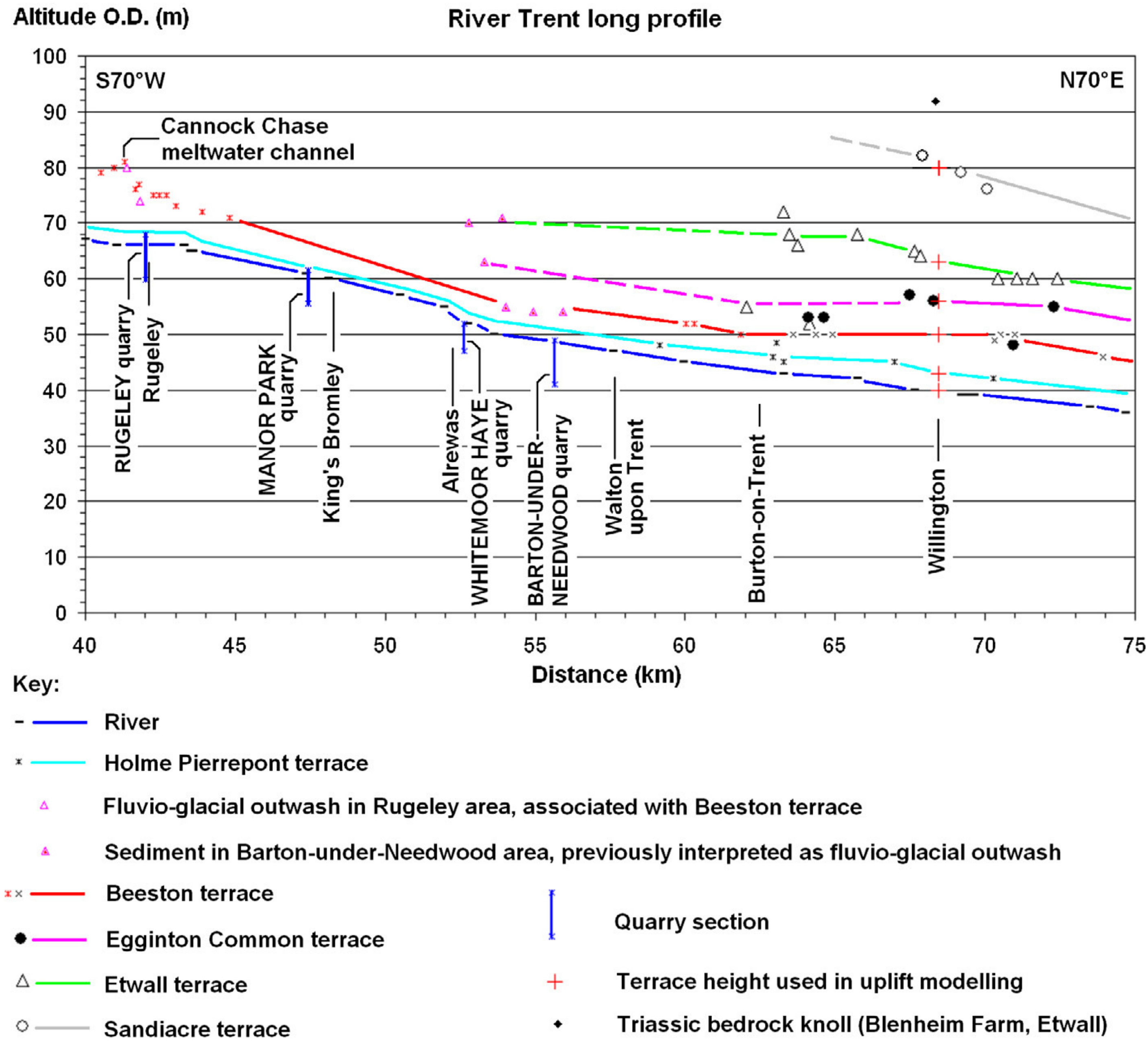

Altitude O.D. (m)
River Trent long profile
S70°W
N70°E
Cannock Chase meltwater channel
RUGELEY quarry
Rugeley
MANOR PARK quarry
King's Bromley
Alrewas
WHITEMOOR HAYE quarry
BARTON-UNDER-NEEDWOOD quarry
Walton upon Trent
Burton-on-Trent
Willington
Distance (km)
Key:
River
Holme Pierrepont terrace
Fluvio-glacial outwash in Rugeley area, associated with Beeston terrace
Sediment in Barton-under-Needwood area, previously interpreted as fluvio-glacial outwash
Beeston terrace
Egginton Common terrace
Etwall terrace
Sandiacre terrace
Quarry section
Terrace height used in uplift modelling
Triassic bedrock knoll (Blenheim Farm, Etwall)

