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Corresponding Author: Prof. David R Bridgland,

Corresponding Author's Institution: Durham University

First Author: David R Bridgland

Order of Authors: David R Bridgland; Rob Westaway

Abstract: Late Cenozoic (and especially Quaternary) fluvial deposits and related landforms provide valuable information about landscape evolution, not just in terms of changing drainage patterns but also documenting changes in topography and relief. Recently compiled records from river systems worldwide have shed much light on this subject, particularly records of terrace sequences, although other types of fluvial archive can be equally informative. Terraces are especially valuable if they can be dated with reference to biostratigraphy, geochronology or by other means. The various data accumulated support the hypothesis that the incision observed from river terraces has been a response to progressive uplift during the Late Cenozoic. This has not occurred everywhere, however. Stacked fluvial sequences have formed in subsiding depocentres and have greater potential for surviving to become part of the longer-term geological record. More enigmatic are regions in the ancient cores of continents (cratons), which show little indication of sustained uplift or subsidence, with fluvial deposits of various ages occurring within a restricted range of elevation with respect to the valley floor. In areas of dynamic crust that were glaciated during the Last Glacial Maximum post-glacial river valleys are typically incised and often terraced in a similar way to valleys on post-Precambrian crust elsewhere, although the terraces and gorges in these systems are very much younger (~15 ka) and therefore the processes have been considerably more rapid. This paper is illustrated with various case-study examples of these different types of archives and discusses the implications of each for regional landscape evolution.

Quaternary fluvial archives and landscape evolution: a global synthesis

David R. Bridgland
Department of Geography, Durham University, DH1 3LE, UK
E-mail: d.r.bridgland@durham.ac.uk

Rob Westaway
School of Engineering, University of Glasgow, Glasgow G12 8QQ, UK
E-mail: robert.westaway@gla.ac.uk;

Also at:
NIReS, Newcastle University, Newcastle-upon-Tyne NE1 7RU, UK
E-mail: r.w.c.westaway@ncl.ac.uk

ABSTRACT

Late Cenozoic (and especially Quaternary) fluvial deposits and related landforms provide valuable information about landscape evolution, not just in terms of changing drainage patterns but also documenting changes in topography and relief. Recently compiled records from river systems worldwide have shed much light on this subject, particularly records of terrace sequences, although other types of fluvial archive can be equally informative. Terraces are especially valuable if they can be dated with reference to biostratigraphy, geochronology or by other means. The various data accumulated support the hypothesis that the incision observed from river terraces has been a response to progressive uplift during the Late Cenozoic. This has not occurred everywhere, however. Stacked fluvial sequences have formed in subsiding depocentres and have greater potential for surviving to become part of the longer-term geological record. More enigmatic are regions in the ancient cores of continents (cratons), which show little indication of sustained uplift or subsidence, with fluvial deposits of various ages occurring within a restricted range of elevation with respect to the valley floor. In areas of dynamic crust that were glaciated during the Last Glacial Maximum post-glacial river valleys are typically incised and often terraced in a similar way to valleys on post-Precambrian crust elsewhere, although the terraces and gorges in these systems are very much younger (~15 ka) and therefore the processes have been considerably more rapid. This paper is illustrated with various case-study examples of these different types of archives and discusses the implications of each for regional landscape evolution.

KEYWORDS

Quaternary; Late Cenozoic; river terraces; landscape evolution; drainage development; uplift

1. Introduction

Sediments from fluvial environments represent an important part of the Quaternary terrestrial record, within which they occur mainly (although not solely) as aggradational river-terrace deposits. These can provide important frameworks for

Quaternary lithostratigraphy, especially where they are well dated with reference to their fossil content or by means of geochronological techniques (e.g., Maddy et al., 1991, 1995; Antoine et al., 2000, 2007; Bridgland, 2000; Bridgland and Maddy, 2002; Nott et al., 2002; Bridgland et al., 2004a; Cordier et al., 2012; Bridgland and Westaway 2008a, b; Westaway et al., 2009a; see also below, section 1.1). The longest sequences, sometimes extending back to the pre-Quaternary, are preserved in regions beyond the reach of the Pleistocene ice sheets, since a common effect of glaciation has been to remove all pre-existing superficial deposits. In areas within the limit of the most extensive glaciations, but beyond the margin of Marine Isotope Stage (MIS) 2 (Last Glacial) ice sheets, the river-terrace record can provide a valuable means for unravelling the history of multiple glacial advances, as in the River Trent in central England (White et al., 2010; Bridgland et al., 2014a, in press; Westaway et al., in press; Rose, in press). Inside the Last Glacial limit, river valleys generally have a terraced form, not dissimilar to those outside the limit, although the terrace sediments invariably post-date the start of MIS 2 deglaciation (Howard et al., 2000a, b; Bridgland et al., 2010, 2011).

NW Europe, including Britain, boasts some of the most important Pleistocene terrace sequences globally, an example being that in the Lower Thames, in which each of the last four 100 ka climate cycles is represented (Fig. 1Ai). Other excellent examples of such archives occur (Fig. 1B–D) in France (Pastre, 2004; Antoine et al., 2007; Cordier et al., 2005, 2006, 2012), Germany (Mania, 1995; Bibus and Wesler, 1995; Schreve and Bridgland, 2002; Bridgland et al., 2004b) and the Netherlands (Van den Berg and van Hoof, 2001; Westaway, 2001). Even better records are known from the rivers flowing southwards to the Black Sea through Ukraine, which have highly informative sequences, dated with reference to a well-established biostratigraphical and magnetostratigraphical framework, that extend back to the Miocene (Matoshko et al., 2002, 2004; Fig. 2). Further south, in the Mediterranean region, there are important and recently documented terrace records in Portugal (Cunha et al., 2005, 2008; Martins et al., 2010a), Spain (Stokes and Mather, 2003; Santisteban and Schulte, 2007; Meikle et al., 2010), Turkey (Demir et al., 2004, 2007a, b, 2012; Bridgland et al., 2007a; Seyrek et al., 2008) and Syria (Demir et al., 2007a; Abou Romieh et al., 2009; Bridgland et al., 2012). Further afield the value of river-terrace sequences is less well established, although there is a long record of terrace studies in the USA, notably from the Mississippi (Saucier, 1996; Blum et al., 2000), Ohio (Westaway, 2007), Susquehanna (Pazzaglia and Gardner, 1993) and Platte (Reed et al., 1965; Osterkamp et al., 1987). Furthermore, there has been considerable recent effort to document the fluvial terraces of the Colorado and the history of down-cutting by this river, with particular emphasis on the formation of the Grand Canyon (e.g., Pederson et al., 2006, 2013; Karlstrom et al., 2008; Lee et al., 2013; see below). The records from the major Chinese rivers are also important (Li et al., 1997; Pan et al., 2009, 2011; Yang, 2006; Vandenberghe et al., 2011; Zhu et al., 2014) and there are records from the southern hemisphere (Bibus, 1983; Bull and Kneupfer, 1987; Bull, 1991; Latrubesse et al., 1997; Hattingh and Rust, 1999; Nott et al., 2002; Westaway, 2006a). This includes exceptional records from Patagonia, where the fluvial archive extends back into the Neogene and is inter-related with evidence for ancient glaciation, preserved by interbedding with volcanic deposits (Mercer, 1976; Martinez and Coronato, 2008).

Much of current knowledge of such archives stems from data compilation during sequential International Geoscience (IGCP) projects: IGCP 449 'Global Correlation of Late Cenozoic Fluvial Sequences' (Bridgland et al., 2007b) and IGCP 518 'Fluvial Sequences as Archives of Landscape and Climatic Evolution in the Late Cenozoic (Westaway et al., 2009a), both undertaken under the aegis of the Fluvial Archives Group (FLAG). For recent reviews of the results from these projects, which demonstrate patterns of variability between Late Cenozoic fluvial sequences in different regions, with different geological characteristics (crustal provinces), see Bridgland and Westaway (2008a, b, 2012) and Westaway et al., 2009a).

This paper is organized thematically, based on the global patterns of fluvial system evolution and style of preservation observed from the IGCP projects. Some systems will therefore appear in more than one thematic section, as the nature of their records have varied over time or between different reaches.

1.1 Dating fluvial sediments

The fluvial sequences that are best constrained in terms of age are generally those with reliable biostratigraphy (see Schreve et al., 2007), although Palaeolithic artefacts found within or in association with fluvial sediments can also provide valuable evidence of age (e.g., Bridgland et al., 2006; Westaway et al., 2006a; Mishra et al., 2007; Pettitt and White, 2012; Bridgland and White, 2014). Various geochronological methods have been applied to fluvial sequences representing the timescales under discussion here: primarily older than can be dated using radiocarbon (but see section 7). It is probably fair to say that none of the geochronological methods is as reliable for dating fluvial sediments as the best evidence from biostratigraphy, but the application of geochronology, where it corroborates biostratigraphical ages or where multiple techniques give concordant results, is of considerable value. In contrast to fluvial deposits, volcanic rocks of the same age-range can be reliably dated using modern variants of the K–Ar technique, such as the Ar–Ar method and the unspiked (or Cassignol) variant of K-Ar, which are capable of dating Middle Pleistocene samples with precision and accuracy of just a few percent, as will be noted in connection with those areas where Quaternary volcanism has led to such rocks being interbedded with fluvial sequences (e.g., Pastre, 2004; Boenigk and Frechen, 2006; Demir et al., 2007a; Seyrek et al., 2008; Westaway et al., 2009b). Uranium-series dating of carbonate inclusions, interbeds, or speleothems can, under optimal circumstances, also provide numerical ages that are both precise and accurate (e.g., Schwarcz et al., 1988; Murton et al., 2001; Mallick and Frank, 2002; Candy et al., 2004). Dating of speleothems can be relevant to the interpretation of fluvial sequences, because the chronology of the former can reflect the history of adjacent valley entrenchment (e.g., Westaway, 2009a, 2010; Bridgland et al., 2014a). Where molluscan fossils are preserved their biostratigraphical value as age indicators can be reinforced by amino-acid dating, which is a measure of protein degradation since death (Bowen et al., 1989; Penkman et al., 2011, 2013).

The only technique that can be used to measure the age of minerogenic (clastic) fluvial sediment directly is luminescence dating; variants of this technique have undergone many refinements in recent years and have been very widely used for dating the last exposure to daylight of sand grains in fluvial sediments (e.g., Murray and Wintle, 2000; Schokker et al., 2005; Briant et al., 2006; Briant and Bateman, 2009; Pawley et al., 2010; Schreve et al., 2013). Its range is dependant on natural radiation doses; in low-dose situations it can provide dates for quartz grains approaching 0.5 Ma (e.g., Pawley et al., 2010), with a promise of longer ranges from feldspar in the future (cf. Wallinga et al., 2001).

2. River terrace sequences: archives of uplift and climate change

There has been considerable and protracted debate in the geological and geomorphological literature about the formation of river terraces, leading to a general consensus that many have been climatically triggered (Tyràček, 1983; Vandenberghe, 1995, 2002, 2003; Bridgland, 1994; Hancock and Anderson, 2002; Starkel, 2003), often in synchrony with glacial–interglacial fluctuation. Also, crucially, there is a growing consensus that terrace formation has been enabled by long-timescale regional (epeirogenic) uplift (Van den Berg, 1994; Maddy, 1997; Maddy et al., 2000; Bridgland, 2000; Bridgland and Westaway, 2008a, b, 2012; Westaway et al., 2009a). A small number of detractors from this interpretation have sustained debate in recent years, mainly on the basis of doubting the requirement for uplift (Hancock and Anderson, 2002; Gibbard and Lewin, 2009) and/or favouring base-level (sea-level) change as a principal driver (Törnqvist and Blum, 1998; Martins et al., 2010a), often through the mechanism of knick-point recession (Rosenbloom and Anderson, 1994; Whipple and Tucker, 1999; Crosby and Whipple, 2006; Bishop, 2007; Roberts and White 2010; see below, section 8.3). It is clear that sea-level fluctuation can be an important local driver in the lower reaches of rivers debouching onto narrow continental shelves, such as the west-flowing rivers of Iberia (e.g., Martins et al., 2010a; Viveen et al., 2012, 2013). It has been widely agreed, however, that the effect of base-level change will seldom be manifest any great distance inland from coastal regions (e.g., Leopold and Bull, 1979; Schumm, 1993).

The evidence in support of uplift as an important and widespread factor in terrace formation is largely empirical. In particular, there is a clear-cut contrast between areas that can be observed to have experienced long-term subsidence and those thought to have been uplifting; the former lack a terraced landscape and fluvial deposits have accumulated in conventional stratigraphical mode, with the most recent beneath the immediate land surface and older ones becoming progressively more deeply buried with age. The fluvial records from subsiding areas of this sort are generally known from boreholes and other subsurface data. They include small basins, often fault-bounded, but are typified by large depocentres in which the accumulation of sediment acts isostatically as a positive feedback mechanism in the subsidence process. The land surface of such subsiding basins is invariable flat, often lacking clear valley geomorphology, with poor separation of multiple river systems where these exist within single basins. An example illustrating fluvial deposition at a particularly large spatial scale is the foreland region of the Himalayas, in which the thickness of sediments laid down in the Ganges system exceeds four kilometres in places (Sinha et al., 2007). The Lower Rhine provides a further example, in which the depocentre underlies the Netherlands and extends onto the continental shelf of the southern North Sea (Caston, 1977; Brunnacker et al., 1982; Ruegg, 1994; Fig. 3), while the Danube also contributes to fluvial depocentres, in particular the Pannonian Basin, beneath the plains of Hungary (Ruszkyczay-Rüdiger et al., 2005; Gábris and Nádor, 2007) and the marginal Black Sea basin (Matoshko et al., 2009). The bodies of accumulating sediment in such depocentres are likely to form the fluvial rocks of the future, since they have a much better chance of surviving to become part of the long-term geological record than the superficial terrace deposits forming elsewhere.

Powerful evidence in support of uplift as a crucial factor in terrace formation comes from areas that can be shown to have experienced neither progressive subsidence nor uplift during the Quaternary. These coincide with the most ancient crust, the Archaean cratons that form the cores of the earliest continents, dating from the dawn of plate tectonics. Long established as ultrastable (Le Gallais and Lavoi, 1982; Gale, 1992; Twidale, 1997; Westaway et al., 2003, 2009a; de Broekert and Sandiford, 2005; Wesselingh et al., 2010; Belton et al., 2004), the fluvial records from such areas demonstrate that the rivers draining them currently flow at levels relatively similar, with respect to the enveloping landscape, to that occupied in early Quaternary or even pre-Quaternary times. This was an observation that came from

IGCP 449 and was documented by Westaway et al. (2003), who cited and illustrated examples from peninsular India and the Kaapvaal Craton of South Africa (Fig. 4; see also below, section 5).

3. Global patterns: late Neogene transition from basin fill to terrace formation

Fluvial sediments representing the infill of large depocentres are considerably more widespread in the comparatively recent (Late Cenozoic but pre-Quaternary) geological record than are modern or geologically very recent (mid–late Quaternary) analogues. This is exemplified by the well-preserved sedimentary sequence from the Tagus in Portugal, recently researched in detail by contributors to the Fluvial Archives Group (Cunha et al., 2005, 2008; Martins et al., 2010a). In the Palaeogene and until the mid-Pliocene separate drainage basins existed either side of the Portuguese Central Mountain Range, the area to the east belonging to the internally draining Madrid Basin; this endoreic basin was progressively captured by drainage to the Atlantic, leading to the formation of substantial westward-flowing rivers, of which the Tagus is the largest (Cunha et al., 2005). Beneath the Lower Tagus the basin sequence fills a syncline formed since the Cretaceous, with mainly marine sedimentation until late in the Miocene, after which fluviatile gravels were deposited as the uppermost basin fill, representing the earliest proto-Tagus. In the Quaternary, the Tagus has incised into the basin-fill sediments to form a classic river-terrace staircase (Cunha et al., 2005, 2008; Martins et al., 2010a; Fig. 5), comparable with those in this and other rivers in Spain (Santisteban and Schulte, 2007).

The sequence of sedimentary and landscape evolution documented in the Tagus, essentially 'basin inversion', is similar to that recorded in many other regions. During IGCP 449 pre-Pleistocene fluvial depocentres were documented from the Czech Republic, Ukraine, Bulgaria and Greece, Turkey, Australia, Argentina, Brazil and Bolivia. Amongst these the record from Ukraine is, as noted above, particularly informative. Here are found, on the interfluve between the River Dniester and its eastern neighbour, the Bug, and beneath later (Pliocene–Early Pleistocene) terrace deposits, basin-fill sediments of the Upper Miocene Balta Series, recording the earliest form of the river systems flowing southwards to the Black Sea (Matoshko et al., 2004). These deposits, which overlie marine sediments of the Paratethys Sea and are in turn overlapped by Pontian (Messinian) deposits, represent numerous cycles of fluvial aggradation in superposition, accruing ~100 m total thickness (Fig. 2B). The disposition of the various well-dated sediments spanning the Miocene–Pleistocene in this area reveals that the Black Sea sedimentary basin, which is still a subsiding depocentre, was much larger prior to the Middle Pliocene, when the hinge between net subsidence and net uplift stabilized at around the modern-day (interglacial) coastline (Matoshko et al., 2009). Thus the Miocene–Pliocene stacked fluvial sequences from areas fringing the Black Sea were formed in this larger Black Sea (or 'Paratethys') Basin (Fig. 2A). The Late Pliocene–Quaternary is represented landward from the coast mainly by terrace deposits, whereas on the northern Black Sea continental shelf it is represented by superimposed unconformity-separated wedges displaying offlap and seaward thickening (Ryan et al., 2003; Matoshko et al., 2009).

Straddling the border between Bulgaria and Greece, the Mesta/Nestos and Strouma/Strymon river systems have records of Miocene–Pliocene fluvio-lacustrine basin filling that was mostly fault controlled. This gave way in the Pleistocene to increased uplift and alternating aggradation and incision, which produced river terraces, particularly after the Mid-Pleistocene Revolution (MPR), ~0.9 million years ago, which saw the change from ~40 ka to 100 ka climatic cycles (Fig. 6). Earlier local literature typically sought explanation for the terraces in the intermittent movement of active faults but, In an IGCP 449 review, Zagorchev (2007) suggested (following Westaway, 2006b, who reviewed the earlier literature) that terrace formation was climatically triggered, as in other regions. In western Turkey Westaway et al. (2004, 2006b) and Maddy et al. (2005, 2007, 2008, 2012) have studied the terraces of the River Gediz system, noting the precursor basin-fill deposits into which these terraces are incised, the predominantly fluvial Hacıbekir and İnay groups, together exceeding 300 m in thickness (Seyitoğlu, 1997; Purvis and Robertson, 2004, 2005; Ersoy et al., 2010). Interfluve plateaux are capped by Lower Pleistocene basalts, which have protected these relatively unconsolidated sediments from erosion (along with early Gediz terrace gravels that separate the basin-fill sediments and lavas (Fig. 7). The basin-fill phase culminated in lacustrine deposition before fluvial incision and inversion began, the timing of the latter being somewhat difficult to determine because of erosion of the uppermost parts of the infill sequence (Maddy et al., 2012).

From the Czech Republic comes a record of a ~40 m thick valley fill, of suggested Miocene–Pliocene age, that occupies a high-level terrace position with respect to the Vltava system, although its affinity to that river is uncertain (Bridgland and Westaway, 2008b; cf. Tyráček et al., 2004). Straddling the northern edge of Prague and 1–4 km to the east of the Vltava, these deposits extend from ~106 to 149 m above the modern floodplain. Since the Pliocene, the river has incised below the level of this thick sequence, establishing a well-developed terrace sequence (Záruba et al., 1977; Tyráček et al., 2004). The pre-incision sediment-accumulation phase is thus a modest example of basin filling and the switch to down-cutting and terrace formation can be regarded as basin inversion, a phenomenon that has been seen to have occurred in many other regions at much the same time (Miocene–Early Quaternary), although its timing cannot always be determined with precision. In the Ukrainian example, where a depocentre still exists, the inversion was apparently time-transgressive.

Turning to the Southern Hemisphere, there are comparable examples from Australia and South America. The Murray–Darling river system of SE Australia is the largest on that dry continent and can be shown to have existed since the Palaeocene; for much of its history it was part of a subsiding depocentre in which several hundreds of metres of fluvial, lacustrine and marine sediments accumulated (Stevenson and Brown, 1989). As in other parts of the world, in the Late Cenozoic the subsidence changed to uplift (basin inversion again: see section 8), with incision into the basin-fill sequence and the formation of both fluvial and marine terraces. On a smaller scale, the River Shoalhaven system, draining to the east coast of Australia ~150 km south of Sydney, reveals evidence for Tertiary valley fill (Nott, 1992; Fig. 8A, B) and subsequent Pleistocene landscape inversion and terrace formation (Nott et al., 2002; Fig. 8C), attributable to the same post-Miocene and accelerating Middle–Late Pleistocene uplift as seen in other parts of the globe (e.g., Bridgland and Westaway, 2008a).

