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A conceptual model of valley incision, planation and terrace formation during cold and arid permafrost conditions of Pleistocene southern England

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ABSTRACT

Staircases of gravelly river terrace deposits in southern England occupy valleys typically underlain by frost-susceptible and brecciated bedrocks. The valleys developed during the Quaternary by alternating episodes of (1) brecciation, incision and planation through the bedrock, forming wide low-relief erosion surfaces; and (2) aggradation in braidplains of gravel a few meters thick that bury the erosion surfaces. A conceptual model to account for some of the terraces proposes that brecciation resulted from ice segregation in the ice-rich layer in the upper meters of Pleistocene permafrost, making them vulnerable to fluvial thermal erosion and therefore predisposing the bedrock to planation. The low gradients of the valleys were adjusted such that rivers transferred fine materials out of the basins but lacked the competence to remove gravel, which therefore accumulated within floodplains. The model challenges the prevailing view of incision during climate transitions. It attributes incision and planation to very cold and arid permafrost conditions, when rivers had limited discharges and hillslopes supplied limited volumes of stony debris into valley bottoms.

Introduction

A fundamental question about the landscape evolution of southern England is: *when and how was the current landscape carved out of the low-relief surface inherited from the Tertiary?* Gibbard and Lewin (2003), reviewing the Tertiary history of the major rivers of this area, inferred that this long period provided remarkably little evidence of fluvial activity; by the end of the Miocene the low-relief landscape was gently graded to the east and southeast, drained by rivers lacking the competence to move coarse material (Belshaw, 2007). They concluded that "today's deeply incised river valleys must be the product of the substantial, rapid climate changes that characterize the Pleistocene". Gibbard and Lewin (2002) reported that there is no evidence for incision during the interglacial episodes of the middle and late Pleistocene. Maddy et al. (2001), Bridgland (2006), Bridgland and Westaway (2008) and Stemerdink et al. (2010) attributed incision to climate transitions: major incision to the time of climate warming from periglacial to interglacial periods, with limited incision occurring during cooling into periglacial periods (Fig. 1). These suggestions implicate Pleistocene ground freezing and thawing as factors effecting incision, but the precise processes and timing remain to be identified. A general discussion of how and when lateral planation and vertical bedrock incision were carried out by European rivers during the Late Cenozoic is provided by Gibbard and Lewin (2009).

This paper focuses on the processes of river and valley incision in Pleistocene southern England (Fig. 2), and their timing in the periglacial–interglacial cycle. We describe the morphology, gradients and stratigraphy of terraces, paying special attention to the broad erosion surface between gravelly terrace deposits and the underlying bedrock. We limit our discussion to bedrocks that are frost-susceptible and brecciated in the upper meters—Mesozoic clays and the Chalk, for example—and interpret the stratigraphy and sediments in the context of fluvial and periglacial processes. We challenge the prevailing view of incision during climate transitions and set out a conceptual model of valley incision, planation and terrace formation during cold and arid permafrost conditions.

Terraces and floodplain deposits

Terrace morphology and gradients

River terraces in southern England are often unpaired, forming wide staircases to one side of the present river course. This is best seen in the lower Thames basin (downstream of London), along the rivers Thames, Medway and Lea (Bridgland et al., 2004; Bridgland, 2006; Howard et al., 2007) and to a lesser extent in the Upper Thames basin, upstream of the Goring Gap (Maddy et al., 1998; Lewis et al., 2001). In the Great Ouse basin between Newport Pagnell and Bedford, and in the Nene basin between Wellingborough and Wansford (Fig. 2) the

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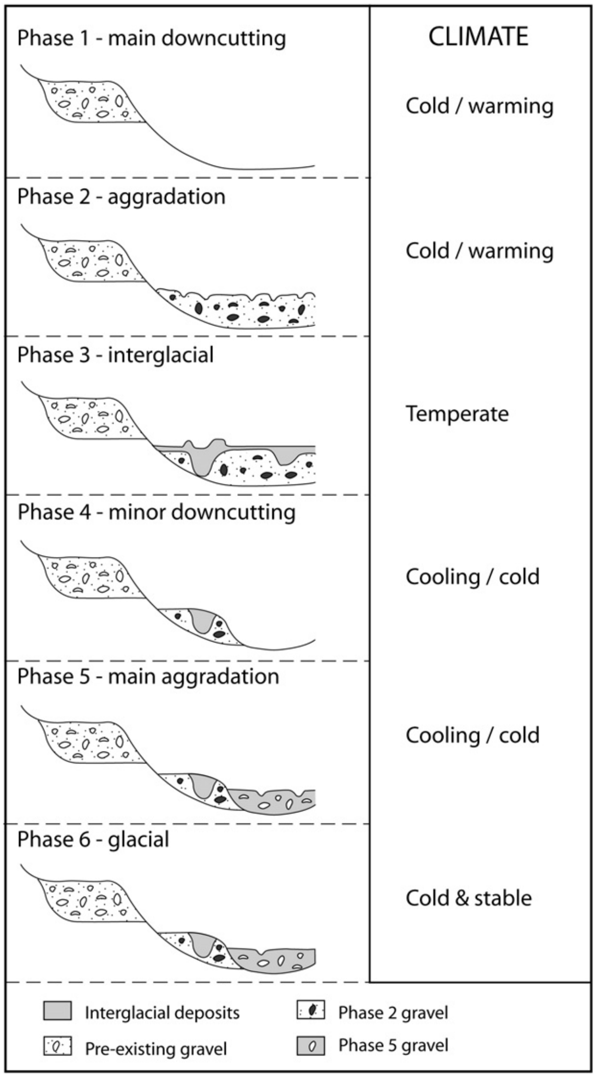

Figure 1. Model of terrace formation of the River Thames. Adapted from Bridgland (2006) and Maddy et al. (2001).

terraces regularly change sides in a constricted valley, but they are still mainly unpaired. However, on several rivers the lowest terrace, often called the Floodplain Terrace, is usually paired. The surface of this terrace is usually less than 2 m above the surface of the current floodplain, and the basal clay tends to continue at the same level below both the Floodplain Terrace and the current floodplain (Collins et al., 1996; Gao et al., 2007).