Figure 7

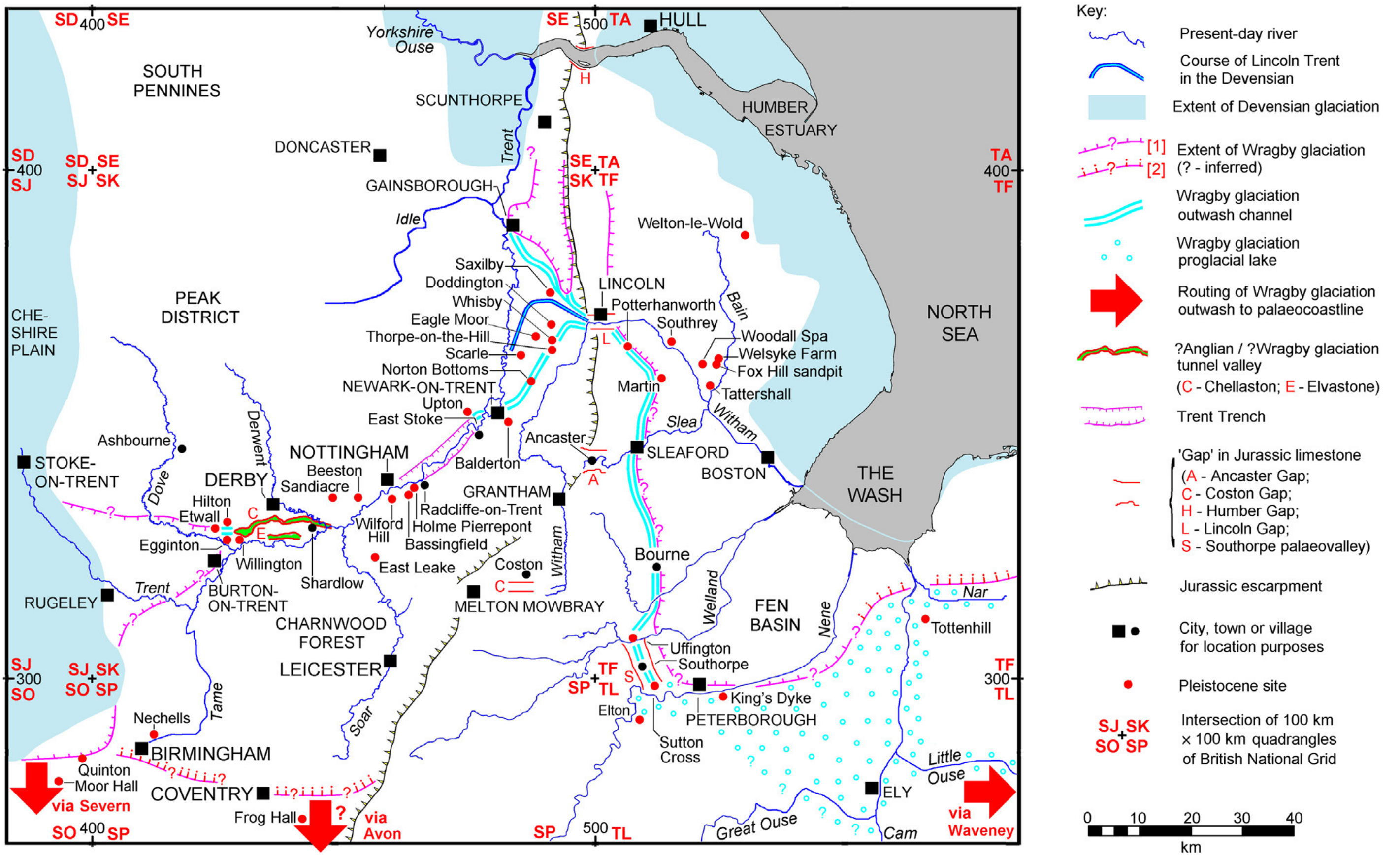


Figure 8

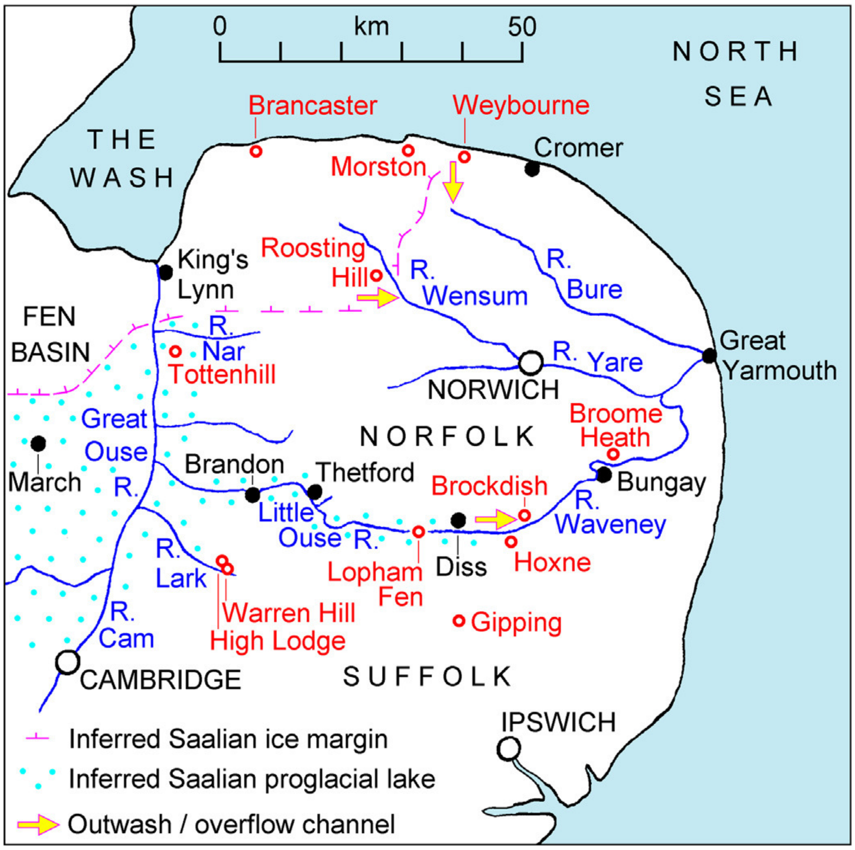
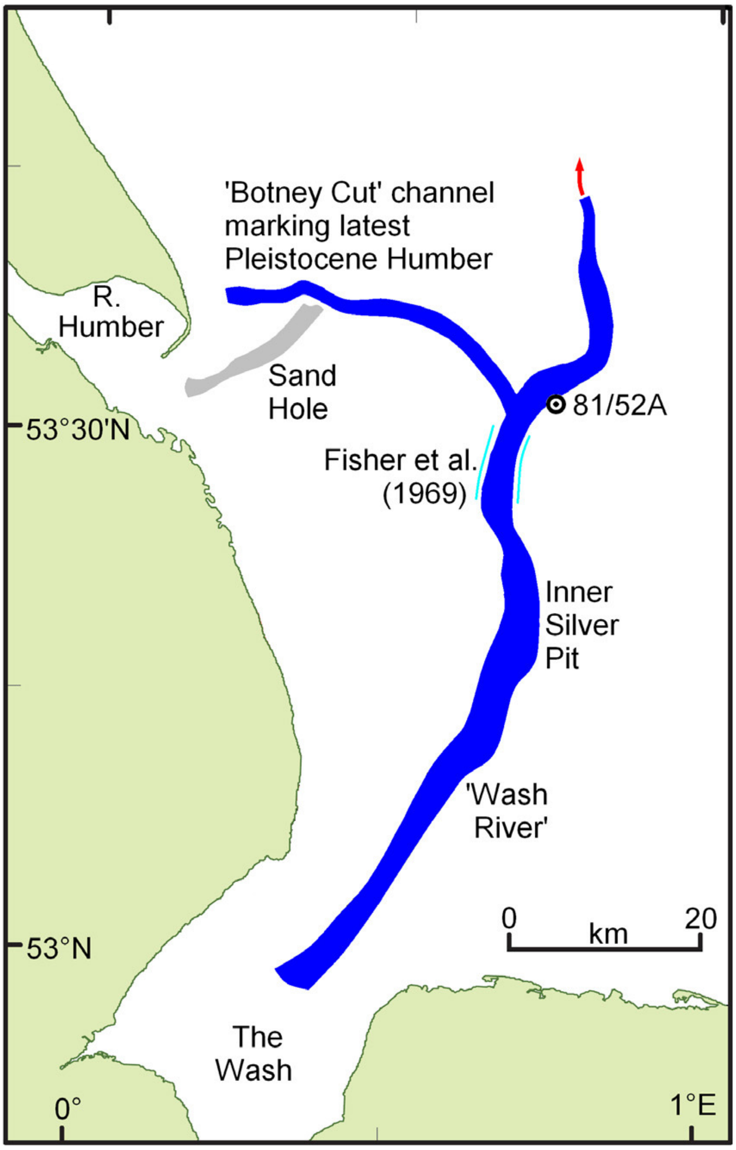


Figure 9


'Botney Cut' channel
marking latest
Pleistocene Humber
R.
Humber
Sand
Hole
81/52A
53°30'N
Fisher et al.
(1969)
Inner
Silver
Pit
'Wash
River'
0
km
20
53°N
The
Wash
0°
1°E
Figure 10