Northern South America is dominated by the Amazon Craton, which would appear, like those elsewhere (see above), to have experienced minimal uplift in the latest Cenozoic. The fluvial evidence for this is, as ever, negative, coming from an absence of river terraces. Nonetheless, on the north side of the Amazon craton (Suriname), staircases of laterized pediments have been formed by the progressive deepening of south–north flowing rivers, as evidenced by Van der Hammen and Wijmstra (1964) and Krook (1975). Immediately west of the craton, around Rio Branco, western Brazil, tributaries of the Amazon such as the Acre and Purus have formed terrace sequences that extend up to 70 m above modern floodplain level (Latrubesse et al., 1997; Westaway, 2006a). These terraces are inset into an older stacked fluvial/lacustrine succession, classified as the Solimões Group and representing depocentre filling that culminated at ~3 Ma (from biostratigraphy and interbedded tuff dated by Ar–Ar; Westaway, 2006a). Once again, it would seem that basin inversion has occurred in the latest Tertiary, perhaps in conjunction with Late Cenozoic global cooling and the onset of glaciation (cf. Westaway, 2001, 2002a; Bridgland and Westaway, 2008a, b; Westaway et al., 2009a). Something similar has occurred further south, in the Eastern Cordillera of the Andes, in central–western Argentina, where the River Mendoza, a tributary of the (Argentinian) Colorado, has incised a sequence of at least six terraces into a stacked accumulation of fluvial conglomerate, the Mogotes Formation, of inferred Pliocene age and extending to 2500 m above sea level, or ~1200 m above present river level (Brunotte, 1983).

It is worth noting that the recent geological history of SE England is essentially similar to the records described above. Here the Pleistocene terraces of the Thames and its tributaries overlie the Palaeogene fill of the London Basin, including possible Thames-precursor fluvial sediments within the Reading Beds on the Chiltern dipslope in Buckinghamshire (cf. Bridgland, 1994, p. 83; Bridgland et al., 2014b); however, the substantial gap in the record between the Eocene and the latest Pliocene prevents the basin-fill and terrace sequences being as satisfactorily linked as they are in the examples described above. This makes it difficult to be confident in straightforward evolution from the Early Cenozoic ‘depobasins’ reconstructed by Gibbard and Lewin (2003) to the present British drainage systems, as their interpretation implies. That interpretation requires the landscape of Britain to have been essentially unchanging over tens of millions of years, if not longer (e.g., Murray, 1992). The recognition of >100 m of Quaternary uplift, based on studies of river terraces (as described in this paper), from the dating of karstic systems in progressively deepening valleys (e.g., Westaway, 2010; Bridgland et al., 2014a), as well as the indication from thermochronology that there has been many hundreds of metres of Cenozoic denudation across much of Britain (e.g., Green, 2002; Green et al., 2012), have largely superseded this type of interpretation.

Even in the most rapidly uplifting crustal provinces at the present day there is evidence of inversion from a basin/valley-fill situation in the pre-Quaternary, followed by the formation of a river-terrace staircase. Thus in the Tibetan reach of the Yarlung Zangpo (uppermost Brahmaputra), Zhu et al. (2014) have described high-level fluvial deposits, forming what they term the highest river terrace, with typical thickness of ~200–350 m. These valley-fill deposits, ~550 m above valley-floor level, represent an ancient high-level terrace intermediate in age between Neogene deposits, which form the Dazhuka Formation and also appear to be restricted to the Yarlung Zangpo valley, and a system of numbered Quaternary terraces representing more conventional incision–aggradation cycles, albeit disrupted by glaciation and the formation of ice- and moraine-dammed lakes. Dated Oligocene–Miocene, the Dazhuka Formation consists of sandstones, conglomerates and volcano-clastic rocks up to 1200 m thick; although it extends along the valley for some 1500 km its interpretation as early Yarlung Zangpo sediment is equivocal (Zhu et al., 2014). Nonetheless, along with the thick high-level terrace deposits, it can be argued to provide evidence of pre-Quaternary valley filling in an area of strong Quaternary uplift, implying that even in the Himalayan Massif terrace formation was preceded by late Neogene ’basin’ inversion.

4. Global patterns: climatically-forced terraces showing acceleration of uplift and increased valley incision in response to greater climatic severity

The IGCP 449 and 518 fluvial archives dataset showed a number of significant patterns of valley incision, as determined from river-terrace preservation, revealing both global similarities and important regional differences. Incision is also implicit in the formation of river gorges, although these cannot readily be dated; terraces are of particular importance, as their sediments can provide a means for dating the incision between the different valley-floor levels thus represented, allowing incision rates and any fluctuation in these to be calculated. Even where no means of numerical dating is available, age models can often be provided for terrace sequences with reference to the fluctuation of Quaternary climate that is recorded in the fluvial sediments. Climatic fluctuation as a driver for terrace formation is an idea that has been promoted since multiple Pleistocene glacials and interglacials were first established (e.g., Zeuner, 1945; Bourdier, 1968; Wymer, 1968), although it fell out of favour during the period when terrestrial sequences were viewed in terms of over-simplified climato-stratigraphical models, in which just 6–7 climate cycles were recognized since the Pliocene (cf. Mitchell et al., 1973). The precise combination of forcing factors that has given rise to the predominant 100 ka climate cyclicity since the MPR is a topic for continued debate (e.g., Maslin and Ridgwell, 2005); nonetheless, with the recognition of nine 100 kyr cycles since ~0.9 Ma (see above; Fig. 6), it became possible to match river terraces to this climatic forcing (Kukla, 1975, 1977, 1978; Green and McGregor, 1980, 1987; Antoine, 1994; Bridgland, 1994, 2000, 2006, 2010; Bridgland and Allen, 1996; Maddy, 1997; Antoine et al., 2000, 2007). Quaternary climatic fluctuation has been inexorably linked with variations in sea level, which have long been regarded as a potential cause of river-terrace formation (e.g., Evans, 1971; Törnqvist and Blum, 1998; Martins et al., 2010a), although the modern-day consensus holds that the effect of climate on river systems is an effective driver irrespective of sea level, and is in any case (as noted above) required as an explanation for terrace formation in areas remote from the coast (Zeuner, 1945; Tyràček, 1983; Starkel, 2003). Indeed, as terraces occur with seemingly equal frequency in central continental areas, where sea-level control is improbable, mechanisms that can explain their formation in such areas are also likely to apply in coastal regions. Evidence that this is the case comes from the recognition that the aggradational braided-river gravels forming the bulk of most terrace sediment sequences, even those near to coasts, have generally been laid down during periods of cold climate (e.g. Rose and Allen, 1977; Green and McGregor, 1980; Gibbard, 1985; Vandenberghe, 1995, 2002; Macklin et al., 2002; Bridgland and Westaway, 2012), when sea level would have been low. At such times, if sea level were the main forcing factor, rivers should have been incising into their valley floors rather than aggrading. Where the continental shelf is wide, seismic profiling has typically demonstrated extensive offshore valley systems, with no marked break of slope at the modern coastline, suggesting (given knowledge of the recent sea-level rise by ~ 130 m from the Last Glacial lowstand) that sea-level fluctuation can readily be accommodated by course lengthening or shortening, with little imperative for aggradation or incision (cf. D'Olier, 1975; Bridgland, 1994, 2002; Bridgland and D'Olier 1989, 1995). During warmer (interglacial) episodes, rivers have typically adopted single-channel regimes, commensurate with incision, which is perhaps why former received wisdom held that incision had taken place during interglacials (Zeuner, 1945; cf. Vandenberghe, 2002), which would have been times of high sea level. This interpretation gave way to the empirical observation, from climatic evidence within fluvial sediment sequences, that valley deepening has predominantly occurred during periods of climatic transition (Vandenberghe, 1995, 2008; Bridgland, 2000; Maddy et al., 2000, 2001). Lewis et al. (2004) sought to clarify the situation, in part by recognizing an ephemeral 'coastal prism' in the lowest reach of the Thames valley, where they considered accretion in response to sea-level highstands to have taken place during interglacial optima, followed by degradation following climatic deterioration: effectively a reinvention of Zeuner's (1945) thallasostatic terraces, although accommodating the key point that the knick-point envisaged at river mouths is, as noted above, rarely observed (see below, section 8.3). Meanwhile the causal relation between sea-level fluctuation and river terraces has remained prominent in text books (e.g., Sparks, 1960; Holmes, 1965; Selby, 1985; Ballantyne and Harris, 1994) and continues to be taught to many students, despite that growing evidence that it is a rare mechanism confined to coastal reaches where the continental shelf is narrow.

With acceptance amongst the majority of the Quaternary fluvial community that climatic change has been a key driver in terrace formation, there has arisen a new debate over when, within the climatic cycle, the incision between distinct terrace levels has taken place. In Britain it has been suggested that down-cutting occurred primarily at glacial-to-interglacial transitions (Maddy, 1997; Maddy et al., 2001; Westaway et al., 2002), although review of the IGCP dataset implies that there are regional variations, with incision at cooling transitions perhaps the most common pattern on the European mainland, evident even in the nearby River Somme, in northern France (Antoine, 1994; Antoine et al., 2000, 2007; Vandenberghe, 2007, 2008; Fig. 1C). There are example sequences with fewer and others with more terraces than the documented number of 100 ka climate cycles with which to correlate them (Bridgland and Westaway, 2008a, b). Nonetheless, an approximate one-to-one match between terraces and glacial–interglacial (Milankovitch 100 kyr) cycles is commonplace. Where there are fewer, this is probably because the river has responded only to the more significant climatic cycles, perhaps those identified by Kukla (2005) as supercycles. Rivers with more terraces than 100 kyr cycles are rarer, although some have produced an extra terrace during MIS 7, which was characterized by warm episodes separated by a significant cold stage: MIS 7e and MIS 7c–7a, separated by relative cold during MIS 7d (Candy and Schreve, 2007). In most previous published interpretations however, the additional cold-climate forcing event has been attributed, probably erroneously, to MIS 7b; these include the Worcestershire–Warwickshire Avon (Maddy et al., 1991; Bridgland et al., 2004a) and, in northern France, the Somme (Antoine, 1994; Antoine et al., 2000) and the Yonne, a tributary of the Seine (Chaussé et al., 2004). More extreme is the record from the erstwhile River Solent, which would appear to have formed a pair of terraces during several of the late Middle Pleistocene 100 kyr climate cycles (Bridgland, 2001; Westaway et al., 2006a). Bridgland and Westaway (2008a) noted that all these examples are from uplifting crustal areas in proximity to the Atlantic margin, where enhanced sensitivity to climatic change might be an anticipated effect of the ocean circulatory system, suggesting that the latter was perhaps a factor that has led to the observed atypical responses.

It has long been recognized that well-separated aggradational river terraces are characteristic of the later parts of the Pleistocene, recording deeper valley incision in many parts of the world at that time, in contrast to the late Tertiary and

Early Pleistocene (Kukla, 1978; Maddy et al., 2000; Bridgland and Westaway, 2008a). Although uplift of an epeirogenic nature was central to early theories of landscape 'rejuvenation', in particular as part of the cycles of erosion theorized by W.M. Davis (1895, 1899), Van den Berg (1994) was perhaps the first to attribute the implicit change in landscape evolution to enhanced uplift rates, whereas Westaway (2001, 2002a) made the important suggestion that the acceleration of uplift was a response to increased climate severity, which he based on a correlation between its timing and established changes in the pattern of climatic fluctuation. The clearest of these correlations is the enhanced uplift that followed the MPR, which, with the change to 100 ka climate cycles (see above), saw an increased severity of glacials. An earlier (late Tertiary) comparable effect has already been mooted in explanation of the start of incision in western Brazil (see above). It can also be seen, and dated with more precision, within the record from the Maas, in the Netherlands, which shows an increase in uplift rate at around the end of the Mid-Pliocene, again coinciding with global cooling (Van den Berg, 1994; Van den Berg and Van Hoof, 2001; Westaway, 2001, 2002a; Westaway et al., 2009a; Fig. 1D). This post-Mid Pliocene phase of enhanced uplift is particularly clearly marked in records from the eastern USA, from the Ohio and Susquehanna Rivers (Westaway, 2007). It has also been recorded from the terrace record of the River Euphrates in southern Turkey (northern Arabian Platform), in a sequence that extends back to the Miocene (Demir et al., 2007a, 2008) and in the northern Black Sea rivers, in which (as in Brazil) it can be invoked as the driver for basin inversion (see above; Fig. 2B).

The data accumulated during the FLAG/IGCP 449 and 518 projects included numerous examples of well-dated terrace sequences that can be used to constrain the timing of the progressive valley incision and the causative uplift they record. Comparison of these data indeed shows that such uplift has proceeded at comparable rates in disparate parts of the world, wherever there is dynamic (non-cratonic) crust that is not loaded by widely accumulating sediment. Thus the uplift in the Murray–Darling since the basin inversion noted above is paralleled by uplift documented from the South Australia–Victoria border region, where it has resulted in marine (coastal) terraces. The implication is that there has been 60–110 m of uplift here since the beginning of the Middle Pleistocene (e.g., Huntley et al., 1993, 1994; Murray-Wallace et al., 1996), at a rate of ~0.07–0.13 mm a^{-1} (Bridgland and Westaway, 2008a), comparable with rates observed in NW Europe (cf. Maddy, 1997; Antoine et al., 2007) and with that calculated for the Vltava (Tyráček et al., 2004), in central Europe. The timing of the uplift in SE Australia is constrained by dated Quaternary basalt of the Mount Gambier/Mount Schank volcanic field (cf. Sheard, 1990) capping Middle Pleistocene marine terrace deposits that fringe the coast ~300 km SSE of the mouth of the Murray.

The majority of well-dated sequences, upon which calculations of uplift rates are based, are from the last 1 Ma and thus correspond with the period characterized by 100 ka climate cycles. Indeed, evidence of terraces from before the MPR, when these cycles began, is much rarer; pre-MPR terraces are often represented by sediment bodies that are likely to represent multiples of the earlier, shorter climate cycles, such as in the Thames (e.g., Maddy et al., 2000). Terrace archives with sufficient resolution to record the shorter, pre-MPR obliquity-driven climate cycles are rare indeed. One such is the record that represents the late Early Pleistocene River Gediz system, in western Turkey, preserved beneath plateaux-capping basaltic lava flows; here individual gravel terraces have been attributed to particular climate cycles between MIS 48 and 28 (Westaway et al., 2004, 2006b; Maddy et al., 2005, 2008, 2012; Fig. 7A).

The IGCP dataset has revealed numerous examples of records suggestive of an acceleration of uplift following the MPR, as summarized by Westaway et al. (2009a), who pointed to a range of case studies. These included the Dniester in the Ukraine (Matoshko et al., 2004, 2009; Fig. 2B), the Vltava and Dyje–Svratka in the Czech Republic (Tyráček et al., 2004; Tyráček and Havlíček, 2009) and the Maas, in the Netherlands (Van den Berg, 1994; Van den Berg and Van Hoof, 2001; Westaway, 2001, 2002a; Fig. 1D). This effect can also be seen in North American records such as those of the South Platte, in the Denver area, the Rio Grande and the Colorado upstream of the Grand Canyon (cf. Bridgland and Westaway, 2008a, b; Westaway et al., 2009a; Fig. 9; see below). Optimal preservation of uplift-generated river terraces occurs in the temperate regions, where glacial–interglacial climatic fluctuation has provided the triggering for terrace-forming processes. Indeed, it has already been noted above that terrace systems are particularly well developed, in terms of numbers of different levels, close to the margin of the climatically sensitive North Atlantic. Thus Phanerozoic crust in the tropics has also uplifted but, without the pronounced climatic fluctuation to trigger episodes of fluvial incision and aggradation, the terrace record in such locations is much sparser (Bridgland and Westaway, 2008a, 2012; cf. Büdel, 1977, 1982).

It is axiomatic that the same uplift that has driven terrace formation will also have forced the incision of gorge reaches through resistant bedrock. Here it is worth considering the recent research on the Colorado sequence in the SW USA (see above), in which emphasis has been given to explanations for the evolution of that river, and the cutting of the Grand Canyon, that call upon tectonic activity and/or other factors that would be unique to the geological history of that location (e.g., Levander et al., 2011; Karlstrom et al., 2012; Lee et al., 2013; Pederson et al., 2013). It is clear from the sedimentary part of the Colorado record (Fig. 9), however, that variations in rates of uplift and fluvial down-cutting can be observed, as elsewhere in the world, and that these can be correlated with the same perturbations of climate change (Bridgland and Westaway, 2008a; Westaway and Bridgland, 2014) so that, rather than being the result of unique circumstances, the formation of the Grand Canyon can evidently be explained in terms of the climatic forcing processes that have been identified from many other systems worldwide.

5. Local patterns: areas not showing the progressive uplift that typifies Pliocene–Quaternary landscape evolution

It has been established, from the data assembled during IGCP 449 and 518, that typical landscape evolution during the Pliocene–Quaternary has involved progressive uplift, with concomitant vertical fluvial incision, giving rise to flights of river terraces and/or (in areas of highly resistant bedrock) deep gorges. As identified already, there are exceptions to this pattern of evolution. The first is represented by depocentres, which are basins, typically tectonically generated, that have been progressively subsiding as a result, at least in part, of the positive-feedback effect of loading by the accumulating sediment. There is another exception, applying to significant areas worldwide: Archaean cratons and similar ultrastable areas, where net vertical movement of valley-floor levels during the course of the last few million years has been neither unidirectional nor by significant amounts (see above). This interpretation is often an inference from the absence of river-terrace staircases and, as a result, not entirely compelling. Critical, therefore, are examples that provide empirical evidence for long-term stability of valley-floor level in cratonic settings, such as the Vaal (Fig. 4). An even more compelling comparison can be made between the major north-shore Black Sea rivers, the Dniester and the Dnieper. The former, flowing southwards along the western edge of Ukraine, has already been seen to possess a well-formed and well-dated terrace staircase, extending back to basin inversion at the end of the Miocene (Matoshko et al., 2002, 2004; Fig. 2B & C). The Dnieper, ~300 km to the east, has a markedly different sedimentary sequence, despite also flowing southwards to the Black Sea; there are Dnieper sediment bodies corresponding in age to the various terraces of the Dniester, but these occupy positions in the landscape that range only between ~40 m below to ~50 m above the modern valley floor and show no clear relation between age and elevation. Indeed, some of the older Dnieper sediment bodies, such as the Lower Pliocene Parafiivka Series and the Upper Pliocene Chernobyl Series, are largely below modern river level (Fig. 2C). This is immediately reminiscent of the Vaal and, in common with that system, the bedrock here is again cratonic, being part of the Ukrainian Shield, although much of it is Early Proterozoic rather than Archaean. The Dniester valley, in contrast, lies to the west of this shield, on younger and more mobile crust of the Dniester–Bug crustal domain (Shchipansky and Bogdanova, 1996). In comparing the records from the neighbouring Dniester and Dnieper systems, both flowing southwards into the Black Sea and clearly within the same climatic zone, the only difference that can explain the marked contrast in the disposition of their fluvial records is crustal type and relative stability, and the effect this has had on uplift history (Westaway and Bridgland, 2014).

Two further examples from this general region of eastern Europe further underline the importance of crustal type in the evolution of landscapes and the development of topography, again using dated fluvial sequences to calibrate the evidence (cf. Bridgland and Westaway, 2008a, b). The first is the River Don, which flows from Russia southwards into the Sea of Azov. It has a combined stacked and terraced sequence that reveals a history of fluctuation between episodes of uplift and of subsidence that, despite not showing the ultra-stability of cratonic regions, has a similar effect in terms of net vertical migration of the valley-floor over the past ~15 Ma (Fig. 2D). Like much of the Dnieper, the Don valley is formed above Lower Proterozoic rocks, in this case of the Voronezh Shield (another part of the East European Platform). The variation in the fluvial records of these three Black Sea rivers, and their potential linkage to crustal characteristics, were discussed at length by Bridgland and Westaway (2008b), who emphasized that histories of uplift and incision from areas of Lower Proterozoic crust were often somewhat intermediate in character between the progressive and sustained movement seen on younger, hotter crustal types and the ultra-stability of the Archaean cratons. Indeed, Westaway (2012) suggested a possible explanation for the apparent alternation between uplift and subsidence in these regions in terms of crustal and lithospheric properties.

The remaining eastern European example is the sequence from the Lower Volga, in its approach to the Caspian Sea. Matoshko et al. (2004) published a transverse section across this part of the Volga that shows a superficial resemblance to that of the Dnieper, although the modern river is incised by only ~20–30 m into a stack of Middle and Upper Pleistocene fluvial sediments some ~100 m thick, with some evidence of repeated cut-and-fill events. This would appear to be a record of modest accumulation coupled with ultra-stability; there has clearly been little vertical migration of the valley floor in this system hereabouts. The Volga here is flowing across the 'Pre-Caspian Block', which has been interpreted as a fragment of oceanic crust that has been incorporated at the edge of the continent and covered in sediments (cf. Şengör et al., 1993; Nikishin et al., 1996). The high density of such oceanic crust would have precluded uplift. The absence of uplift of cratonic areas is not attributable to high density, however. On the contrary, cratons are generally formed of typically low-density continental crust that lacks the hot mobile lower crustal layer seen elsewhere on the continents. As argued previously by the present authors, flow within this lower crustal layer provides a highly plausible mechanism for driving progressive uplift, arguably as a coupled response to the increasing severity of climatic fluctuation during the past few million years, perhaps operating as positive feedback to isostasy in relation to redistribution of material by erosion in uplifting areas and sedimentation in adjacent depocentres (cf. Westaway, 2001, 2002b, c; Westaway et al., 2002). The plausibility of this mechanism is increased by the fact that it provides an explanation for the apparent coupling between changes in the style and severity of climatic fluctuation and increases in rates of uplift (e.g., Westaway, 2002a; Bridgland and Westaway, 2008a, b, 2012; Westaway et al., 2009a). This and other potential mechanisms for explaining the empirical records provided by fluvial sequences will be discussed in the synthesis section, below.

Another region with a fluvial record that points to reversals in vertical crustal motion during the Late Cenozoic, again seemingly related to crustal type, is the northern Arabian Platform, as represented by the River Euphrates in NE Syria (Fig. 10). The crust in this region is of Late Proterozoic age, having consolidated during the latest Precambrian 'Pan-African' orogeny but, like older Proterozoic crust elsewhere, it consists of a thick basal mafic layer overlain by a relatively thin layer of mobile felsic lower crust (cf. Demir et al., 2007b). Geochronological constraint has been provided recently by Ar–Ar dating of basalt flows that cap Euphrates terrace deposits between Raqqa and Deir ez-Zor (Demir et al., 2007a; Fig. 10). The resultant enhanced interpretation recognizes relative stability of the landscape here before ~3 Ma, followed by a phase of fluvial incision, then further relative stability before renewed incision, starting at ~2 Ma, which saw the river cut down to ~30 m below its present level. Aggradation of a 40–45 m thick deposit of gravel, which gives rise to Euphrates terrace QfII, took place in the late Early Pleistocene, after which renewed incision began, at around the start of the Middle Pleistocene, eventually reaching the present level of the river (Demir et al., 2007b). Reversals in vertical crustal motion are thus evident in the mid- and latest Early Pleistocene, as part of a more complex uplift history than was envisaged by Demir et al (2007a). Upstream of Raqqa, the Early Pleistocene incision did not reach below the present river level (Demir et al., 2007b); for example, at Birecik, southern Turkey (Fig. 10A), ~40 m of gravel between ~50 and ~50 m above river level represents the same episode of aggradation as the QfII deposit in Syria (Demir et al., 2008; compare Fig. 10A and B). Thus the same Early Pleistocene reversals in vertical crustal motion are evident upstream in Turkey, although greater subsequent uplift there means that the evidence is preserved higher within the landscape. This is consistent with the general southward tilt of the northern Arabian platform, indicative of a southward decrease in uplift (Arger et al., 2000; Demir et al., 2012).

At Diyarbakır, also near the northern margin of the Arabian Platform, terrace gravels of the River Tigris extend up to ~200 m above the modern valley floor; gravels ~70 m above the river are capped by distinct basalt flows, dated to ~1.20 and ~1.05 Ma (Bridgland et al., 2007a; Westaway et al., 2009b; Fig 10C), showing vertical crustal motion here to have been very low before increasing significantly at around the MPR. No Early Pleistocene reversal of vertical crustal motion, on the scale observed in the Euphrates, is evident at Diyarbakır, however, probably because the crust is somewhat hotter than further south in Syria (apparently with a slightly thicker mobile lower-crustal layer), possibly due, at least in part, to proximity to the much hotter crust of the Anatolian province (e.g., Tezcan, 1995).

Britain and NW Europe lack cratonic crust, but nonetheless there are areas of relative stability, signified by fluvial records that are demonstrably indicative of lower rates of vertical movement than elsewhere. A westward increase in crustal stability has been recognized in the British Isles, thought to relate to the westward constriction and eventual disappearance of the mobile lower-crustal layer that gives rise to crust of high stability in Ireland (e.g., Westaway, 2010; Green et al., 2012). However, as the whole of Ireland was glaciated during MIS 2 (e.g., Hiemstra, et al., 2006; Ó Cofaigh and Evans, 2007; Ó Cofaigh et al., 2008), there is no pre-Last Glacial Maximum (LGM) fluvial record and therefore no possibility of testing this suggestion using Late Cenozoic fluvial sequences. In central England, Quaternary fluvial deposits around Leicester identify a localized area of slow uplift: ~20 m in ~0.5 Ma, roughly half the amount evident ~35 km further north in the Nottingham area (Bridgland et al., 2014a), the latter being more typical of Britain as a whole. Indeed, Westaway (2011) identified a region of relative crustal stability in the southern part of the East Midlands, in the Milton Keynes–Northampton area, and this is probably a southern extension of the slowly uplifting crustal region seen at Leicester. In the Milton Keynes–Northampton area Lower Palaeozoic ‘basement’ is present in the shallow subsurface and has not been deeply buried by subsequent sedimentation, providing evidence of crustal stability extending back into deep geological time. In the Leicester area, Precambrian basement crops out in the Charnwood Forest inlier, conforming to a similar pattern and providing a further indication of crustal stability that has been a characteristic of this region over geological timescales and which has influenced both crustal structure and Quaternary landscape evolution. The evidence from this area includes the back-tilting of the lower Middle Pleistocene Bytham Sand and Gravel as it extends west–east beneath the modern valley of the River Wreake, a tributary of the Soar (Bridgland et al., 2014a; cf. Rice, 1991; Rose, 1994). Although deposited by an eastward-flowing river, the slow uplift of the Leicester area in comparison with that to the east of Melton Mowbray has resulted in this linear sediment body having a gentle east-to-west tilt, in apparent conflict with palaeocurrent and clast-provenance evidence, both pointing to transport from the west (Fig. 11).

6. Local patterns: areas showing unusually rapid uplift during the Middle – Late Pleistocene

The localities that have yielded IGCP 449 and 518 project data are invariably in regions with moderate to low uplift rates during the late Quaternary. This is somewhat counter-intuitive, given the widespread perception that the most significant research problems in geomorphology relate to largest-scale topography, which is related to the fastest uplift, in regions like the Tibetan Plateau. However, optimal long-term preservation of sedimentary evidence, including river-terrace deposits, is unlikely in areas that are uplifting extremely rapidly, if only because of concomitant rapid erosion (cf. Veldkamp and Van Dijke, 2000; Westaway et al., 2009a). Nonetheless, there are well-constrained fluvial sequences that establish localized areas of atypically fast uplift, in comparison with the established norm for post-Precambrian continental crust (see above) of ~0.07–0.13 mm a^{-1}. At the upper end of the range for Europe is the Middle Rhine, where the well-dated terrace sequence implies 200m of uplift since the late Early Pleistocene, at ~0.2mm a^{-1} (Westaway, 2001, 2002a). Similar rates have been calculated for the region around the NE corner of the Mediterranean, based on separate studies of terrace evidence from rivers flowing through Turkey and Syria: from NW to SE, these are the Ceyhan (Seyrek et al., 2008), Orontes (Bridgland et al., 2012) and Kebir (Bridgland et al., 2008), all of which record more rapid uplift than is typical (Fig. 12), resulting in sequences that do not extend back beyond the Middle Pleistocene. For example, in the Ceyhan, uplift rates of up to ~0.4 mm a^{-1} are evident from heights of fluvial terraces, the succession being constrained by Ar–Ar dating (~280 ka) of basalt capping a terrace assigned to MIS 10, into which younger terraces are inset (Seyrek et al., 2008). These rivers traverse the boundary zone between the Turkish, African and Arabian plates (e.g., Westaway, 2004; Duman and Emre, 2013) and so the local effects of active faults accommodating the plate motions are superimposed onto the more general effect of erosional isostasy (see below) in driving the uplift. Numerical modelling by Seyrek et al. (2008) suggested, however, that although the development of the topography in this region was initiated by the onset of the present phase of plate motions in the Mid-Pliocene, the resulting uplift has been driven primarily by erosion and is thus a consequence of the effect of climate on erosion rates, albeit with an initial tectonic trigger.

Rates of regional uplift of 0.2mm a^{-1} are by no means extreme in global terms. For the area around Auckland, North Island of New Zealand, Claessens et al. (2009) reconstructed post MPR uplift rates of ~0.4 mm a^{-1}, based on analysis of fluvial and marine datasets (although the latter provide the clearest evidence); the South Island has experienced even faster uplift, up to ~1 mm a^{-1}, determined from last-interglacial (MIS 5e) marine terraces (e.g., Kim and Sutherland, 2004; Cooper and Kostro, 2006). Because of this rapid uplift the longer-timescale record from the South Island is poor (cf. Westaway et al., 2009a).

The Grand Canyon of the Colorado River in the SW USA, perhaps the most famous fluvial landform in the world, is one for which rapid uplift is a prerequisite for its formation. Until recently, the chronology of the ~1500 m of fluvial entrenchment represented by this spectacular landform was unclear. Recent investigations, including thermochronology (e.g., Karlstrom et al., 2012) and the dating of speleothems that chart the water-table lowering in the surrounding strata (e.g., Karlstrom et al., 2008) now constrain this incision history. It is evident that post-

Pliocene incision rates have been low (~0.08 mm a^{-1}) in the western (upstream) part of the canyon, increasing to rather higher values (~0.2 mm a^{-1}) in its eastern part (Fig. 9A). Thus, much of the incision of the western Grand Canyon pre-dates the integration of drainage that formed the modern Colorado River at ~6 Ma. Conversely, most if not all of the entrenchment of the eastern part of the canyon post-dates the formation of the Colorado. Some 400 km upstream of the Grand Canyon, in the vicinity of Grand Junction in SW Colorado state, there has been ~1500 m of fluvial incision since a basalt eruption at ~10 Ma; furthermore, a terrace deposit ~100 m above the modern river contains tephra from the ~0.6 Ma Yellowstone eruption (e.g., Karlstrom et al., 2012). As a result, it has been argued (e.g., Karlstrom et al., 2012; Donahue et al., 2013) that rates of fluvial incision have remained constant, at ~0.15 mm a^{-1}, since ~10 Ma. Bridgland and Westaway (2008a) deduced variable uplift rates for this locality, indicative of phases of climatic forcing; however, it is now clear (Donahue et al., 2013; Westaway and Bridgland, 2014) that much of this variability relates to tributary deposits and can be ascribed to changes in the local drainage geometry due to tributary diversion or capture events. Nonetheless, at Grand Junction there are terraces 163–175, 80–100, 64–67, 24–37 and 3–5 m above the modern river, respectively assigned to MIS 22, 16, 12, 6 and 2 (Scott et al., 2002; Westaway and Bridgland, 2014). Thus the rate of fluvial incision, which can be taken as a proxy for uplift, was rather higher in the early Middle Pleistocene than subsequently, behaviour that is to be expected if the uplift is a consequence of erosional isostasy, with erosion rates increasing in response to the MPR (e.g., Westaway, 2002c; see above).

Most recently, Pederson et al. (2013) have proposed that the rate of fluvial incision increases upstream in the uppermost Grand Canyon to ~0.4 mm a^{-1}, based on OSL and cosmogenic dating of terrace deposits, notably around Lee’s Ferry, Arizona (Fig. 9C). The highest of these, some 200 m above the modern river, probably date from MIS 12 or thereabouts. Pederson et al. (2013) also reported similar rates of incision/uplift at sites in SE Utah, upstream of Lee’s Ferry, before the uplift rates taper further upstream to the aforementioned lower values calculated at Grand Junction (Fig. 9A). Their deduction, that the rapid uplift of this region is essentially the isostatic response to widespread erosion of unlithified Mesozoic and

Cenozoic sediments, supports the conclusions reached previously by Bridgland and Westaway (2008a) and Westaway et al. (2009a). The common preservation of post-MPR terrace deposits in this region, despite the rapid uplift, is perhaps a consequence of the arid climate.

In the upper Colorado, in the Glenwood Canyon area of the western Rocky Mountains, ~200 km upstream of Grand Junction (Fig. 9A), the pattern of fluvial incision is different again. Here, heights of dated basalt flows indicate an increase in time-averaged incision rates from ~0.02 mm a^{-1} during ~7.8–3.0 Ma to ~0.24 mm a^{-1}, time-averaged since ~3.0 Ma (Kunk et al., 2002). Using speleothem data, Polyak et al. (2013) resolved the younger part of this incision history into a phase at ~0.3 mm a^{-1} between ~3 and ~0.9 Ma, decreasing to ~0.15 mm a^{-1} since 0.9 Ma. They attributed this decrease in uplift rate, evidently coincident with the MPR, to decreasing erosion in response to increased local aridity, thus demonstrating a counter example in relation to the usually evident trend in temperate latitudes for increased uplift in response to enhanced cold-climate erosion, while nonetheless indicating an influence of climate change on uplift rates. The present geometry of the Colorado River has existed since ~6 Ma, since when, according to data from thermochronology, there has been cooling by ~45 ^{o}C of the rocks at present river level at Lee's Ferry (from ~60 ^{o}C to the modern-day ~15 ^{o}C; Lee et al., 2013). Given the present-day ~25 ^{o}C km^{-1} geothermal gradient, this equates to ~1.8 km of denudation at a time-averaged rate of ~0.3 mm a^{-1}. As noted above, the post-early Middle Pleistocene uplift rate in this locality has been ~0.4 mm a^{-1}, somewhat higher than the ~0.3 mm a^{-1} rate time-averaged since ~6 Ma. The difference (+ ~0.1 mm a^{-1}); cf. Pederson et al., 2013) provides further evidence that uplift rates have varied over time, as a result, the present authors would suggest, of climatic forcing of erosion rates.

Another case-study of a rapidly uplifting region was provided by Demir et al. (2009), who studied the terraces of the River Murat, a major tributary of the Euphrates in eastern Anatolia. As in the Colorado, river terraces formed in an Early Pleistocene forebear of the Murat valley are preserved high on the side of the modern valley (now inundated by a reservoir) thanks to burial beneath erosion-resistant basalt. This Çakmaközü Basalt, which has been dated to 1818 ± 39 ka (unspiked K–Ar: Demir et al., 2009), cascaded over four terraces at ~ 20 m intervals before covering the palaeo-Murat valley. The disposition of these terraces and of the dated basalt, as well as Mid-Pliocene lake sediments ~500 m higher in the landscape, together indicate an uplift rate of ~0.5 mm a^{-1} during the Late Pliocene and Early Pleistocene. As elsewhere, this uplift is interpreted as an isostatic response to erosion, indirectly driven by climate change, and not related to tectonic activity, despite the location close to the East Anatolian Fault Zone (Demir et al., 2009; Westaway et al., 2009a).

Perhaps the most rapid uplift thus far demonstrated from fluvial archives is in the Middle Yangtze in Yunnan, SW China, which was visited in 2006 by an IGCP 518 field excursion). Westaway (2009a) used the sporadically preserved fluvial terraces in this predominantly gorge reach of the Yangtze to estimate a late Quaternary uplift rate of ~0.8 mm a^{-1}, roughly double the average uplift in this region (calculated from a range of evidence, including thermochronology). Westaway interpreted the implicit acceleration of uplift as an isostatic response to the enhanced erosion resulting from the East Asian Monsoon, leading, by way of positive feedback, to crustal thickening from inflowing mobile lower crustal material.

7. The short-timescale records from areas glaciated during MIS 2

This paper has thus far been concerned with longer-timescale Quaternary sequences from regions generally beyond the influence of the Pleistocene ice sheets. As already noted in the case of Ireland, there are widespread parts of the continents, particularly in the Northern Hemisphere, where glaciation during MIS 2 has removed any evidence of earlier Quaternary fluvial archives. A notable exception, in limestone areas, is the survival of karstic evidence of underground drainage, which can provide well-dated constraints on valley incision and causative uplift, despite the valleys themselves having been glaciated and ‘wiped clean’ (by glacial erosion) of earlier (pre-MIS 2) fluvial archives (see Westaway, 2009b). A notable feature of the British landscape within the MIS 2 glacial limit is that it is superficially similar, in terms of the incised nature of its river valleys and the occurrence of terraces on their flanks, to the area beyond (generally to the south) of this limit. Factored into this comparison, naturally, must be a consideration of the more upland character of the northern regions that were glaciated during MIS 2, which is partly a result of their geology. It may also owe much, however, to the repeated glaciation experienced and the degree to which the erosion thus generated has driven isostatic compensation of the sort invoked to explain the longer-timescale uplift of unglaciated regions, potentially at a faster rate because of the accelerated erosion likely to have resulted from glacial processes. As in the unglaciated regions, this isostatic mechanism will presumably have brought about permanent effects only in lithospheric provinces that are post-cratonic, having hot and mobile lower crustal layers that can respond the loading and unloading effect. In cratonic regions, which include the two great Northern Hemisphere ice-gathering centres, Fennoscandia and Laurentia, the isostatic compensation will have taken place entirely in the mantle, as modelled in respect of glacio-isostasy by Lambeck (1995) and Peltier (2002), in which case the effect is unlikely to have been permanent, with unloading having led to complete recovery of previously depressed areas (cf. Bridgland et al., 2010).

Areas that were glaciated during MIS 2 and which have comparable crustal characteristics to Britain are scarce; these need to be not just young, dynamic continental crust but also distant from plate boundaries and active faults, where tectonic uplift could be invoked in explanation of fluvial incision. The high plains of Canada, which show clear evidence of post-glacial fluvial incision in the form of canyons and terrace systems (Jackson et al., 1982; Rains and Welch, 1988; Rains et al., 1994; Evans et al., 2004), are on post-cratonic Precambrian crust, which explains the occurrence of such evidence, which has sometimes been attributed to glacio-isostatic effects (Bryan et al., 1987; Campbell, 1997; Oetelaar, 2002), but disqualifies the area as a direct analogue for Britain. A similar observation can be made with regard to Michigan, where post-glacial incision below terraces representing MIS 2 deglaciation have been observed on Proterozoic crust flanking the Laurentian craton (Arbogast et al., 2008). On the eastern side of North America, however, the area of the Appalachian Mountains is not just an analogue for Britain's Phanerozoic continental crust: before early Mesozoic Atlantic rifting these two areas were contiguous. In the northern Appalachians, terraces of the upper reaches and headwaters of the Ohio River system show deformation that is attributed to deposition during periods of crustal depression resulting from Laurentide ice sheets, this area being at the periphery of these repeated Middle and Late Pleistocene glaciations, which have repeatedly diverted the Ohio headwaters (Jacobson et al., 1988; Westaway, 2007). This would indeed appear to be a good analogue for the post-glacial glacio-isostatic effect recognized in northern Britain, albeit representative of more than the last glaciation.

Modelling of British Late Pleistocene glacio-isostasy has hitherto been based on the same mantle-compensation of crustal loading and unloading by ice sheets as in the above-mentioned cratonic regions (e.g., Peltier et al., 2002; Shennan et al., 2006) and has predicted minimal post-MIS 2 isostatic rebound in northern England, seemingly precluding uplift thus generated as a cause of the post-glacial valley incision that is a characteristic feature in that region. This incision is manifest in the profusion of gorges and terrace reaches to be seen in eastern Pennine rivers such as the Ure, Swale, Wear and Tyne and in the deeply incised minor valleys known as 'denes' that drain to the North Sea coast, such as Castle Eden Dene and Hawthorn Dene (Beaumont, 1970; Yorke, 2008; Bridgland et al., 2010, 2011). Bridgland et al. (2010, 2011) noted a similarity between all these systems in that the glaciated landscape into which they are incised is typically ~30 m above the modern valley floors, with terraces preserved sporadically on the valley sides that, where datable, range in age from Lateglacial to Late Holocene (even post-Medieval sediments can occupy low-level terrace situations in such valleys). It was also apparent that earlier versions of many of these incised valleys existed, generally filled with MIS 2 glacial sediments but often re-excavated by the post-glacial incision. This suggests that the equilibrium position of the pre-glacial valley floors, in terms of position within the landscape, was the level represented by the bases of the buried valleys, the infill (and the accumulation of sediments represented by the 'glaciated plateau' into which renewed incision has occurred) having taken place during glacio-isostatic depression. The rivers have largely succeeded in returning to these supposed equilibrium levels since deglaciation, usually by cutting though the unconsolidated valley-fill deposits but sometimes, as with the famous Durham Meander of the River Wear, departing from the pre-glacial course and cutting a new valley in bedrock.