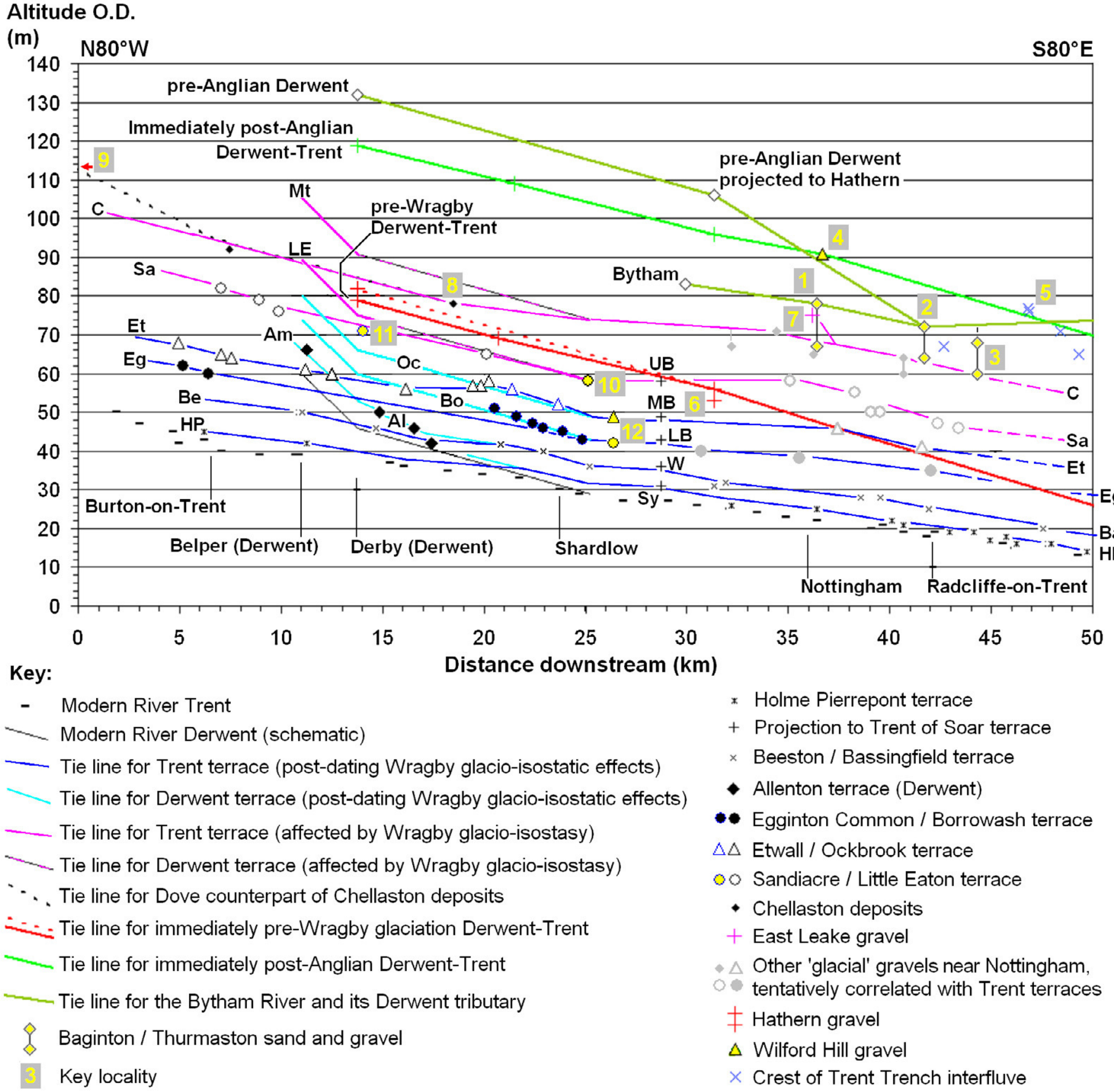
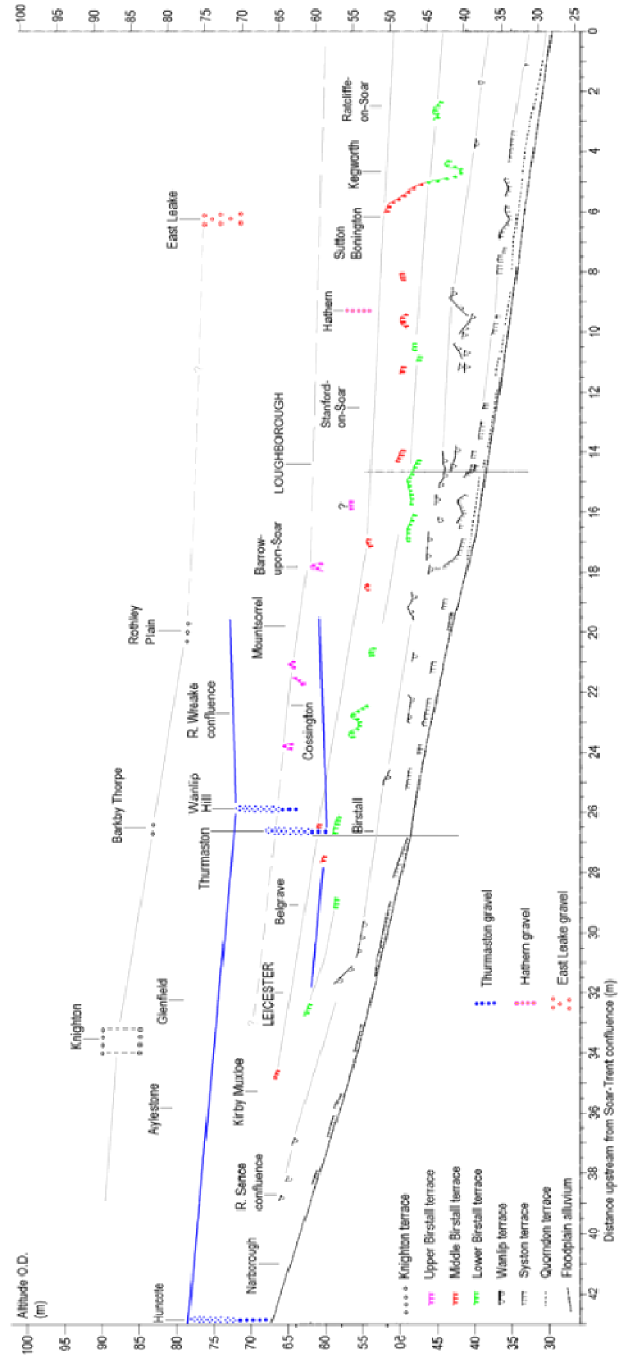