Bridgland et al. (2010, 2011) concluded that glacio-isostatic rebound (uplift) would have been an essential driver for this renewed post-glacial incision, regarding this as an isostatic mechanism largely driven by lower-crustal mobility and, therefore, peculiar to the younger and more dynamic crust of areas like northern Britain. This type of isostatic compensation is likely to have taken place rapidly after deglaciation and been localized within areas formerly beneath ice sheets (Bridgland et al., 2010). A test of this idea, utilized by Bridgland et al. (2010), is that the viscosity distribution required in the lower crust beneath northern England to account for this component of glacio-isostasy is consistent with that deduced from modelling of longer-timescale vertical crustal motions in the same region (Westaway, 2009b).

Bridgland et al. (2010, 2011) were reporting on detailed work undertaken in the River Ure system of North Yorkshire, which, thanks to the dating constraints now available, provides an intriguing comparison with the larger eastward-draining River Thames further south (see Fig. 13). Similar records of post-LGM incision have been reported from the Wharfe (Howard et al., 2000b) and, on the western side of the Pennines, from the Mosedale Beck, NE Lake District (Boardman, 1994, 1997, 2002) and from the River Ribble, in Lancashire (Chiverrell et al., 2007, 2009). Comparable post-LGM terrace sequences have been reported from SW Scotland , in the valley of the Kirtle Water, by Tipping (1995, 1999) and from SW Ireland by Harrison et al. (2002), who showed that the valley of the River Gaddagh, which drains from the Macgillycuddy's Reeks into the Atlantic, has been incised by ~10 m into the glacial sediments of the region, with five cut-and-fill terraces inset into the fill of an early post-glacial valley incision to below modern floodplain level. Bridgland et al. (2010) proposed that the ~30 m of postglacial fluvial entrenchment apparent in NE England was a consequence of a component of ~30 m of localized uplift as a result of inward flow of lower-crustal material in response to unloading of the Earth's surface due to the unloading accompanying deglaciation, this component being in addition to the predominant glacio-isostatic response that occurred in the mantle. The reduction in this effect from ~30 m in NE England to ~10 m in SW Ireland is consistent with a westward reduction in the mobility of the mobile lower-crustal (or mid-crustal) layer associated with a westward increase in the thickness of the layer of mafic underplating at the base of the crust (e.g., Westaway, 2010; Green et al., 2012). SW

Ireland is known to be underlain by a thick layer of underplating, its top at a minimum depth of ~22–23 km (e.g., Masson et al., 1998). Preliminary geothermal calculations by Westaway (2010) indicate that, at most, only a thin layer of mobile lower crust can be expected here, although estimation of its thickness on this basis is difficult due to uncertainties in geothermal data (Westaway and Bridgland, 2012). It is not fully established how much of the layer of underplating, the thickness of which generally increases westward beneath Britain and Ireland (e.g., Westaway, 2010; Green et al., 2012), was emplaced as a result of the Palaeogene British Tertiary Igneous Province (BTIP) magmatism and how much relates to earlier magmatic episodes. The origin of this layer beneath SW Ireland has been discussed, for example, by Klemperer et al. (1991) and Masson et al. (1998); the BTIP magmatism is known to have been broadly synchronous with significant vertical crustal motions and changes to sedimentary environments in localities now offshore of SW Ireland, attributed to thermal effects of the Iceland mantle plume (e.g., Jones et al., 2001; cf. McDonnell and Shannon, 2001), so additions to the thickness of magmatic underplating might well be expected at this time.

Thus, as Bridgland et al. (2010, 2011) have pointed out, the fluvial records inside and outside the MIS 2 ice limit can be contrasted, as is exemplified in eastern England by drainage to the North Sea. In NE England, within the MIS 2 ice limit, Lateglacial–Holocene deposits form terrace sequences, with glacial outwash gravels typically forming the highest terrace, up to 30 m above river level. Lateglacial and early Holocene deposits are preserved as intermediate terraces but no older, 'pre-glacial' terraces survive, the MIS 2 glaciation having destroyed any that once existed. Beyond the MIS 2 ice limit, latest Devensian and Holocene fluvial deposits are restricted to the floodplain and any buried channel deposits that underlie the floodplain, although sometimes the latter continue above modern river level to form the lowest terrace; older terraces, dating back to the Middle Pleistocene and earlier, will typically form the majority of the record in these areas (Bridgland et al., 2010; see examples in Fig. 1). As noted already, and illustrated in Fig. 13, there is a degree of similarity in the geomorphological character of the valleys in locations outside and inside the Last Glacial limit; valley incision and aggradational terraces are common to both, and of comparable scale (in terms of heights above valley floor: Fig. 13), with the essential difference being in the age of the deposits in question. Beyond the MIS 2 ice limit, it is clear that the terraces have formed in relation to climatic triggering at Milankovitch timescales and that the uplift recorded is generally in the order of 0.04–0.1 mm a^{-1}., whereas minimum rates of uplift indicated by the post-glacial terraces in NE England suggest uplift of ~30 m during ~ 15 ka, or 2 mm a^{-1}. The general similarity of the landscapes and relief in, for example, SE and NE England probably delayed recognition by early Earth scientists of the very different ages of the glacial deposits in these regions and of their greatly different Quaternary history; these similar landscapes could be readily reconciled with 'monoglacial' theory.

8. Discussion: implications of fluvial archives for an understanding of landscape evolution and the mechanisms that have driven it

An important implication of the patterns detected amongst Late Cenozoic fluvial records as a result of the above-mentioned IGCP projects is that crustal type, i.e., cratonic, Early Proterozoic, dynamic (post-Precambrian) or highly dynamic and relatively hot, has a very large and hitherto largely overlooked influence on landscape evolution. Crustal properties are therefore potentially implicated in the causation of differences between regions that have hitherto been attributed to other factors, such as proximity to tectonic plate boundaries or to active fault zones, or characteristics particular to different climatic regions. In the last case, for example, Büdel (1977, 1982) developed a view, implicit from his theories of 'climatic geomorphology', that river terraces did not occur in the tropical zone. Data from IGCP 449 have, however, contributed to the falsification of any such hypothesis, showing instead that terrace sequences do indeed occur in tropical regions. Thus Veldkamp et al. (2007) reported on a long-timescale terrace system of the River Tana in Kenya, dated with reference to Quaternary volcanic activity in the catchment. The River Niger in eastern Mali and southern Niger (Beaudet et al., 1981; Bergoeing and Gilliard, 1997) and the Acre and Purus (Amazon tributaries), mentioned above, are further examples of tropical rivers with terraces. A long-timescale river-terrace staircase has also been reported from the Awash in Ethiopia, where the component deposits are important sources of Lower Palaeolithic artefacts, as with many river terraces in the temperate latitudes of Europe. As with several other examples discussed above, volcanic activity has both helped to preserve the Awash sequence and provided means for dating it (e.g., Chavaillon et al., 1979; Gallotti et al., 2010). Bridgland and Westaway (2008a) concluded that Büdel, who worked extensively in Amazonia and central Africa, was misled by the stability of these areas, which results from their cratonic crust (in the latter case the Archaean Congo Craton) and has prevented river terrace formation.

The recognition of meaningful patterns of variability between Late Cenozoic fluvial sequences in different regions, or crustal provinces, has repercussions for determining the most plausible mechanisms at work. The present authors have previously noted the importance, as a key factor governing this variability, of the thickness of the mobile lower crustal layer (e.g., Westaway et al., 2003, 2009a; Bridgland and Westaway, 2008b; Westaway and Bridgland, 2014). The isostatic response to inflow, beneath an area of crust, of material in this mobile layer, driven by lateral pressure gradients created by erosion (and, in many cases, by sediment loading elsewhere), is envisaged as a general mechanism that can explain the uplift of the area concerned (e.g., Westaway, 2001, 2002a, b, c); indeed, this can provide an explanation in each of the localities thus far reviewed. An important point is that this mechanism can enable, by means of positive feedback, uplift at rates that exceed the spatially-averaged erosion rate of the uplifting region, as has been demonstrated for some of the study regions discussed (e.g., Westaway et al., 2006a; Westaway, 2009a), notwithstanding the difficulties that often arise over precise estimation of amounts, timings and rates of erosion (cf. Maddy et al., 2012). An important influence is the loading effect of sedimentation in depocentres driving outward flow of mobile lower crust from beneath such areas and beneath uplifted regions in the vicinity (cf. Westaway, 2002c). By the equivalent opposite process, erosion of uplifting areas can induce inward lower-crustal flow, providing positive feedback that further sustains the uplift.

Global syntheses (e.g., Zhang et al., 2001) have established that rates of sedimentation in many depocentres worldwide increased in response to climate change around the start of the Pleistocene (i.e., at the end of the Mid-Pliocene

Climatic Optimum) and at times during the Quaternary, especially as regards offshore depocentres. In the light of experience of modelling such effects (e.g., Westaway, 2002c), it can be envisaged that the isostatic response to such increases, mediated by lower-crustal flow, will have resulted in faster uplift of eroding onshore regions that act as sediment sources, potentially leading to the late Neogene inversion, or switch from sedimentation to erosion, widely observed in many former smaller onshore depocentres, located in areas that began to experience more general uplift in response to accelerated surface processes brought about by Pleistocene cooling (see above).

The fact that changing patterns of river-terrace formation can be interpreted as matching changes in patterns of Late Cenozoic and Quaternary climatic fluctuation (see above), presumably via the broad effects of climate on sediment mobility, supports the deduction that climatic forcing is the mechanism responsible for driving the observed uplift. This deduction is strengthened if the onshore uplift is paired with corresponding subsidence in adjacent depocentres, such correlations being indicative of surface processes coupled by induced lower-crustal flow being the causative mechanism (e.g., Westaway, 2002c). The aforementioned River Dniester (Fig. 2a) provides a particularly outstanding example of a fluvial terrace staircase demonstrating such phases of synchrony, associated with faster erosion during the Pontian salinity crisis of the Black Sea basin, at the end of the Mid-Pliocene Climatic Optimum, around the climate deterioration that occurred at ~2 Ma, and following the MPR (e.g., Bridgland and Westaway, 2008b; Westaway et al., 2009; Westaway and Bridgland, 2014). Other examples of similar effects recently recognized include, first, the rapid uplift of the eastern Anatolian Plateau following the Mid-Pliocene Climatic Optimum, which Demir et al. (2009) envisaged to be sustained by the inflow of lower crust from beneath the adjacent Black Sea depocentre. A second example is the pairing of the uplift of northern England and Scandinavia, again since the Mid-Pliocene Climatic Optimum, with subsidence of the North Sea basin, recognized by Westaway (2009b). In eastern England, this gradual transition from onshore uplift to offshore subsidence is reflected in the downstream convergence of the terraces of the early Middle Pleistocene Bytham River (Westaway, 2011). A third example is the rapid uplift of southern Italy since the

MPR, paired to synchronous faster subsidence in its offshore surroundings (Westaway and Bridgland, 2007). This latter region, of course, is located within a plate boundary zone but in the view of Westaway and Bridgland (2007) the effects of the active faulting that accommodates the plate motions are superimposed onto a background of regional-scale vertical crustal motions caused by the isostatic response to climate change.

As already discussed, the Colorado dataset provides evidence of uplift rates varying over time in response to climatic forcing, which elsewhere in the world is indicative of isostatic compensation involving lower-crustal flow induced by surface processes. However, unlike the examples of paired uplift and subsidence noted above, in this particular locality it is not at all clear from where the lower-crustal material required by this general mechanism, to sustain the observed uplift, might have flowed. In principle, it might have originated from beneath endoreic depocentres in the Basin and Range Province to the west, or the northern Rio Grande Rift to the east (Fig. 9(a)), or possibly from beneath coastal regions flanking the Gulf of California or the Pacific Ocean to the southwest. No quantitative modelling of this effect is therefore possible at this stage; however, the evident mismatch between the zone of most rapid uplift revealed by the fluvial evidence and the sites where 'tectonic' forcing of this uplift have been proposed (Fig. 9(a)) raise doubts as to the validity of the 'tectonic' forcing mechanism (cf. Levander et al., 2011; Karlstrom et al., 2012). For this and for many other examples worldwide, the limited information currently available precludes any definitive conclusion being reached; hence the emphasis in the present study on localities for which the strength of the available evidence allows clear conclusions to be drawn.

The combination of thick lithosphere and low radioactive heat production in the upper crust means that a mobile lower-crustal layer is absent in Archaean cratons, which are ultrastable, as already noted. This correlation between ultra-stability and the absence of a mobile layer lends weight to the argument that the significant rates of vertical crustal motion observed in other crustal provinces are feasible as consequence of the presence there of this mobile layer (Westaway et al., 2003).

In Phanerozoic crustal provinces the mobile lower crustal layer may be ~10 km thick or more (e.g., Westaway, 2002a, b, c). On the other hand, this layer is typically thinner in Early/Middle Proterozoic crustal provinces, where its vertical extent is constricted by a mafic layer at the base of the crust (e.g., Westaway and Bridgland, 2014). Similar mafic layers, added to the base of the crust by magmatism associated with mantle plumes, may likewise constrict the thickness of the mobile lower crustal layer in younger crustal provinces. Westaway (2012) has suggested that the occurrence of reversals in vertical crustal motion, as are observed in the Dnieper and Don (see above; Fig. 2), results from the interaction between isostatic compensation of erosion by lower-crustal flow and by deformation within the mantle lithosphere. He suggested that if the mobile lower-crustal layer is thin (≤~6 km thick) these two isostatic responses will be sufficiently separated in time as to produce a reversal in the sense of vertical crustal motion. In the Early Proterozoic crust of Eastern Europe, the mobile lower-crustal layer is clearly thin, due to the low heat flow linked to high lithospheric thickness (cf. Westaway and Bridgland, 2014). As in other regions of Early Proterozoic basement, the basal crust of the East European Platform consists of a mafic layer which does not flow and helps to constrict the overlying layer of mobile, felsic, continental crust.

A significant discussion point here is the chain of cause and effect, specifically the temporal relation between forcing and response. In many well-documented cases, there is evidence for accelerated uplift following the MPR, suggesting that the greater climatic severity resulted in increased erosion rates, which have resulted in turn in increased uplift rates. This cause-and-effect sequence can be explained by isostatic modelling in which changes in erosion rates induce flow in the mobile lower-crustal layer, the time-lag between the increase in erosion rates and the onset of the uplift response depending on crustal properties, in particular the thickness of the mobile layer (e.g., Westaway, 2002a, b, 2007, 2012). The idea that rates of uplift and erosion are interrelated is well understood but the chain of cause and effect is not. For example, as part of a study of Cenozoic denudation in the British Isles, Jones et al. (2002) considered the rate of sediment flux into an offshore basin to be related to the size of the corresponding drainage catchment and the rate of denudation therein; they suggested (after Reading, 1991; Burgess and Hovius, 1998) that the lag time between an increase in denudation rate and the corresponding increase in offshore sediment flux would be < 100 ka, thus concluding that sediment-flux histories in the basins surrounding Britain and Ireland could be directly related to uplift of their sediment source regions. Quaternary fluvial datasets, by virtue of their much better time resolution compared with most other types of geological record, indicate that erosion forces uplift rather than the other way round, although positive feedback effects are clearly important. Such datasets therefore illustrate the general manner in which these processes interact in a manner that is important for many other aspects of Earth Science beyond the Quaternary.

As the Colorado example illustrates, many previous authors have sought explanations for well-developed river-terrace sequences in terms of plate tectonic activity or proximity to active fault systems, notwithstanding that such mechanisms would be problematic as explanations for what can now be confirmed as widespread phenomena. Indeed, it is apparent that typical rates of uplift in tectonically active regions, such as the Mediterranean, are often entirely comparable with those far from plate boundaries (Bridgland and Westaway, 2008a, b; 2012; Westaway and Bridgland, 2009). The effects of Quaternary tectonic activity can be detected in the former regions, where they appear as disruptive influences, perturbing the more systematic results of background epeirogenic uplift. Thus the Middle Euphrates in eastern Syria has terraces that are significantly deformed by tectonic movements of fault belts that were hitherto not known to have been active during the Quaternary (Abou Romieh et al., 2009). There are numerous other examples, from various parts of the world, of fluvial sequences affected by active faulting, many with longer research pedigrees. For example, Krzyszkowski et al. (1998, 2000) documented displacement of the terraces of the River Nysa Kłodzka as they pass across the Sudeten Boundary Fault, SW Poland, while Krzyszkowski and Biernat (1998) reported similar deformation of the terraces of the left bank tributary of the Nysa Kłodzka, the Bystrzyca, related to the same cause. Other examples of rivers affected by active faulting have been discussed by Bridgland and Westaway (2012). A particularly dramatic example is provided by the Yangtze in Yunnan, SW China, which has developed a pronounced knickpoint where it passes through Tiger-Leaping Gorge, a zone of localized rapid uplift associated with active faulting (Westaway, 2009a). Likewise, uplift rates within the Colorado catchment taper from the aforementioned ~0.4 mm a^{-1} value upstream of the Grand Canyon to much lower values further west, beyond the downstream end of the Grand Canyon, through a combination of tilting and active faulting (e.g., Bridgland and Westaway, 2008a; Karlstrom et al., 2008; Lee et al., 2013; Pederson et al., 2013).

8.1 Further application: Poland

Poland can be offered as a further case study area, not used to establish the varieties of preservation styles and processes, but one with variable styles of fluvial archive preservation that can be interpreted according to the suggested mechanisms and relationships proposed in this paper. The above Polish examples of tectonically deformed Quaternary river terraces are from the extreme SW of that country, although much of Poland lacks well-developed terrace records and is instead characterized by stacked sequences indicative of subsidence (cf. Marks, 2004), particularly in the vicinity of salt diapirism (such as at Bełchatow, near Łodz: Krzyszkowski, 1995; Krzyszkowski and Szuchnik, 1995), or more enigmatic fluvial archives suggestive of fluctuations between uplift and subsidence. Terraces are also found in SE Poland, near the border with Slovakia, where they are documented from the catchments of the Dunajec (Zuchiewicz 1992) and the San (Starkel, 2003), both tributaries of the Vistula. These records are from crust that was affected by the Caledonian orogeny and is bordering on the Western Carpathian Mountains, products of Cenozoic plate motions. It is thus conventional 'young crust' and progressive uplift during the Quaternary would be anticipated. Indeed, the headwaters of the San and Dniester (see above) are very close together, near the point where the borders of Poland, Ukraine and Slovakia meet, the two rivers flowing in opposite directions, and comparable records would thus be expected from these neighbouring systems. Further downstream in the Vistula system is crust that forms part of the East European Platform, which was seen above to be characterized by evidence for alternating pattern of uplift and subsidence brought about by low crustal heat flow and/or thick lithosphere (cf. Bridgland and Westaway, 2008b; Westaway and Bridgland, 2014). The Vistula has been much affected by glaciation and its lower catchment covered in glacigenic sediments, so the fluvial record is less well documented. Nonetheless, evidence for the type of record observed in the Don valley (Fig. 2d) and the Arabian Platform (Fig. 10) has been determined from subsurface data, which show terracing of Pliocene and Early Pleistocene valley floors, with multiple subsequent channelling and burial (Mojski, 1982). This suggests that, although the sedimentary stacking might well reflect proximity to the Baltic Basin, some of the characteristics of Central Poland that might traditionally have been accredited to the effects of glaciation, or glaciation interspersed with marine transgression (e.g., Marks, 2004) result instead from the characteristics of the crust. Like other northern European rivers, the Vistula has also experienced glacial diversion, its lower course reflecting the geometry of retreat of the Scandinavian Ice Sheet at the end of the LGM, as well as glacio-lacustrine influences (Kozarski, 1988; Marks, 2004). There are perhaps transitions within this system between three provinces: first an upstream, uplifting province, with well-developed terraces, then a central province in which the comparative stability of the East European Platform is important in determining the characteristics of landscape and fluvial archive disposition, giving way northwards to the increasing influence of the Baltic Basin and to the effects of repeated glaciation. Thus fluvial archives from Poland are readily explained within the framework established from elsewhere and can be reconciled with the mechanisms that have been proposed above for the translation of the various forcing factors and influences into different patterns of evolution and preservation.

Implicit in the use of heights of river terraces as a proxy for uplift is the notion that each fluvial terrace deposit was emplaced under an equivalent hydrological regime. This has been stated many times (e.g., Maddy, 1997; Westaway et al., 2002); moreover it is well established on theoretical grounds that for a given upstream catchment area the longitudinal gradient at which a river is in equilibrium depends on the hydrology. In detail, rivers in equilibrium are known to adopt long profiles along which the Shields stress parameter for the entrainment of bedload maintains a particular threshold value (e.g., Shields, 1936; Paola and Mohrig, 1996; Mueller et al., 2005). In cases where river terraces converge downstream and pass into stacked sequences, such as the Rhine in NW Germany (Fig. 3) or where the early Middle Pleistocene Bytham River approached the subsiding North Sea Basin, in coastal East Anglia (Westaway, 2009c), it is apparent that the observed convergence results from tapering between uplift and subsidence. However, in cases where river terraces converge downstream with the modern long profile of a river, the interpretation is potentially more problematic. A case in point is provided by the Platte River in the U.S. Midwest, which has well-developed terraces in its upper reaches but these converge downstream and grade to the modern river level. This effect has been interpreted by Bridgland and Westaway (2008) and Westaway and Bridgland (2014) as a consequence of uplift tapering downstream as the river passes into regions of progressively greater crustal stability. Duller et al. (2012), however, have attributed the effect to post-Pliocene changes in hydrology and thus in equilibrium longitudinal gradient.

8.2 Overview: application to Britain

Already encountered as the source of important examples of classic river-terrace staircases (Fig. 1A) and concerning the formation of post-glacial terraces (section 7), Britain has long been regarded as a crustally stable region, due to its minimal seismicity. However, the evidence from river terraces (as well as from raised beaches and from cave levels) is clearly indicative of significant Quaternary uplift across much of the British land mass. The regions where the highest uplift rates have been demonstrated, which include NE England (~0.2 mm a^{-1}), the Yorkshire Dales and Peak District (~0.15 mm a^{-1}) and much of the Hampshire Basin (~0.1 mm a^{-1}), are also regions with relatively high heat flow and, therefore, relatively high temperatures at mid- and lower-crustal depths (e.g., Westaway et al., 2006a; Westaway, 2009b); significant mobility in the mobile lower-crustal layer can thus be envisaged. This correlation between uplift rates and heat flow, and the general observation that the resulting crustal deformation is largely aseismic, are consistent with the view that the deformation results from lower-crustal flow. Indeed, Westaway (2009b) envisaged that erosion onshore and sediment loading offshore have created a horizontal pressure gradient that acts to drive lower-crustal material from under offshore depocentres to beneath the land, the principal effect being that the many hundreds of metres of Quaternary sediment-loading in the southern North Sea are being accommodated by a component of westward lower-crustal flow to beneath the land area of Britain.

It is also apparent that there has been fault activity in Britain, in addition to the predominant epeirogenic uplift. This was first recognized from the warping of terraces of the drowned River Solent across the Portsdown Anticline by Westaway et al. (2006b). They observed an additional ~10 m of uplift (additional elevation) in respect of terraces in close proximity to the anticlinal axis, interpreted as resulting from vertical slip by that amount (during the past 1 Ma) on the blind reverse fault beneath this structure. Harding et al. (2012) further refined the modelled displacement of terraces by this fault, suggesting that since 0.9 Ma it has contributed an additional 26 m of uplift, occurring at a uniform rate, in respect of the River Test terraces at Chilworth, Hampshire; for the Solent River terraces at Porchester the additional uplift, during the same interval, has been 20 m. A further example of active fault movement within the supposedly stable British land area was recognized by Westaway (2010), from the anomalously rapid uplift of the Mendip Hills in Somerset, revealed by the disposition of cave levels: attributed to slip on an underlying blind reverse fault. The localized slip on these faults can be inferred to have occurred in order to accommodate changes in the state of stress in the adjoining crust that result from lateral variations in rates of surface processes or of the accompanying lower-crustal flow (cf. Westaway, 2006c). Following similar usage by others (e.g., Kaufman and Royden, 1994), this deformation mechanism should be termed ‘atectonic’, so as to distinguish it from the tectonic deformation that is caused by plate motions.

8.3 An alternative mechanism: knick-point recession

The relation between river-terrace sequences and landscape evolution has not been universally acknowledged thus far, let alone the evidence provided for crustal processes and the mechanisms by which the effects of climatic change on the Earth’s surface can influence such processes. Other proposed linkages between rivers and landscape evolution exist, including the notion (deeply engrained in the ‘theoretical geomorphology’ literature) that much can be discerned from knickpoints in river long profiles. These short steep reaches are hypothesized to have formed in relation to a fall in base level, such as would occur at the coast in response to sea-level fall, and then propagated upstream over periods as long as millions of years (e.g., Bishop, 2007; Pritchard et al., 2009; Roberts and White, 2010; Hartley et al., 2011; cf. Bridgland and Westaway, 2012). The coincidence of knickpoints with outcrops of resistant bedrock has led others to suggest that such hard rocks will tend to give rise spontaneously to steeper gradient channels, irrespective of base-level influences, a view that is often countered by the suggestion that knickpoint recession is slower in such durable substrates and that this can explain the above-noted coincidence (see discussion of the Colorado, above).

While knick-points, typically marked by waterfalls and/or cataracts, can obviously erode in an upstream direction, it is doubtful whether any meaningful long-timescale record can be determined from them; most attempts to do so have largely ignored the effects of climatic fluctuation during the Quaternary, which will have caused numerous changes in base level and will also have had a profound influence on fluvial discharge and catchment processes (sufficient, indeed, to drive terrace generation with or without knickpoints). In this regard, the Colorado has once again featured in recent discussions (e.g., Bridgland and Westaway, 2012; Pederson and Tressler, 2012; Pederson et al., 2013; Fig. 9). The longitudinal profile of this river has a long 'knickzone' within the Grand Canyon, with a relatively steep downstream gradient; further upstream the gradient is much less. Theoretical geomorphological considerations lead to the association of steep reaches with relatively rapid fluvial incision and uplift; elaborate tectonic explanations have thus been proposed for uplift in the vicinity of the Grand Canyon on the basis that it is where Late Cenozoic uplift has been concentrated (e.g., Levander et al., 2011; Karlstrom et al., 2012). The realisation that rates of fluvial incision (regarded here as a proxy for uplift) are higher upstream of the Grand Canyon than within the canyon itself (e.g., Bridgland and Westaway, 2012; Pederson and Tressler, 2012; Pederson et al., 2013; Fig. 9A) undermines this prediction; the steep longitudinal gradient of the Grand Canyon is evidently a consequence of valley constriction as the river flows through highly lithified Palaeozoic and Precambrian rocks, thus simply reflecting an effect observed worldwide (Bridgland and Westaway, 2012). Despite their ingrained use, therefore, the theoretical geomorphological techniques can predict effects that are at odds with observations, as in this high-profile example. It is apparent that approaches based on the analysis of evidence from fluvial archives can provide better understanding of the long-timescale behaviour of river systems, as the examples presented in this review have illustrated.

9. Conclusions

The idea that the history of landscape evolution during the Quaternary can be discerned from the disposition of river terraces, implicit in the concept of 'denudation chronology' that was prevalent in the early–mid 20th Century, has been reinvigorated by the application of empirical data, as reviewed in this paper. Much has been learned, particularly from the comparison of sequences in different parts of the world. The main driver of fluvial activity, and of the changing activity required to form flights of terraces (even against a background of uplift), is seen to be climate change; it can be assumed that rivers respond rapidly to this and achieve equilibrium within each climate cycle. Conversely, the empirical evidence for correlation of terraces with climate cycles, which is available from the best-dated and most informative sequences (especially those richest in palaeontological evidence), is clearly suggestive of a causative mechanism.

Comparison of worldwide data has produced unforeseen but valuable information about the influence of crustal type on the development of relief and provides important corroboration of the role of lower-crustal mobility in the generation of sustainable epeirogenic uplift, by way of a positive feedback mechanism that enhances erosional isostasy. Such uplift is seen as essential for the formation of long sequences of river terraces, in which the river has become increasingly incised into the landscape, far below the level of its earlier deposits (also true of gorge reaches in resistant bedrock types). Areas in which this type of uplift has not occurred can be recognized from the different patterns of fluvial archive preservation they display, with the matching of such patterns to different crustal types further underling the importance of rheological mechanisms.

These ideas arise from datasets that are unusually precise in the Earth sciences. They are testable, in that they allow predictions about expected patterns of fluvial archive preservation.

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Figures:

Figure 1 . Quaternary terrace staircases: classic examples from NW Europe

A . River Thames (idealized sequences, after Bridgland, 2010)

i Middle Thames, west of London, the most complete sequence

ii Lower Thames, east of London – excellent preservation of the last four 100 ka climate cycles; also with archaeological (Lower and Middle Palaeolithic) evidence.

iii NE Essex, lower Middle Pleistocene – MIS 11 sequence

B. River Wipper, Thuringia, Germany: non-idealized cross section of a meander core at Bilzingsleben (after Mania, 1995; reproduced from Bridgland et al., 2004b)

C. River Somme, northern France, with preservation of climato-stratigraphical and Palaeolithic archaeological evidence, well constrained by geochronology (after Antoine et al., 2007; reproduced from Bridgland, 2010).

D. River Maas, Maastricht, Netherlands (after Van den Berg, 1994, and Westaway, 2002a, b). One of the longest terrace records in the world, with dating from biostratigraphy and geochronology, including palaeomagnetism. Reproduced from Bridgland and Westaway (2008b) with updated MIS attributions.

[Needs 2 page spread; two separate images]

Figure 2. Fluvial records from the East European Plain (after Matoshko et al., 2002, 2004; reproduced from Bridgland and Westaway, 2008b; see that publication for further explanation). For key, see Fig. 1 (only new ornaments shown here).

A. Location map, showing the key crustal blocks mentioned in the text;

B. Generalized transverse profile through the Middle–Lower Dniester terrace sediments, which are inset into Miocene fluvial basin-fill deposits;

C. Transverse profile, ~240 km long, across the Middle Dnieper basin, ~100 km downstream of Kiev;

D. Transverse profile through the deposits of the Upper Don near Voronezh;

E. Cross section through the sediments of the Lower Volga, in the region of the Pre-Caspian Block.

Figure 3. Fluvial archives in subsiding depocentres: the Lower Rhine (For key, see Fig. 1).

A. Schematic long profile of Rhine deposits beneath the central Netherlands and the submerged Rhine valley beneath the southernmost North Sea, showing stratigraphical relations with submerged terrace deposits of the River Thames (from Westaway and Bridgland, 2010).

B. Cross section through Late Pleistocene Rhine–Meuse palaeochannels beneath the central Netherlands (after Busschers et al. (2007). For location see D.

C. Stratigraphy of stacked Lower Rhine deposits in relation to terraces further upstream (extracted from Bridgland and Westaway, 2008b).

D. Map of the Lower Rhine channel system in the latest Pleistocene, showing the location of B (after Busschers et al., 2007).

Figure 4. Fluvial archives from cratonic crustal regions: The Vaal, South Africa

A. Map showing the course of the River Vaal through the Archaean Kaapvaal Craton. Excavated for alluvial diamonds, the fluvial archives here have been studied in some detail, as they are important sources of early Palaeolithic artefacts (Butzer et al., 1973; Helgren, 1977, 1978). Minimal vertical crustal movement over the last several Ma is indicated, in marked contrast with records from outside the craton, such as that from the coastal Sundays River system (indicated), in which an extensive terrace staircase has formed (Hattingh, 1994; Hattingh and Rust, 1999) on younger and more dynamic crust and is suggestive of ~450 m of uplift during only ~3 Ma (cf. Westaway et al., 2003; Bridgland and Westaway, 2008a, b).

B. Transverse profile through the 'terraces' of the Vaal within the Kapvaal Craton (after Helgren, 1978). Of the three 'Younger Gravel' members, A and B are thought to be Early Pleistocene, whereas C is biostratigraphically dated to the mid-Middle Pleistocene. In the Riverton Formation, Members I and II have yielded Acheulian artefacts (Middle Pleistocene); Member III has yielded Middle Palaeolithic artefacts, suggesting a late Middle Pleistocene or early Late Pleistocene age; and Members IV and V are Holocene (cf. Westaway et al., 2003).

C. Longer-timescale record from this area (after De Wit et al., 1997). The Wedburg and Proksch Koppie gravels have been attributed to the Miocene (e.g. De Wit et al., 1997; cf. Butzer et al., 1973; Helgren, 1978) and the Nooitgedacht gravel is thought to date from the Late Cretaceous or early in the Cenozoic (De Wit, 2004).

Figure 5. The Lower Tagus, Portugal: a Neogene basin inverted and incised in the Pleistocene. From Martins et al. (2010b) with modifications.

Figure 6. The marine oxygen isotope record for the last 1.8 Ma, based on the LR04 benthic δ18O stack constructed by Lisiecki and Raymo (2005) by the graphic correlation of 57 globally distributed benthic records. Note the change from shorter ~40 ka to longer ~100 ka cycles at the 'Mid-Pleistocene Revolution' (at around the transition from the Early to the Middle Pleistocene, which coincides with the Matuyama–Bruhnes magnetic reversal, in MIS 19). Reproduced from Bridgland et al. (2014).

Figure 7. The River Gediz, western Turkey.

A. Transverse profile through the Early Pleistocene terrace staircase as preserved beneath basalt the capping of the Burgaz Plateau. After Maddy et al. (2008, their figure 7).
B. The Burgaz Plateau at Kale Tepe, viewed from the south.
C. Transverse profile across the Gediz terrace staircase near Eynehan and Karabeyli, redrawn from Westaway et al. (2004), after Seyitoğlu (1997).

Figure 8. The River Shoalhaven, SE Australia, showing detail of Neogene valley fills and Pleistocene terraces incised through these (or into 'basement' where the Pre-

Quaternary and post Quaternary valleys diverge). There is sporadic preservation of even earlier valley-fill deposits. The Neogene evolution of the river system was complicated by basaltic eruptions that produced lava dams, leading to episodes of lacustrine deposition in the palaeovalleys and accounting for the siltstone facies (Nott, 1992).

A. Cross section, compiled from borehole data, through valley fill of the Mongarlowe Palaeochannel, near its confluence with the palaeo-Shoalhaven (after Nott, 1992, with modifications; for location see D);

B. Schematic cross section through valley fill (Nadgigomar Subgroup) of the palaeo-Shoalhaven in the region of Spa Creek (see D), showing dissection by the modern river (after Nott, 1992, with modifications);

C. Schematic cross section through the Shoalhaven at Larbet, showing post-inversion terraces (modified from Nott et al., 2002; inferred MIS correlations from Bridgland and Westaway, 2008a). For key to terrace colours see Fig. 1.

D. Map of the Middle Shoalhaven, showing the footprints of Late Cenozoic (Oligocene) palaeovalleys, as well as the outcrops of older valley-fill sediments. The Oligocene basalt flows that dammed the system are also shown.

Figure 9. Evidence from the Colorado catchment (for key to terrace colours see Fig. 1).

A. Map of the middle reach of the river within and upstream of the Grand Canyon. Modified from Figure 4(a) of Pederson et al. (2013), who provided sources of information for rates of fluvial incision (taken as a proxy for uplift); interpretation from Westaway and Bridgland (2014) is also incorporated. These rates are based on a variety of dating methods (luminescence and cosmogenic dating of terrace deposits; U-series dating of speleothems; tephrochronology; Ar–Ar dating of basalt flows that cap terrace deposits) and are time-averaged for different intervals during the Pleistocene. Note that the fastest uplift rates occur upstream of the Grand Canyon, in a region of widespread Late Cenozoic erosion, and also upstream of the regions with anomalous crustal and mantle properties (arising from earlier tectonic history) recognized by Levander et al. (2011), indicating the role of erosional isostasy in the uplift history in this region. The Pleistocene diversion of the Gunnison River into the Colorado at Grand Junction (Donahue et al., 2013), depicted here schematically, means that the incision by this tributary has not always served as a proxy for uplift (Westaway and Bridgland, 2014). However, the 0.15 mm a^{-1} uplift rate indicated, based on ~100 m of incision below the level of a terrace deposit containing tephra from the ~0.6 Ma Lava Creek B eruption of Yellowstone (reported by Donahue et al., 2013), represents a span of time well after this diversion, for which incision can indeed provide a proxy for uplift.

B. Schematic transverse profile across the Colorado terrace staircase in the vicinity of Grand Junction, Colorado, based on data from Scott et al. (2002), as interpreted by Westaway and Bridgland (2014).

C. Transverse profile across the Colorado terrace staircase at Lee's Ferry, Arizona, modified from Fig. 2 of Pederson et al. (2013) to show interpreted MIS correlations for the emplacement of the terrace deposits.

Figure 10. Records from Mesopotamia (for key to terrace colours see Fig. 1):

A. Idealized transverse profile through the terrace staircase of the River Euphrates in the Birecik area, ~50 km north of the Turkey–Syria border. Holocene flood deposits that overlie the terraces assigned to MIS 6 and 2 (cf. Kuzucuoğlu et al., 2004) are omitted (modified from Demir et al., 2008). Note that deposits considered, by analogy with Syria, to be Middle Pliocene are found up to ~200 m above the present river level here (e.g., Minzoni-Deroche and Sanlaville, 1988), much higher than their counterparts further downstream.

B. Idealized transverse profile through the terrace staircase of the Euphrates between Raqqa and Deir ez-Zor, showing Ar–Ar dating of basalts; Euphrates deposits above the level of the Halabiyeh upper gravel are omitted. Modified from Demir et al. (2007b).

C. Idealized transverse profile across the River Tigris at Diyarbakır, SE Turkey, showing the chronological constraint provided by multiple Ar–Ar dated basalts. Modified from Westaway et al. (2009b).

Figure 11. Back-tilting of the Bytham Formation as a result of differential crustal properties in Midland England. Modified from Rose (1994); additional data in blue. For explanation see Bridgland et al. (2014, chapter 6). For key to terrace colours see Fig. 1.

Figure 12. Records from areas of rapid uplift:

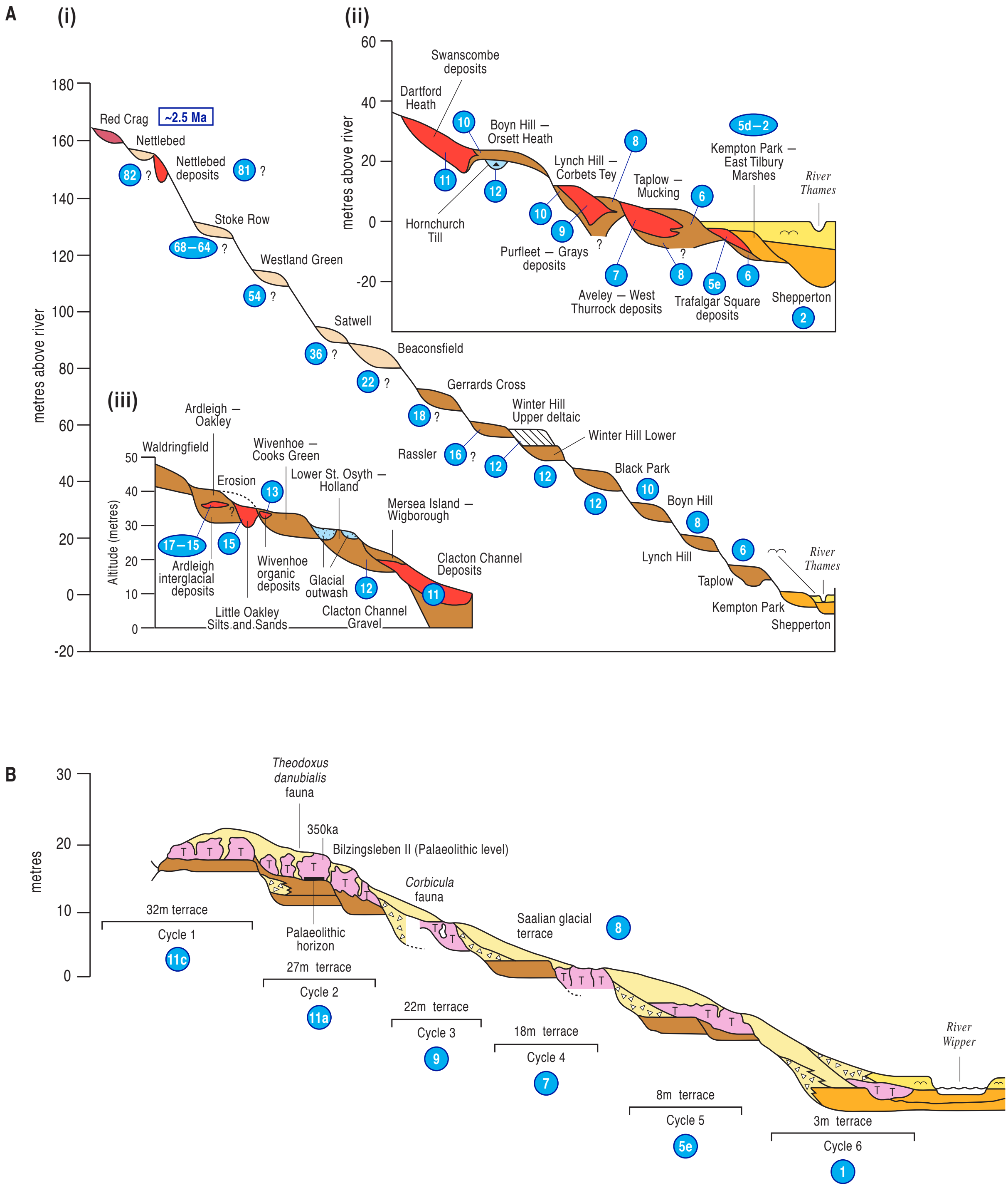
A. Cross-section across the Ceyhan valley at the location of the Aslantaş Dam, showing the disposition of basalt, dated to ~270 ka, and colluvial and terrace deposits. Modified from figure 8 of Seyrek et al. (2008).