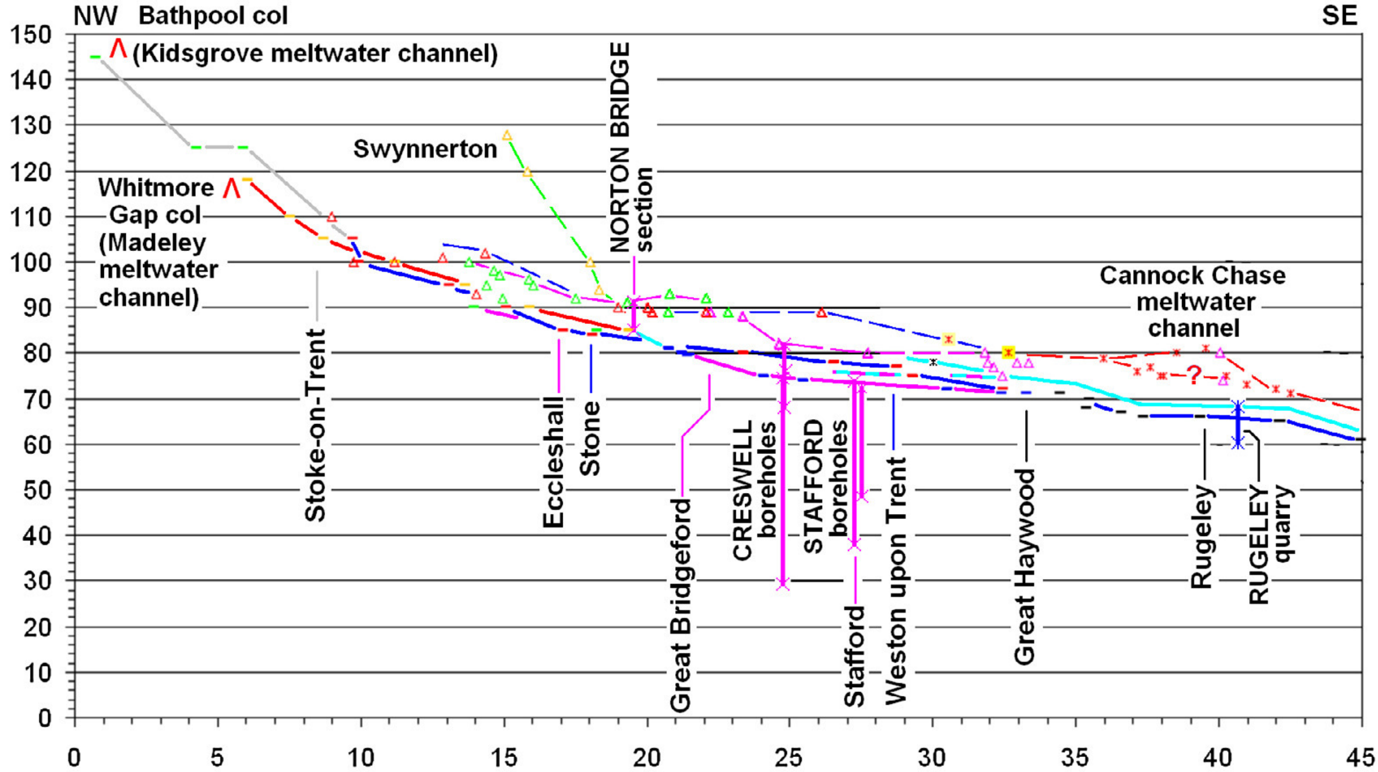