B. NE–SW longitudinal profile of the Nahr el Kebir terraces (modified from Bridgland et al., 2008). Note the combination of deformed coastal terraces, from interglacials, and steeply graded colder-climate gravel terraces, with intersect with the much shallower downstream gradient of the modern (Holocene) valley floor. For key to terrace colours see Fig. 1.

Figure 13. Comparison between areas inside and outside the LGM ice limit in eastern England: contrasting terrace staircases illustrated at the same vertical scale.

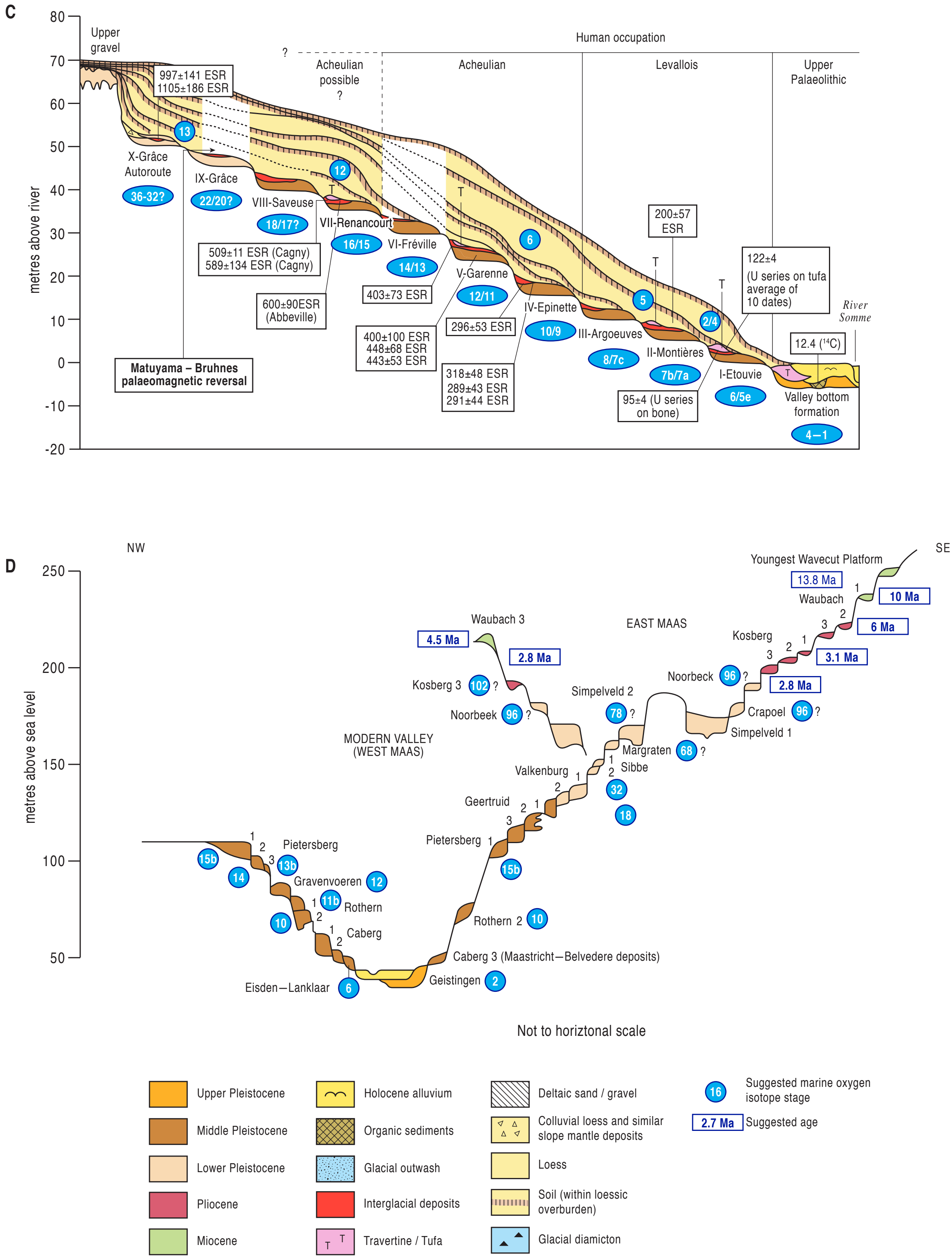
A. the ~450 ka record in the Lower Thames (after Bridgland, 2006; see also Fig. 1Aii)
B. the <20 ka record in the Middle Ure, North Yorkshire, showing incision into the landscape subsequent to LGM deglaciation (after Bridgland et al., 2011).

Figure 1A



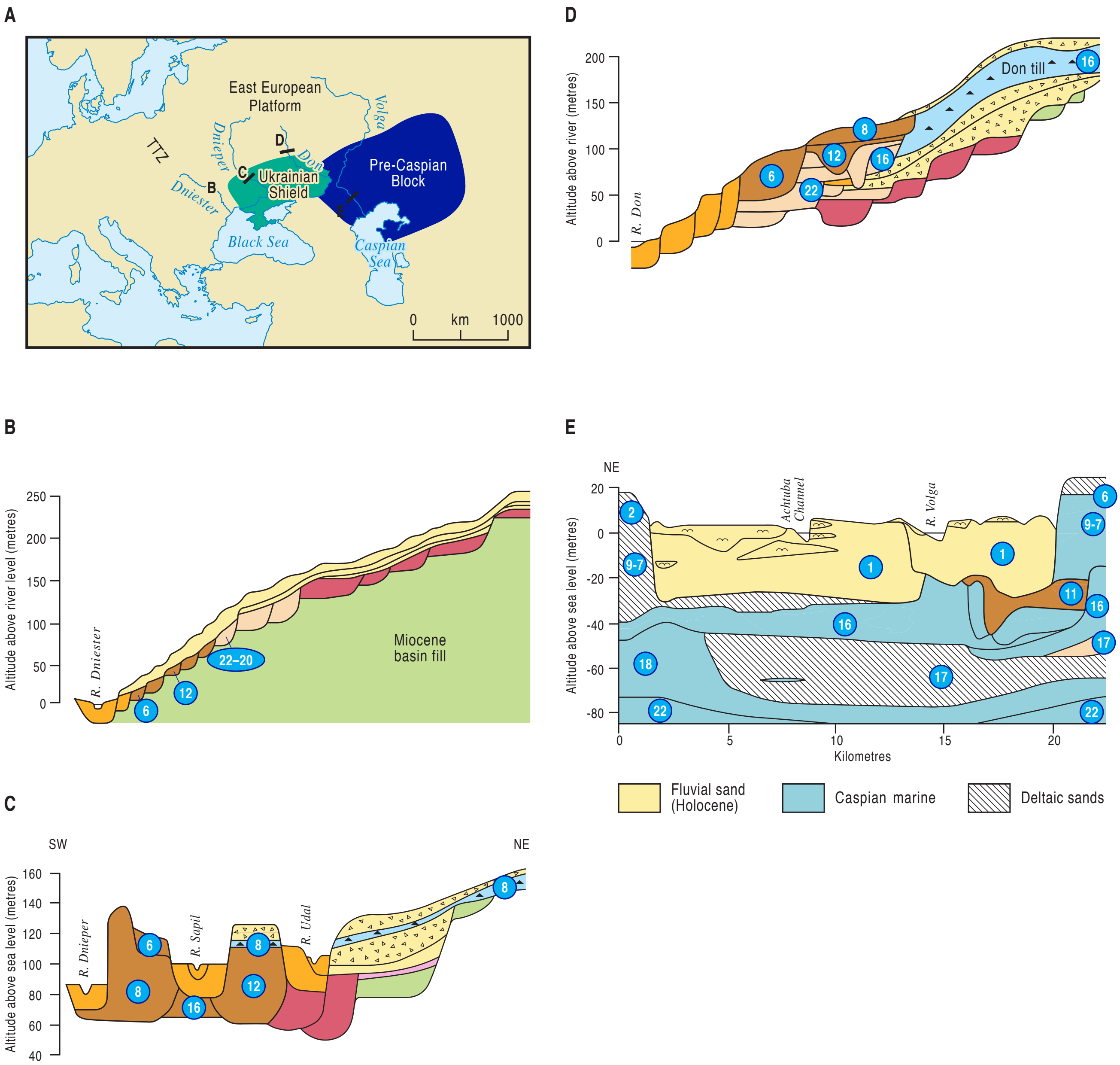
A
(i)
(ii)
(iii)
metres above river
Red Crag
~2.5 Ma
Nettlebed
Nettlebed deposits
Stoke Row
Westland Green
Satwell
Beaconsfield
Gerrards Cross
Winter Hill Upper deltaic
Winter Hill Lower
Rassler
Black Park
Boyn Hill
Lynch Hill
Taplow
Kempton Park
Shepperton
River Thames
Swanscombe deposits
Dartford Heath
Boyn Hill – Orsett Heath
Hornchurch Till
Lynch Hill – Corbets Tey
Purfleet – Grays deposits
Taplow – Mucking
Aveley – West Thurrock deposits
Kempton Park – East Tilbury Marshes
Trafalgar Square deposits
Altitude (metres)
Ardleigh – Oakley
Waldringfield
Wivenhoe – Cooks Green
Erosion
Lower St. Osyth – Holland
Mersea Island – Wigborough
Ardleigh interglacial deposits
Wivenhoe organic deposits
Glacial outwash
Clacton Channel Deposits
Little Oakley Silts and Sands
Clacton Channel Gravel
B
metres
Theodoxus danubialis fauna
350ka
Bilzingsleben II (Palaeolithic level)
Palaeolithic horizon
Corbicula fauna
Saalian glacial terrace
32m terrace
Cycle 1
27m terrace
Cycle 2
22m terrace
Cycle 3
18m terrace
Cycle 4
8m terrace
Cycle 5
3m terrace
Cycle 6
River Wipper

Figure 1B



C
metres above river
Upper gravel
Human occupation
?
Acheulian possible ?
Acheulian
Levallois
Upper Palaeolithic
997±141 ESR
1105±186 ESR
X-Grâce Autoroute
36-32?
IX-Grâce
22/20?
VIII-Saveuse
18/17?
VII-Renancourt
16/15
VI-Fréville
14/13
V-Garenne
12/11
IV-Epinette
10/9
III-Argoeuves
8/7c
II-Montières
7b/7a
I-Etouvie
6/5e
Valley bottom formation
4—1
13
12
6
5
2/4
T
509±11 ESR (Cagny)
589±134 ESR (Cagny)
600±90ESR (Abbeville)
403±73 ESR
400±100 ESR
448±68 ESR
443±53 ESR
296±53 ESR
318±48 ESR
289±43 ESR
291±44 ESR
200±57 ESR
95±4 (U series on bone)
122±4 (U series on tufa average of 10 dates)
12.4 (14C)
River Somme
Matuyama – Bruhnes palaeomagnetic reversal
D
NW
SE
metres above sea level
Youngest Wavecut Platform
13.8 Ma
Waubach
10 Ma
6 Ma
Kosberg
3.1 Ma
2.8 Ma
Noorbeck
96
Crapoel
Simpelveld 1
EAST MAAS
Waubach 3
4.5 Ma
2.8 Ma
Kosberg 3
102
Noorbeek
Simpelveld 2
78
Margraten
68
Sibbe
32
18
MODERN VALLEY (WEST MAAS)
Valkenburg
Geertruid
Pietersberg
15b
Rothern 2
10
Caberg 3 (Maastricht—Belvedere deposits)
Geistingen
2
Eisden—Lanklaar
6
Caberg
Rothern
11b
Gravenvoeren
12
13b
14
Not to horiztonal scale
Upper Pleistocene
Middle Pleistocene
Lower Pleistocene
Pliocene
Miocene
Holocene alluvium
Organic sediments
Glacial outwash
Interglacial deposits
Travertine / Tufa
Deltaic sand / gravel
Colluvial loess and similar slope mantle deposits
Loess
Soil (within loessic overburden)
Glacial diamicton
16
Suggested marine oxygen isotope stage
2.7 Ma
Suggested age

Figure 2



A
East European Platform
TTZ
Dniester
Dnieper
Don
Volga
B
C
D
E
Ukrainian Shield
Pre-Caspian Block
Black Sea
Caspian Sea
0 km 1000
B
Altitude above river level (metres)
R. Dniester
Miocene basin fill
C
SW
NE
Altitude above sea level (metres)
R. Dnieper
R. Sapil
R. Udal
D
Altitude above river (metres)
R. Don
Don till
E
NE
Altitude above sea level (metres)
Achtuba Channel
R. Volga
Kilometres
Fluvial sand (Holocene)
Caspian marine
Deltaic sands

Figure 3

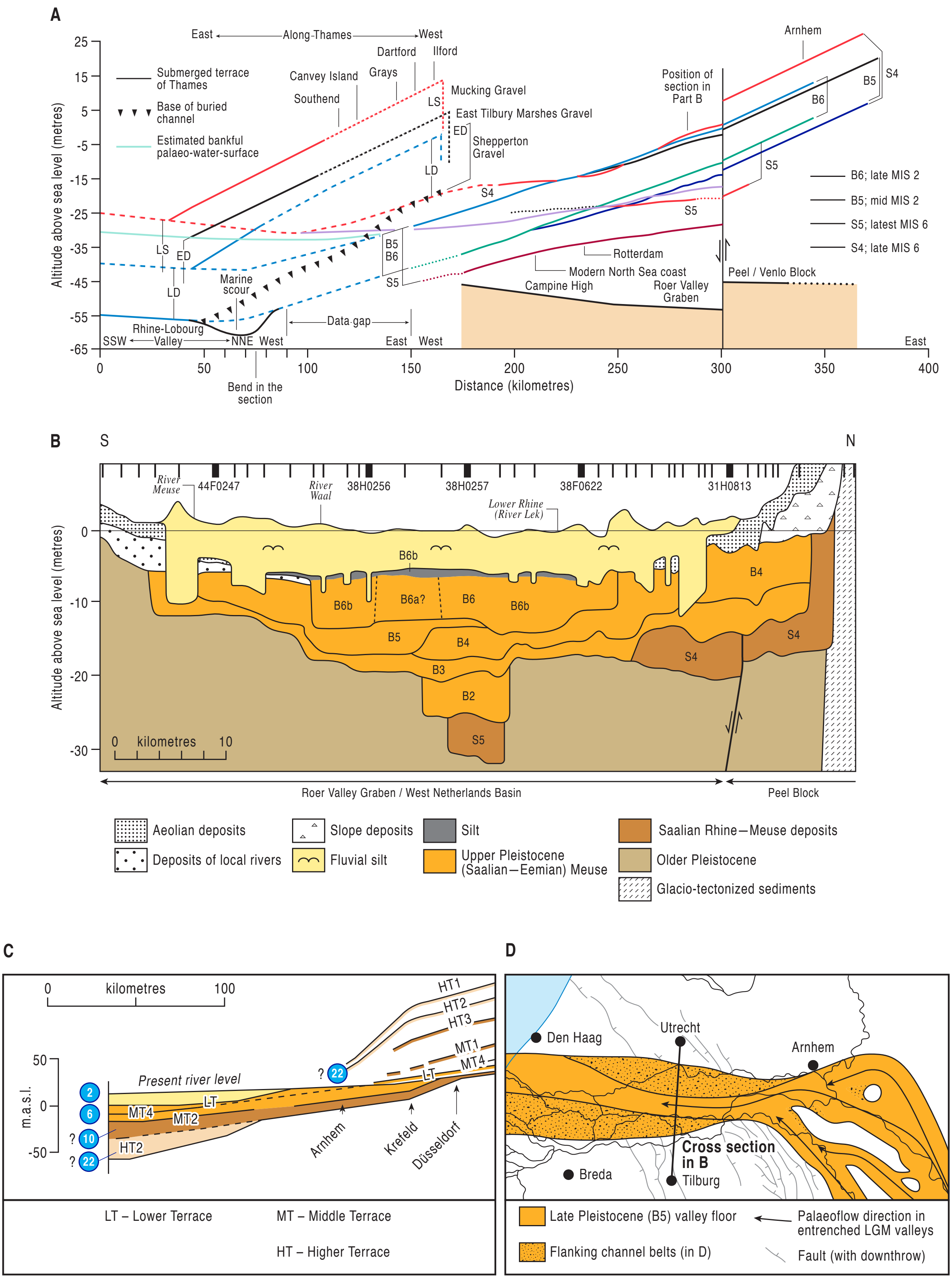

A
East ← Along-Thames → West
Altitude above sea level (metres)
Submerged terrace of Thames
Base of buried channel
Estimated bankful palaeo-water-surface
Canvey Island
Southend
Grays
Dartford
Ilford
Mucking Gravel
East Tilbury Marshes Gravel
Shepperton Gravel
LS
ED
LD
Marine scour
Rhine-Lobourg Valley
SSW
NNE
West
East
Data gap
Bend in the section
Position of section in Part B
Arnhem
Rotterdam
Modern North Sea coast
Campine High
Roer Valley Graben
Peel / Venlo Block
B6; late MIS 2
B5; mid MIS 2
S5; latest MIS 6
S4; late MIS 6
Distance (kilometres)
B
S
N
River Meuse
44F0247
River Waal
38H0256
38H0257
Lower Rhine (River Lek)
38F0622
31H0813
B6b
B6a?
B6
B5
B4
B3
B2
S5
S4
kilometres
Roer Valley Graben / West Netherlands Basin
Peel Block
Aeolian deposits
Deposits of local rivers
Slope deposits
Fluvial silt
Silt
Upper Pleistocene (Saalian—Eemian) Meuse
Saalian Rhine—Meuse deposits
Older Pleistocene
Glacio-tectonized sediments
C
kilometres
m.a.s.l.
Present river level
HT1
HT2
HT3
MT1
MT4
MT2
LT
Arnhem
Krefeld
Düsseldorf
LT – Lower Terrace
MT – Middle Terrace
HT – Higher Terrace
D
Den Haag
Utrecht
Arnhem
Breda
Tilburg
Cross section in B
Late Pleistocene (B5) valley floor
Flanking channel belts (in D)
Palaeoflow direction in entrenched LGM valleys
Fault (with downthrow)