Altitude O.D. (m)
N80°W
S80°E
pre-Anglian Derwent
Immediately post-Anglian Derwent-Trent
pre-Anglian Derwent projected to Hathern
pre-Wragby Derwent-Trent
Bytham
Mt
C
LE
Sa
Et
Am
Eg
Oc
Be
Bo
HP
Al
UB
MB
LB
W
Sy
Ba
Burton-on-Trent
Belper (Derwent)
Derby (Derwent)
Shardlow
Nottingham
Radcliffe-on-Trent
Distance downstream (km)
Key:
Modern River Trent
Modern River Derwent (schematic)
Tie line for Trent terrace (post-dating Wragby glacio-isostatic effects)
Tie line for Derwent terrace (post-dating Wragby glacio-isostatic effects)
Tie line for Trent terrace (affected by Wragby glacio-isostasy)
Tie line for Derwent terrace (affected by Wragby glacio-isostasy)
Tie line for Dove counterpart of Chellaston deposits
Tie line for immediately pre-Wragby glaciation Derwent-Trent
Tie line for immediately post-Anglian Derwent-Trent
Tie line for the Bytham River and its Derwent tributary
Baginton / Thurmaston sand and gravel
Key locality
Holme Pierrepont terrace
Projection to Trent of Soar terrace
Beeston / Bassingfield terrace
Allenton terrace (Derwent)
Egginton Common / Borrowash terrace
Etwall / Ockbrook terrace
Sandiacre / Little Eaton terrace
Chellaston deposits
East Leake gravel
Other 'glacial' gravels near Nottingham, tentatively correlated with Trent terraces
Hathern gravel
Wilford Hill gravel
Crest of Trent Trench interfluve

Figure 11



Altitude O.D. (m)
Distance upstream from Soar-Trent confluence (m)
Huncote
Narborough
R. Sence confluence
Aylestone
Knighton
Kirby Muxloe
Glenfield
LEICESTER
Belgrave
Barkby Thorpe
Thurmaston
Wanlip Hill
Birstall
R. Wreake confluence
Cossington
Rothley Plain
Mountsorrel
Barrow-upon-Soar
LOUGHBOROUGH
Stanford-on-Soar
Hathern
Sutton Bonington
East Leake
Kegworth
Ratcliffe-on-Soar
Knighton terrace
Upper Birstall terrace
Middle Birstall terrace
Lower Birstall terrace
Wanlip terrace
Syston terrace
Quorndon terrace
Floodplain alluvium
Thurmaston gravel
Hathern gravel
East Leake gravel

Figure 12



Key:

Upper Trent downstream of Sow confluence
- Modern river
- Holme Pierrepont Terrace
- Beeston Terrace
- Outwash deposits in Rugeley area