Figure 4

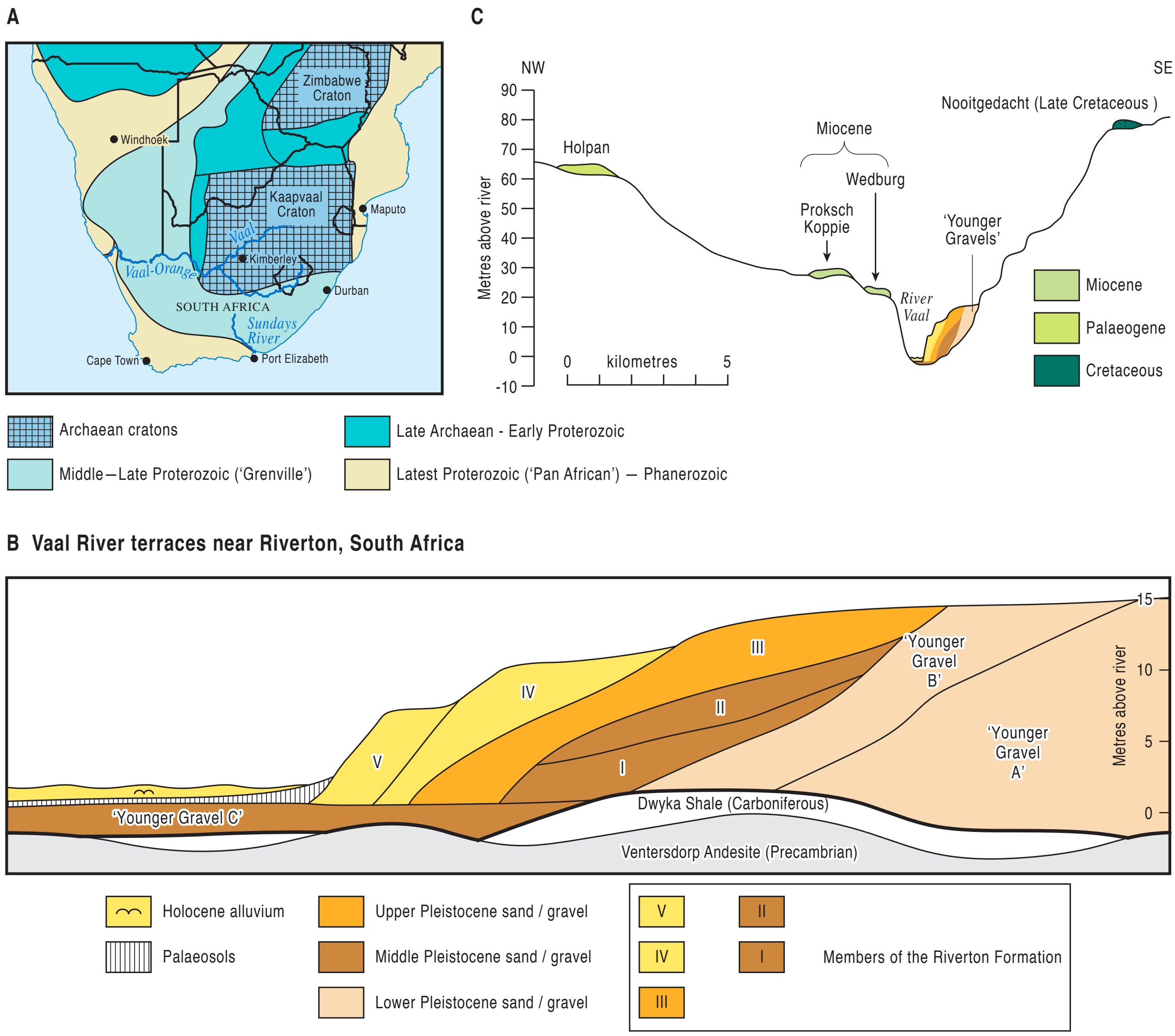

A
Zimbabwe Craton
Windhoek
Kaapvaal Craton
Maputo
Vaal
Kimberley
Vaal-Orange
Durban
SOUTH AFRICA
Sundays River
Cape Town
Port Elizabeth
Archaean cratons
Late Archaean - Early Proterozoic
Middle—Late Proterozoic ('Grenville')
Latest Proterozoic ('Pan African') — Phanerozoic
C
NW
SE
Metres above river
90
80
70
60
50
40
30
20
10
0
-10
Holpan
Miocene
Wedburg
Proksch Koppie
Nooitgedacht (Late Cretaceous)
'Younger Gravels'
River Vaal
0
kilometres
5
Miocene
Palaeogene
Cretaceous
B Vaal River terraces near Riverton, South Africa
III
IV
V
II
I
'Younger Gravel B'
'Younger Gravel A'
'Younger Gravel C'
Dwyka Shale (Carboniferous)
Ventersdorp Andesite (Precambrian)
Metres above river
15
10
5
0
Holocene alluvium
Palaeosols
Upper Pleistocene sand / gravel
Middle Pleistocene sand / gravel
Lower Pleistocene sand / gravel
V
IV
III
II
I
Members of the Riverton Formation

Figure 5

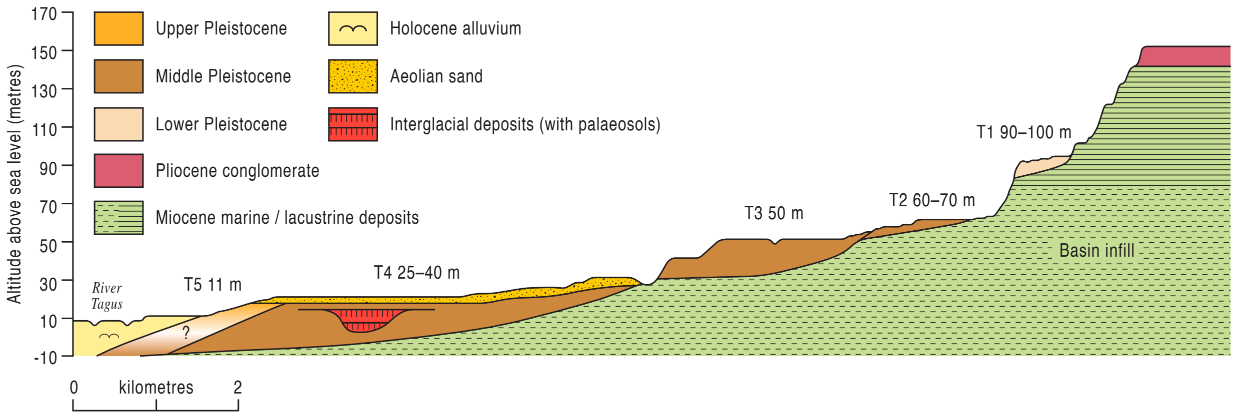

Altitude above sea level (metres)
170
150
130
110
90
70
50
30
10
-10
Upper Pleistocene
Middle Pleistocene
Lower Pleistocene
Pliocene conglomerate
Miocene marine / lacustrine deposits
Holocene alluvium
Aeolian sand
Interglacial deposits (with palaeosols)
River Tagus
T5 11 m
T4 25–40 m
T3 50 m
T2 60–70 m
T1 90–100 m
Basin infill
?
0
kilometres
2

Figure 6

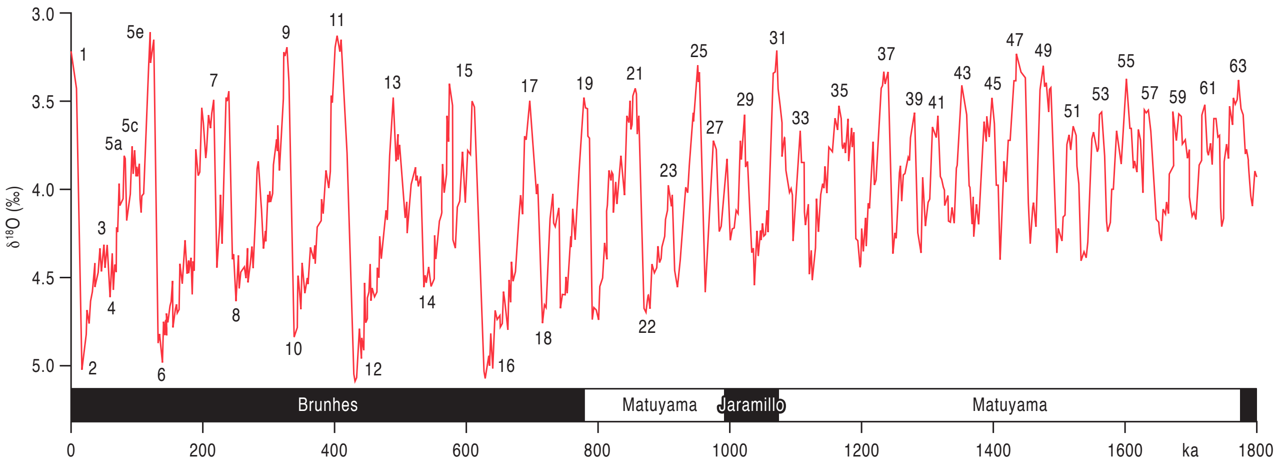

δ18O (‰)
3.0
3.5
4.0
4.5
5.0
1
2
3
4
5a
5c
5e
6
7
8
9
10
11
12
13
14
15
16
17
18
19
21
22
23
25
27
29
31
33
35
37
39
41
43
45
47
49
51
53
55
57
59
61
63
Brunhes
Matuyama
Jaramillo
Matuyama
0
200
400
600
800
1000
1200
1400
1600
ka
1800

Figure 7

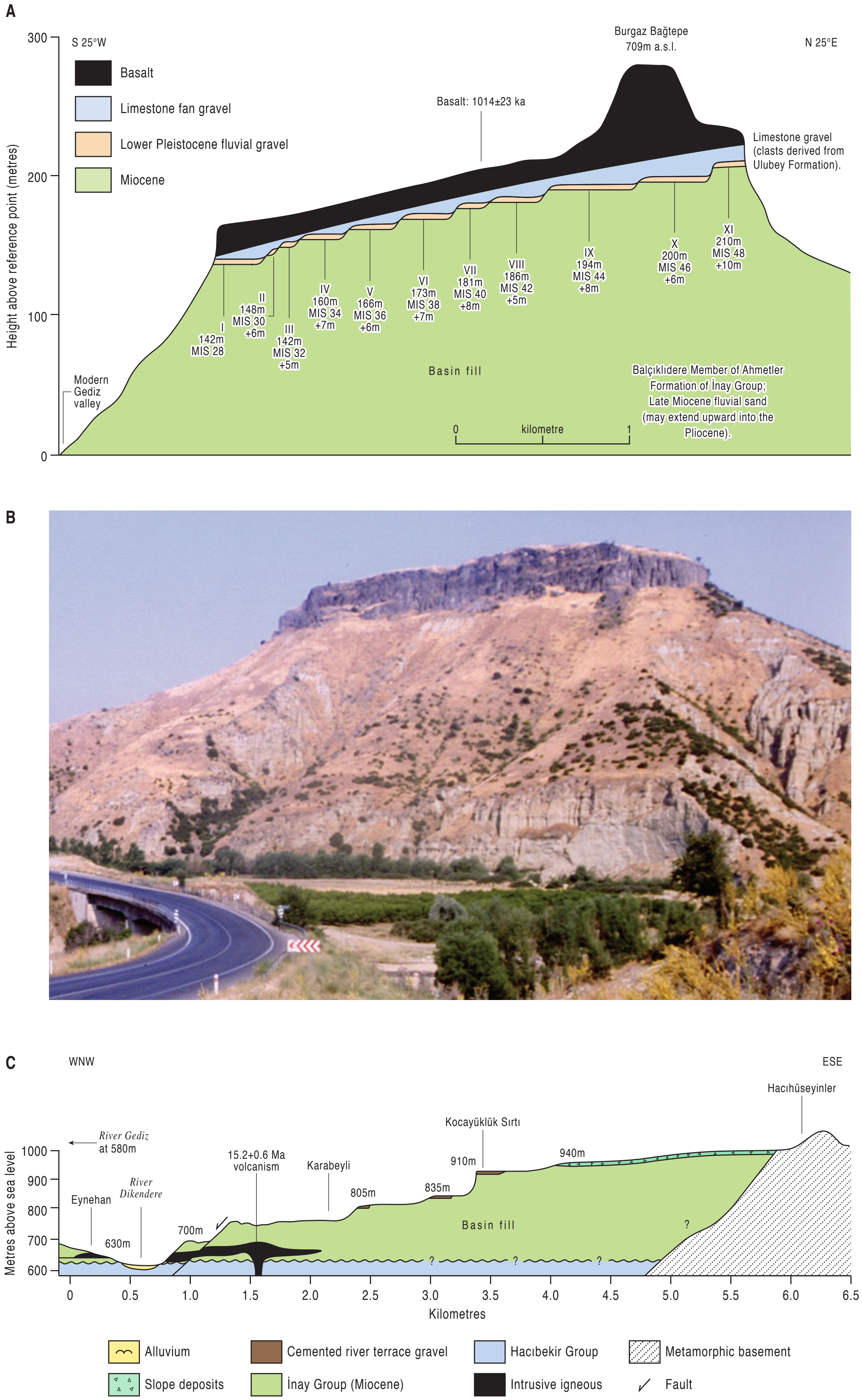

A
S 25°W
N 25°E
Height above reference point (metres)
300
200
100
0
Basalt
Limestone fan gravel
Lower Pleistocene fluvial gravel
Miocene
Burgaz Bağtepe
709m a.s.l.
Basalt: 1014±23 ka
Limestone gravel
(clasts derived from
Ulubey Formation).
Modern
Gediz
valley
I
142m
MIS 28
II
148m
MIS 30
+6m
III
142m
MIS 32
+5m
IV
160m
MIS 34
+7m
V
166m
MIS 36
+6m
VI
173m
MIS 38
+7m
VII
181m
MIS 40
+8m
VIII
186m
MIS 42
+5m
IX
194m
MIS 44
+8m
X
200m
MIS 46
+6m
XI
210m
MIS 48
+10m
Basin fill
Balçıklıdere Member of Ahmetler
Formation of İnay Group;
Late Miocene fluvial sand
(may extend upward into the
Pliocene).
0
kilometre
1
B
C
WNW
ESE
Metres above sea level
1000
900
800
700
600
River Gediz
at 580m
Eynehan
River
Dikendere
630m
700m
15.2+0.6 Ma
volcanism
Karabeyli
805m
835m
910m
Kocayüklük Sırtı
940m
Hacıhüseyinler
Basin fill
0
0.5
1.0
1.5
2.0
2.5
3.0
3.5
4.0
4.5
5.0
5.5
6.0
6.5
Kilometres
Alluvium
Slope deposits
Cemented river terrace gravel
İnay Group (Miocene)
Hacıbekir Group
Intrusive igneous
Metamorphic basement
Fault

Figure 8

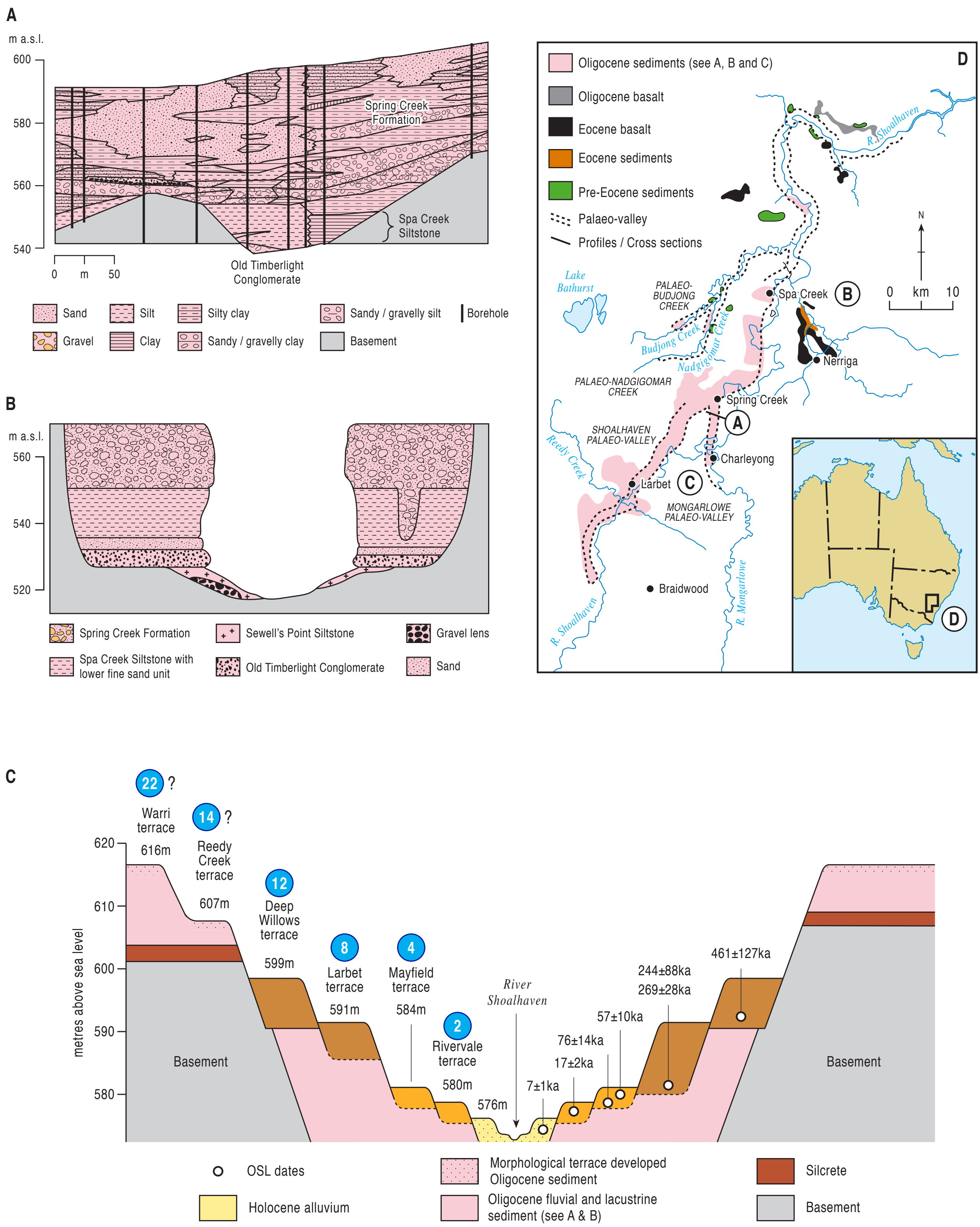
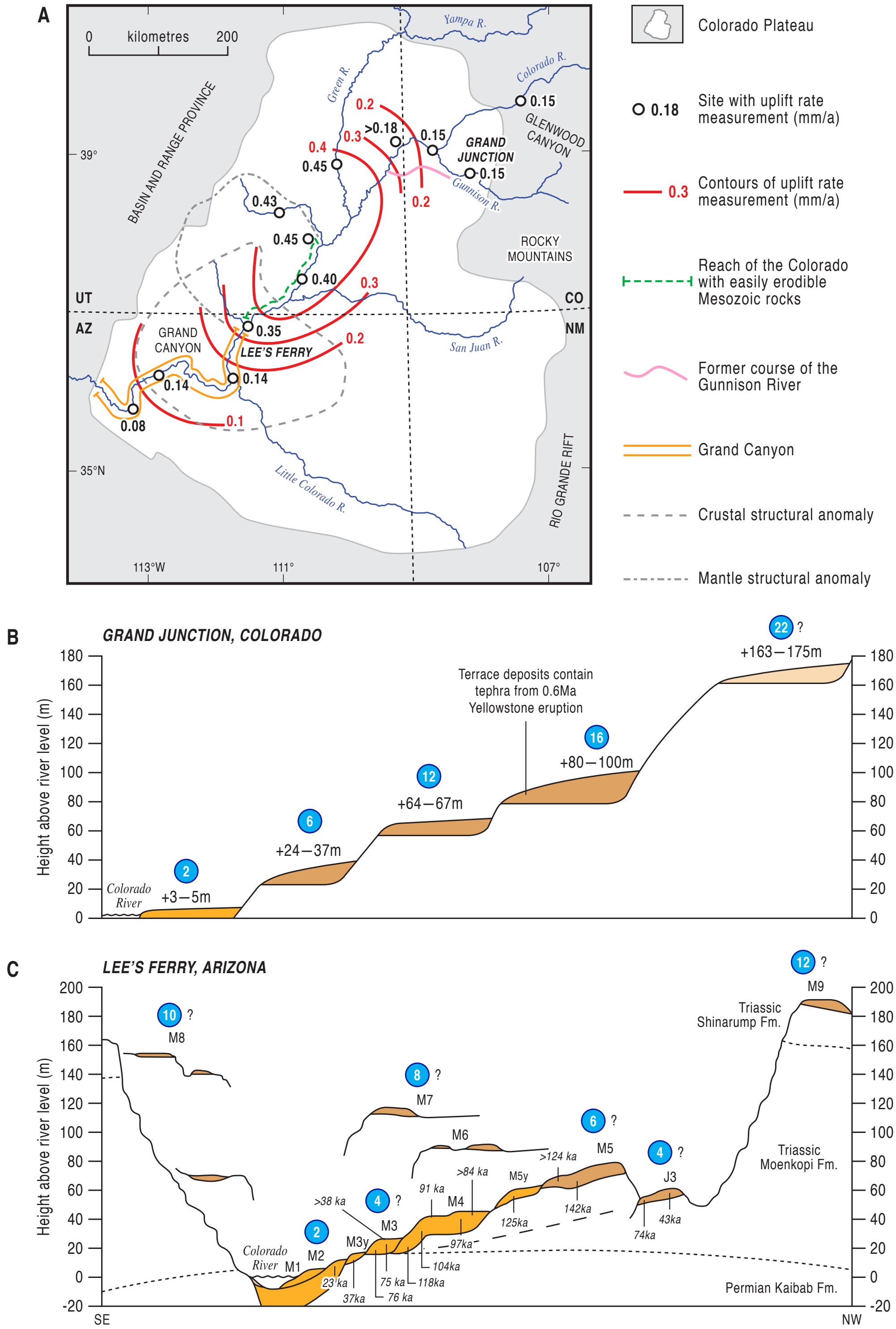
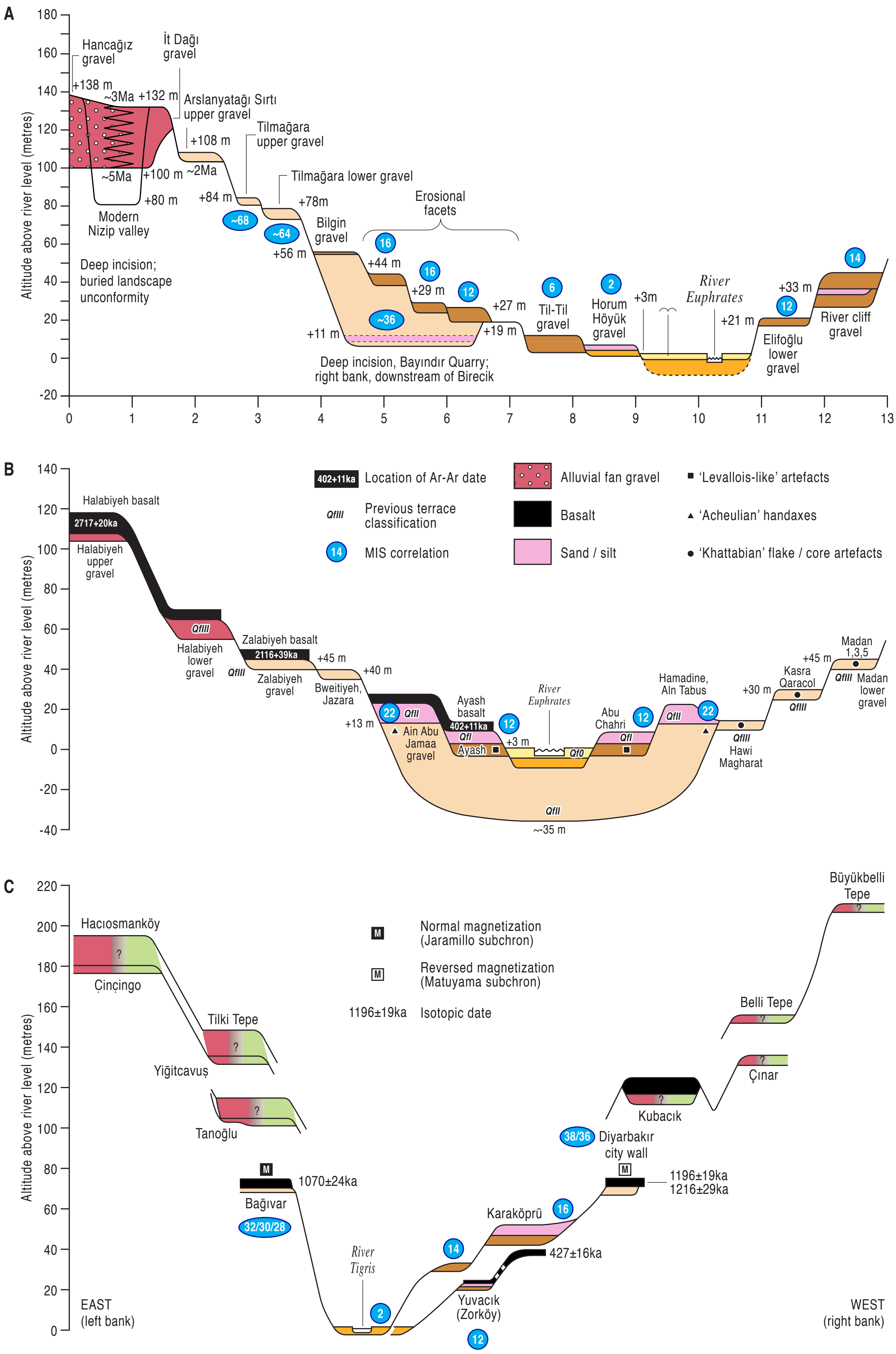

A
m a.s.l.
600
580
560
540
Spring Creek Formation
Spa Creek Siltstone
Old Timberlight Conglomerate
0 m 50
Sand
Silt
Silty clay
Sandy / gravelly silt
Borehole
Gravel
Clay
Sandy / gravelly clay
Basement
B
m a.s.l.
560
540
520
Spring Creek Formation
Sewell's Point Siltstone
Gravel lens
Spa Creek Siltstone with lower fine sand unit
Old Timberlight Conglomerate
Sand
D
Oligocene sediments (see A, B and C)
Oligocene basalt
Eocene basalt
Eocene sediments
Pre-Eocene sediments
Palaeo-valley
Profiles / Cross sections
R. Shoalhaven
Lake Bathurst
PALAEO-BUDJONG CREEK
Budjong Creek
Nadgigomar Creek
PALAEO-NADGIGOMAR CREEK
Spa Creek
B
Nerriga
Spring Creek
A
SHOALHAVEN PALAEO-VALLEY
Reedy Creek
Charleyong
Larbet
C
MONGARLOWE PALAEO-VALLEY
Braidwood
R. Mongarlowe
R. Shoalhaven
N
0 km 10
D
C
22 ?
Warri terrace
616m
14 ?
Reedy Creek terrace
607m
12
Deep Willows terrace
599m
8
Larbet terrace
591m
4
Mayfield terrace
584m
2
Rivervale terrace
580m
576m
River Shoalhaven
7±1ka
17±2ka
76±14ka
57±10ka
244±88ka
269±28ka
461±127ka
metres above sea level
620
610
600
590
580
Basement
Basement
OSL dates
Holocene alluvium
Morphological terrace developed Oligocene sediment
Oligocene fluvial and lacustrine sediment (see A & B)
Silcrete
Basement

Figure 9



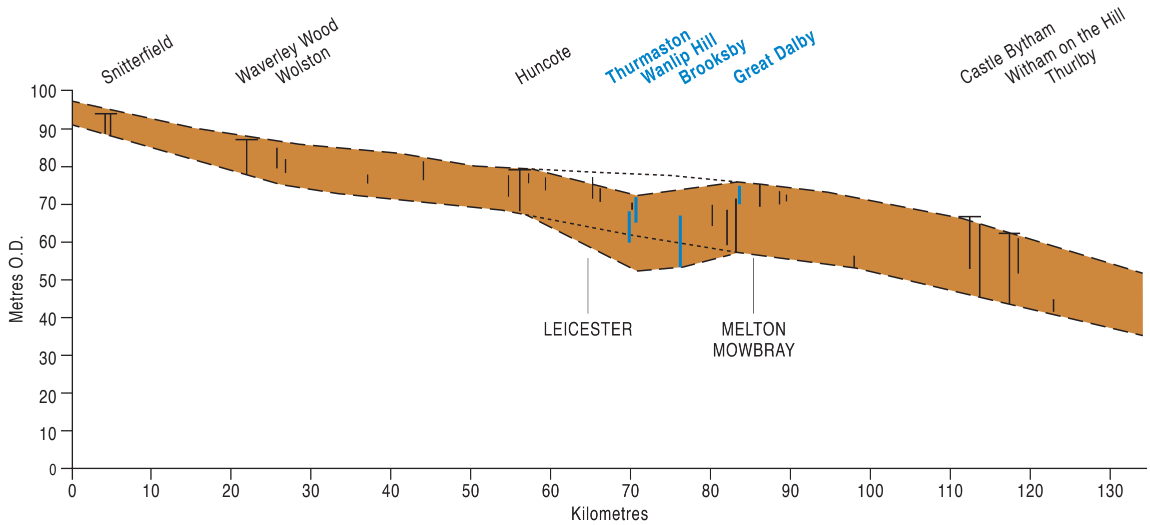
A
0 kilometres 200
BASIN AND RANGE PROVINCE
GRAND CANYON
LEE'S FERRY
GRAND JUNCTION
GLENWOOD CANYON
ROCKY MOUNTAINS
RIO GRANDE RIFT
UT
AZ
CO
NM
39°
35°N
113°W
111°
107°
Yampa R.
Green R.
Colorado R.
Gunnison R.
San Juan R.
Little Colorado R.
Colorado Plateau
Site with uplift rate measurement (mm/a)
Contours of uplift rate measurement (mm/a)
Reach of the Colorado with easily erodible Mesozoic rocks
Former course of the Gunnison River
Grand Canyon
Crustal structural anomaly
Mantle structural anomaly
B
GRAND JUNCTION, COLORADO
Height above river level (m)
Colorado River
+3–5m
+24–37m
+64–67m
+80–100m
+163–175m
Terrace deposits contain tephra from 0.6Ma Yellowstone eruption
C
LEE'S FERRY, ARIZONA
Height above river level (m)
Triassic Shinarump Fm.
Triassic Moenkopi Fm.
Permian Kaibab Fm.
Colorado River
M1 M2 M3y M3 M4 M5y M5 M6 M7 M8 M9 J3
SE
NW

Figure 10



A
Altitude above river level (metres)
Hancağız gravel
İt Dağı gravel
+138 m
~3Ma +132 m
Arslanyatağı Sırtı upper gravel
Tilmağara upper gravel
+108 m
~5Ma
+100 m ~2Ma
Tilmağara lower gravel
+80 m
+84 m
+78m
Modern Nizip valley
Bilgin gravel
Erosional facets
+56 m
Deep incision; buried landscape unconformity
+44 m
+29 m
+27 m
+19 m
+11 m
Til-Til gravel
Horum Höyük gravel
+3m
River Euphrates
+21 m
+33 m
Elifoğlu lower gravel
River cliff gravel
Deep incision, Bayındır Quarry; right bank, downstream of Birecik
B
Halabiyeh basalt
2717±20ka
Halabiyeh upper gravel
QfIII
Halabiyeh lower gravel
Zalabiyeh basalt
2116±39ka
+45 m
Zalabiyeh gravel
+40 m
Bweitiyeh, Jazara
+13 m
Ain Abu Jamaa gravel
Ayash basalt
402±11ka
Ayash
+3 m
River Euphrates
Qf0
Abu Chahri
Hamadine, Aln Tabus
+30 m
Hawi Magharat
Kasra Qaracol
+45 m
Madan 1,3,5
Madan lower gravel
~-35 m
402±11ka Location of Ar-Ar date
QfIII Previous terrace classification
14 MIS correlation
Alluvial fan gravel
Basalt
Sand / silt
'Levallois-like' artefacts
'Acheulian' handaxes
'Khattabian' flake / core artefacts
C
Hacıosmanköy
Çinçingo
Tilki Tepe
Yiğitcavuş
Tanoğlu
Bağıvar
1070±24ka
32/30/28
River Tigris
Yuvacık (Zorköy)
427±16ka
Karaköprü
Diyarbakır city wall
38/36
1196±19ka
1216±29ka
Kubacık
Çınar
Belli Tepe
Büyükbelli Tepe
Normal magnetization (Jaramillo subchron)
Reversed magnetization (Matuyama subchron)
1196±19ka Isotopic date
EAST (left bank)
WEST (right bank)

Figure 11



Snitterfield
Waverley Wood
Wolston
Huncote
Thurmaston
Wanlip Hill
Brooksby
Great Dalby
Castle Bytham
Witham on the Hill
Thurlby
LEICESTER
MELTON MOWBRAY
Metres O.D.
Kilometres

Figure 12

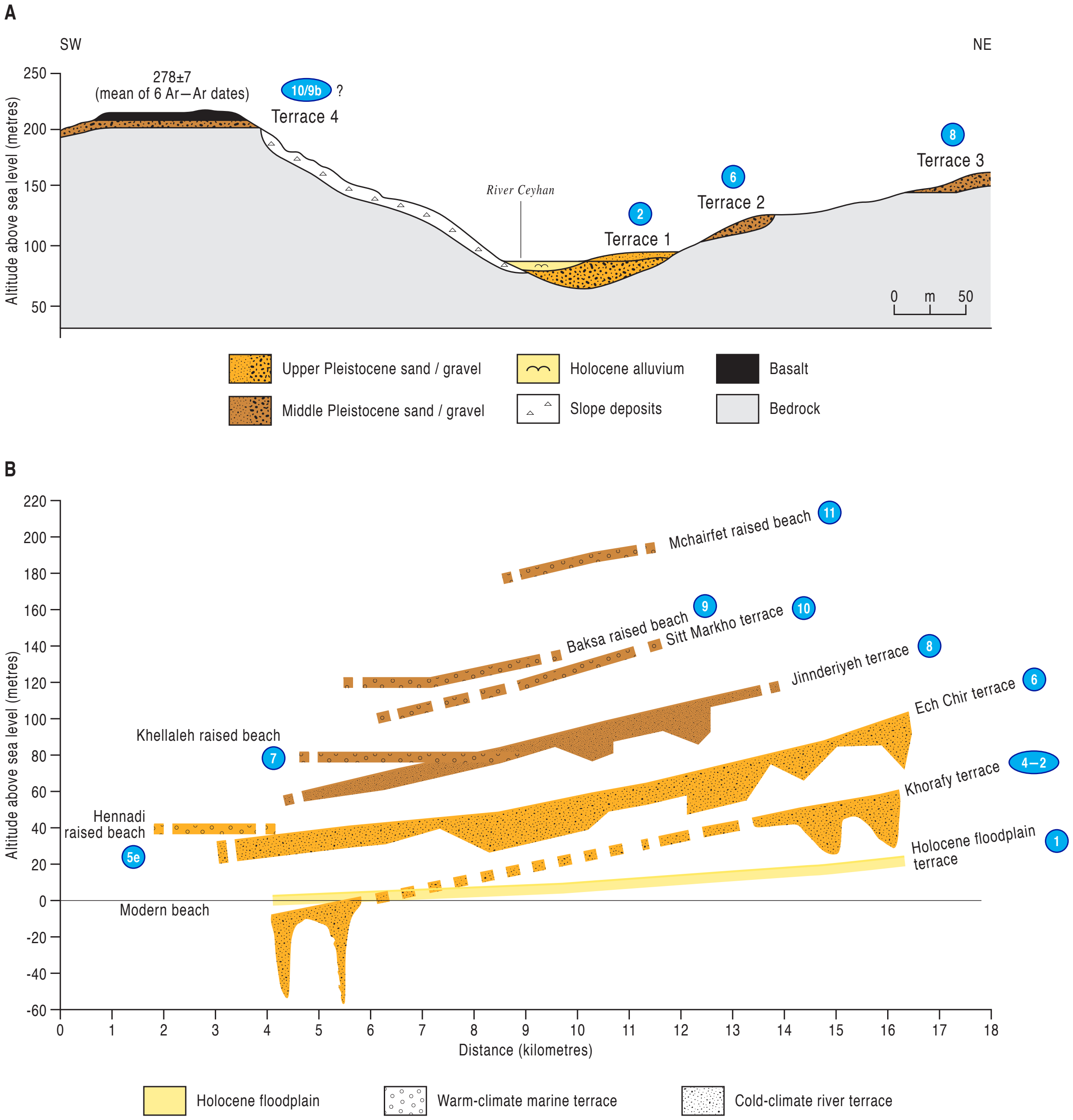
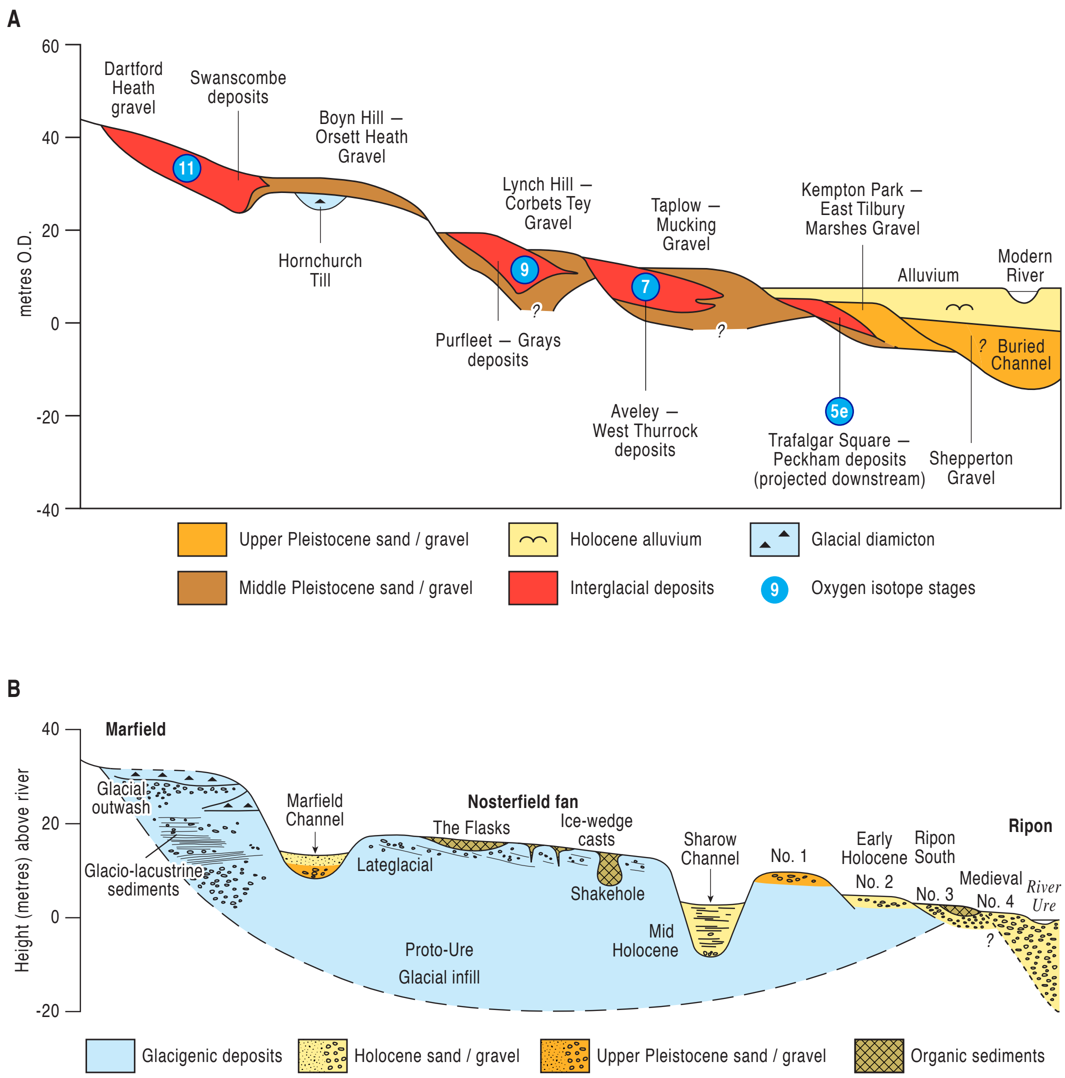

A
SW
NE
278±7
(mean of 6 Ar—Ar dates)
10/9b ?
Terrace 4
River Ceyhan
2
Terrace 1
6
Terrace 2
8
Terrace 3
Altitude above sea level (metres)
250
200
150
100
50
0 m 50
Upper Pleistocene sand / gravel
Middle Pleistocene sand / gravel
Holocene alluvium
Slope deposits
Basalt
Bedrock
B
Mchairfet raised beach 11
Baksa raised beach 9
Sitt Markho terrace 10
Jinnderiyeh terrace 8
Ech Chir terrace 6
Khorafy terrace 4–2
Holocene floodplain terrace 1
Khellaleh raised beach
7
Hennadi raised beach
5e
Modern beach
Altitude above sea level (metres)
220
200
180
160
140
120
100
80
60
40
20
0
-20
-40
-60
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18
Distance (kilometres)
Holocene floodplain
Warm-climate marine terrace
Cold-climate river terrace

Figure 13


A
60
40
20
0
-20
-40
metres O.D.
Dartford Heath gravel
Swanscombe deposits
Boyn Hill – Orsett Heath Gravel
Hornchurch Till
Lynch Hill – Corbets Tey Gravel
Purfleet – Grays deposits
Taplow – Mucking Gravel
Aveley – West Thurrock deposits
Kempton Park – East Tilbury Marshes Gravel
Alluvium
Modern River
? Buried Channel
Trafalgar Square – Peckham deposits (projected downstream)
Shepperton Gravel
11
9
7
5e
?
?
Upper Pleistocene sand / gravel
Holocene alluvium
Glacial diamicton
Middle Pleistocene sand / gravel
Interglacial deposits
Oxygen isotope stages
B
40
20
0
-20
Height (metres) above river
Marfield
Glacial outwash
Glacio-lacustrine sediments
Marfield Channel
Nosterfield fan
The Flasks
Ice-wedge casts
Lateglacial
Shakehole
Sharow Channel
Mid Holocene
No. 1
Early Holocene No. 2
Ripon South
Ripon
No. 3
Medieval No. 4
River Ure
?
Proto-Ure Glacial infill
Glacigenic deposits
Holocene sand / gravel
Upper Pleistocene sand / gravel
Organic sediments