R. Sow upstream of Meece Brook confluence
- Modern river
- Surface of outwash

Etruria Valley / Cliffe Vale
- Modern valley floor (Fowlea Brook)

Λ Col of glacial meltwater channel

Upper Trent upstream of Sow confluence
- Modern river
- Surface of outwash
- Holme Pierrepont Terrace

R. Sow downstream of Meece Brook confluence
- Modern river
- Surface of outwash
- Lowest outwash terrace flat

Meece Brook
- Modern valley floor

Swynnerton outwash deposits
- Surface of outwash

Figure 13

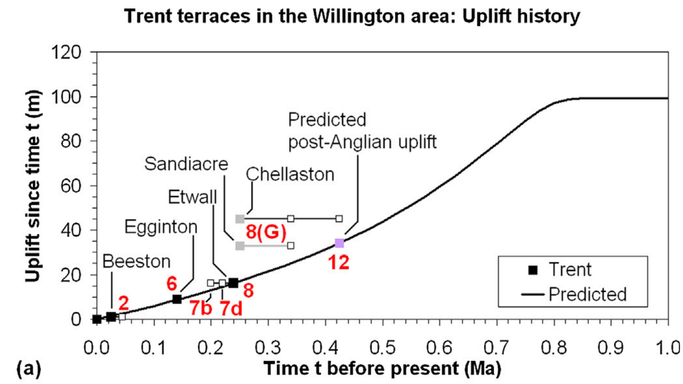


Figure 14

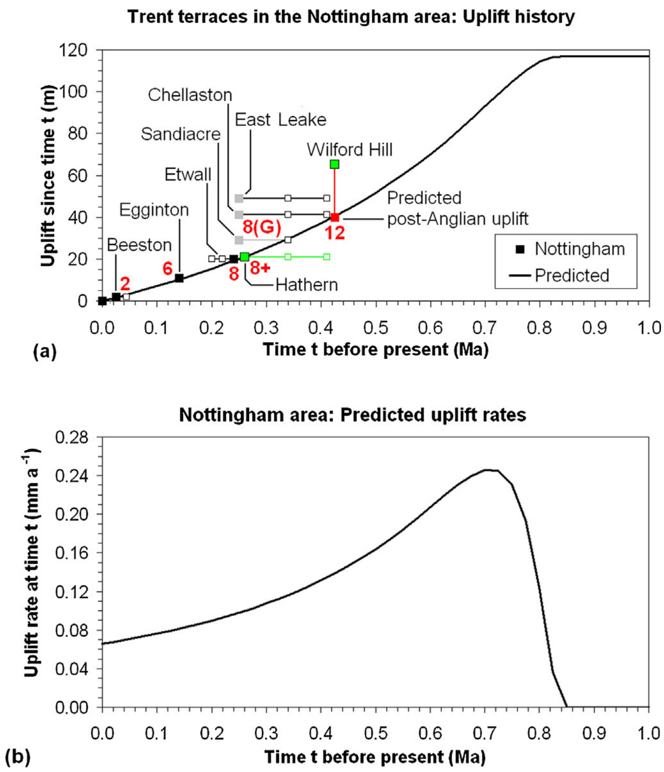


Figure 15

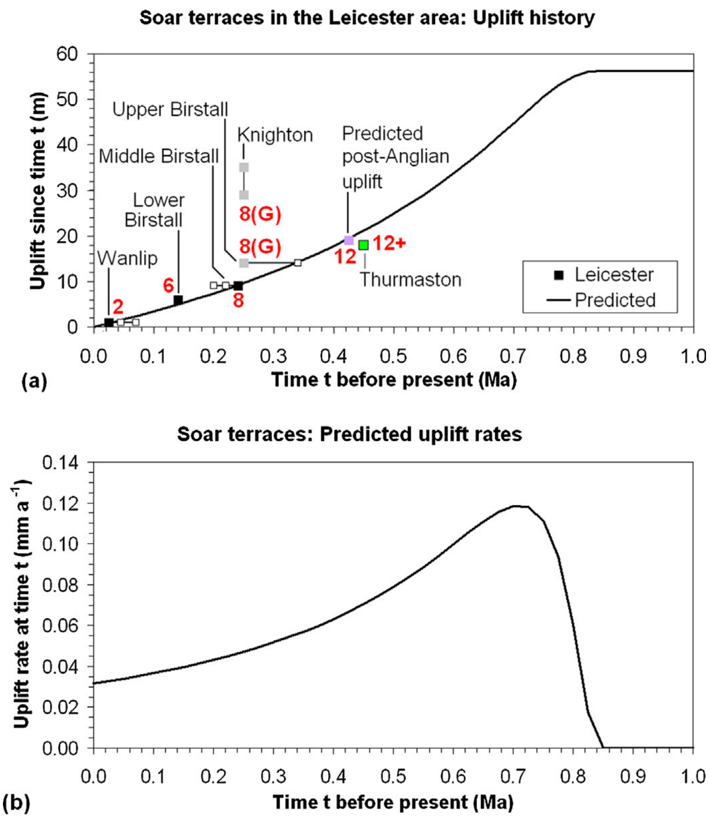


Figure 16

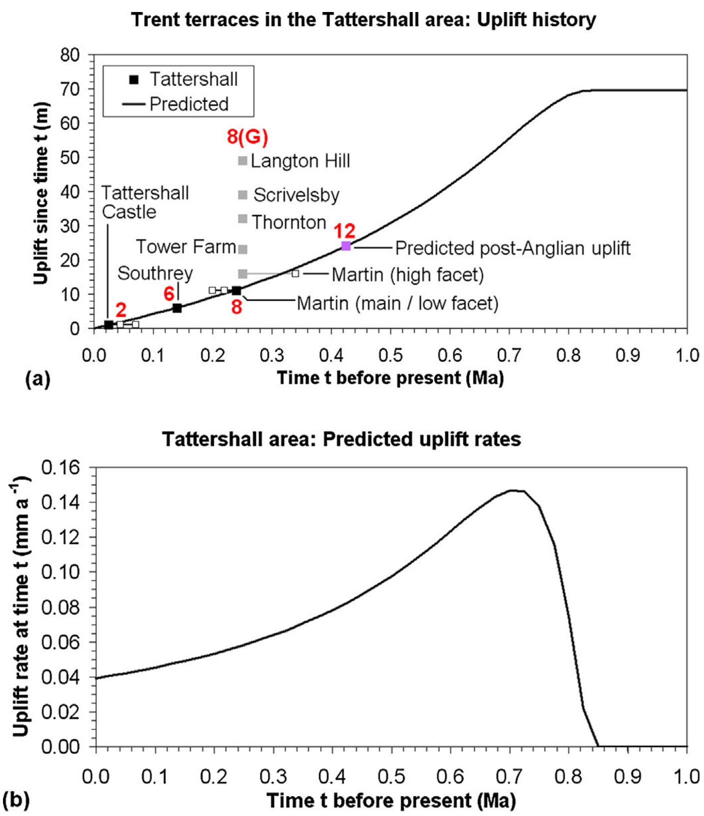
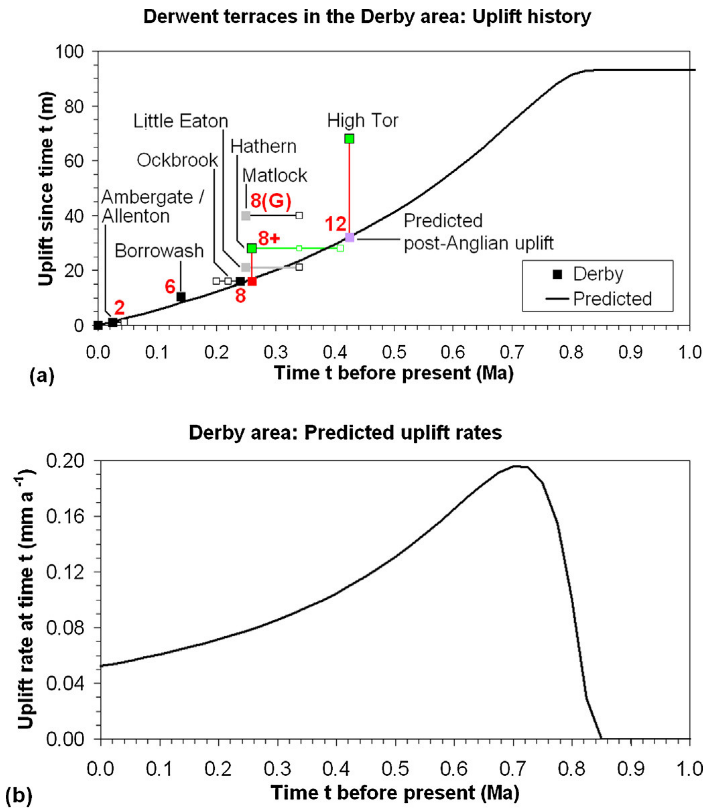


Figure 17



Derwent terraces in the Derby area: Uplift history
Uplift since time t (m)
Time t before present (Ma)
Little Eaton
High Tor
Hathern
Ockbrook
Matlock
8(G)
Ambergate /
Allenton
12
Predicted
post-Anglian uplift
8+
Borrowash
6
8
2
Derby
Predicted
(a)
Derby area: Predicted uplift rates
Uplift rate at time t (mm a-1)
Time t before present (Ma)
(b)

Figure 18

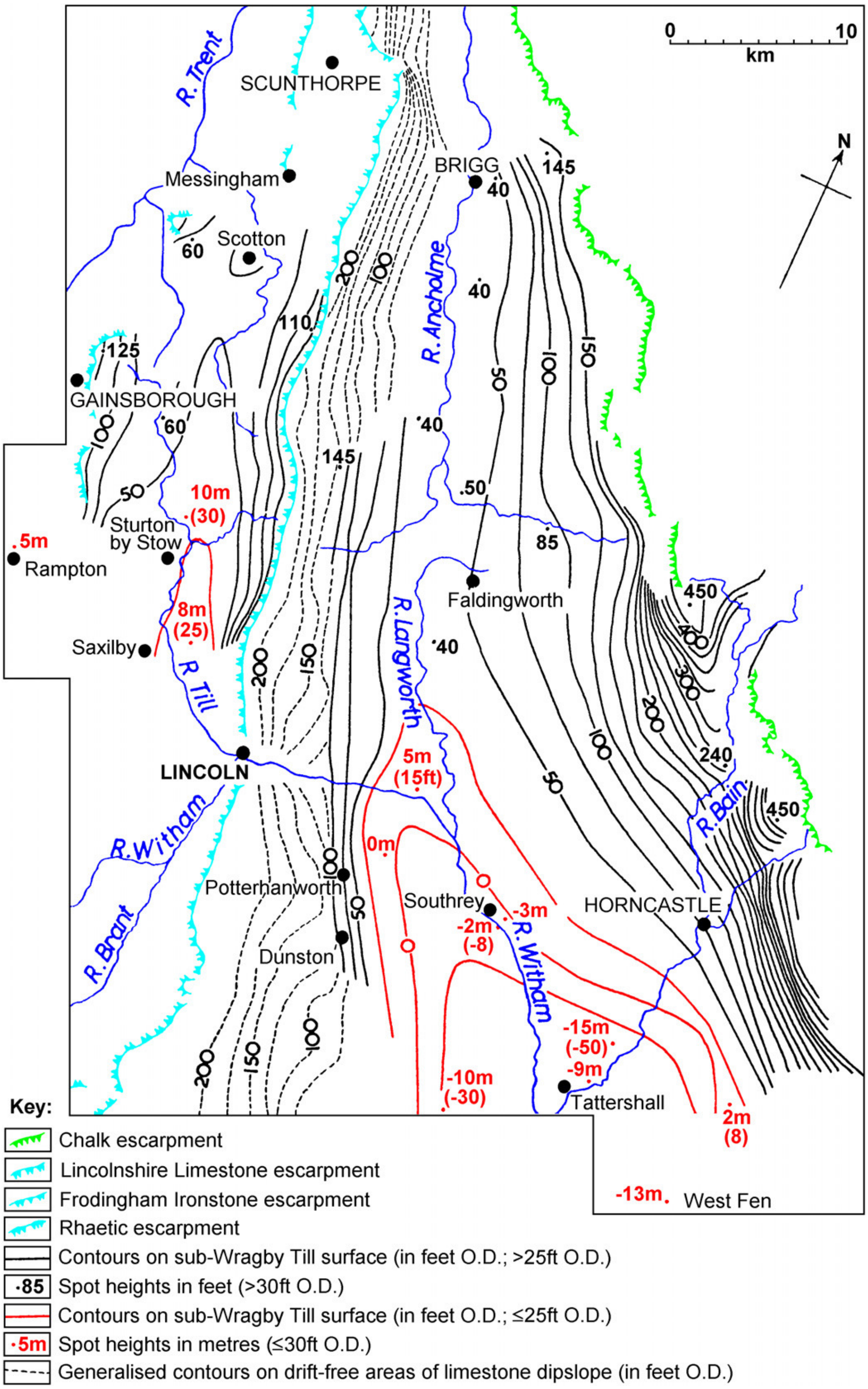
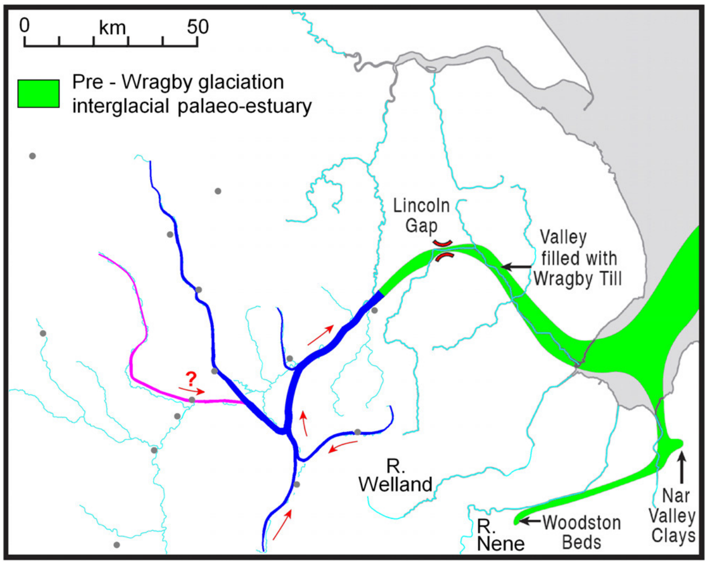

0
10
km
N
R. Trent
SCUNTHORPE
Messingham
BRIGG
145
40
Scotton
60
200
100
R. Ancholme
40
110
125
GAINSBOROUGH
60
100
50
150
100
40
145
.50
50
10m
(30)
Sturton
by Stow
.5m
Rampton
85
Faldingworth
450
8m
(25)
Saxilby
R Till
200
150
R. Langworth
40
400
300
200
100
5m
(15ft)
50
240
LINCOLN
R. Witham
R. Bain
450
0m
100
Potterhanworth
50
Southrey
-3m
-2m
(-8)
R. Witham
HORNCASTLE
R. Brant
Dunston
-15m
(-50)
-9m
200
150
100
-10m
(-30)
Tattershall
2m
(8)
-13m.
West Fen
Key:
Chalk escarpment
Lincolnshire Limestone escarpment
Frodingham Ironstone escarpment
Rhaetic escarpment
Contours on sub-Wragby Till surface (in feet O.D.; >25ft O.D.)
·85 Spot heights in feet (>30ft O.D.)
Contours on sub-Wragby Till surface (in feet O.D.; ≤25ft O.D.)
·5m Spot heights in metres (≤30ft O.D.)
Generalised contours on drift-free areas of limestone dipslope (in feet O.D.)

Figure 19



0
km
50
Pre - Wragby glaciation
interglacial palaeo-estuary
Lincoln
Gap
Valley
filled with
Wragby Till
?
R.
Welland
R.
Nene
Woodston
Beds
Nar
Valley
Clays

Figure 20