Each terrace consists of a terrace surface that is subaerially exposed and a bedrock bench that is buried by the terrace sediments (Fig. 3; Gibbard, 1985, p. 5; 'strath terrace' of Gibbard and Lewin, 2009, Fig. 5b). The terrace surface represents the remnants of a floodplain surface, and has often been later disturbed by post-depositional processes such as cryoturbation and soil formation while the edge near the valley side may be buried under solifluction deposits. The bench is broadly horizontal to gently undulating ('a pediment' of Castleden, 1977, 1980) and represents an erosion surface.

The terrace staircases occupy valleys whose cross-profile tends to be relatively flat and shallow. Channel and terrace gradients tend to be very low in relation to present-day peak discharges. Values of 0.14% and 0.15% have been obtained for the Kennet valley terraces (Cheetham, 1980), 0.11% for the ancestral Milton River (Belshaw et al., 2005), and 0.045% for the floor of the present Nene valley below Northampton. The Upper Thames between Cricklade and Eynsham has a gradient of 0.032%, and the lower Thames between Staines and Teddington has one of 0.023%.

Terrace stratigraphy

The stratigraphy of the terrace deposits found in the river valleys in southern England is shown in idealized form in Figure 3 (Gibbard, 1985).

Clay or chalk bedrock typically underlies the terrace deposits. Examples include the Jurassic Oxford and Lias clays, and the Eocene London Clay, all of which tend to be silty and thus frost-susceptible. Brecciation is common in the upper few meters of the bedrock, often becoming finer towards the top, where horizontal platy fragments commonly 5–20 mm wide and a few mm thick grade downward into larger fragments (Fig. 4). The upper layer is sometimes deformed into gentle folds. The bedrock is unconformably overlain by gravel.

The gravel deposit directly above the bedrock surface is, in the cases quoted, a basal lag of large clasts derived from the bedrock: for example, sandstone boulders ≤ 1 m median diameter derived from the London Clay (Collins et al., 1996); calcareous nodules from the Lias Clay (Smith, 1999); and septarian nodules from the Oxford Clay (Maddy et al., 1998). Fine sediments and organic materials—some of temperate affinities—are often associated with the lag deposits (Maddy et al., 1998; Smith, 1999).

The main gravel sheet, usually 2–5 m thick, is composed of a succession of thin sheets of sand and gravel interrupted by small cut-and-fill structures in multiple channels (Figs. 3 and 5). Less common are channels filled with massive gravels cut more deeply into the underlying gravels (Seddon and Holyoak, 1985; Maddy et al., 1998; Briant et al., 2005). Such channels occasionally extend into clay bedrock to form an irregular, narrow channel incision (Gibbard, 1985, p. 5) that contrasts with the flattish erosion surfaces which extend laterally for hundreds of meters (Castleden, 1980; Bridgland, 2006). Intraformational ice-wedge pseudomorphs—some multi-stage (Maddy et al., 1998; Seddon and Holyoak, 1985)—and involutions are common in the gravels (Fig. 5). Organic deposits and clastic silty sediments, which occur occasionally in pockets, lenses or channel fills between the layers of gravel, suggest that the whole aggradation accumulated sporadically over a long period of time.

Sheets of diamicton ('head deposits') may interdigitate with the gravel or overlie it, feathering out at some distance from the source hillslopes (Fig. 3; Gibbard, 1985, p. 5). Silty or clayey sediments ('brickearth') sometimes mantle the terrace staircase.

Requirements for a model of terrace formation

A model of terrace formation must explain the features above and meet some additional requirements. It is widely assumed that (1) terrace gravels are the remnants of former floodplain accumulations left on valley-side benches by incision into the valley floor, and (2) the area has experienced uplift (Maddy, 1997; Stemerdink et al., 2010).

The current model (Bridgland, 2006) incorporates these assumptions and provides a valuable stratigraphic scheme for understanding terrace deposits. It assumes that channel incision through the floodplain floor is caused by high-energy fluvial events and that valley incision is caused by channel migration. The processes of gravel movement and incision, however, are not considered explicitly and the timing of major incisions is loosely constrained to climate transitions. In order to investigate these issues, we first discuss the movement of gravel in low-gradient rivers.

Flow and gravel movement in low-gradient rivers

The drag and lift forces capable of moving gravel-sized particles along the river bed commence when river flow is close to 'critical' (Helley, 1969). Critical flow occurs when the velocity of flow downstream is equal to the velocity at which a gravity wave can travel upstream (see 'Froude number' in Chorley et al., 1984, p. 282). Under these conditions river flow is composed entirely of small, rapidly rotating eddies with axes parallel to the direction of flow. The eddies impart the considerable lift forces needed to entrain gravel.

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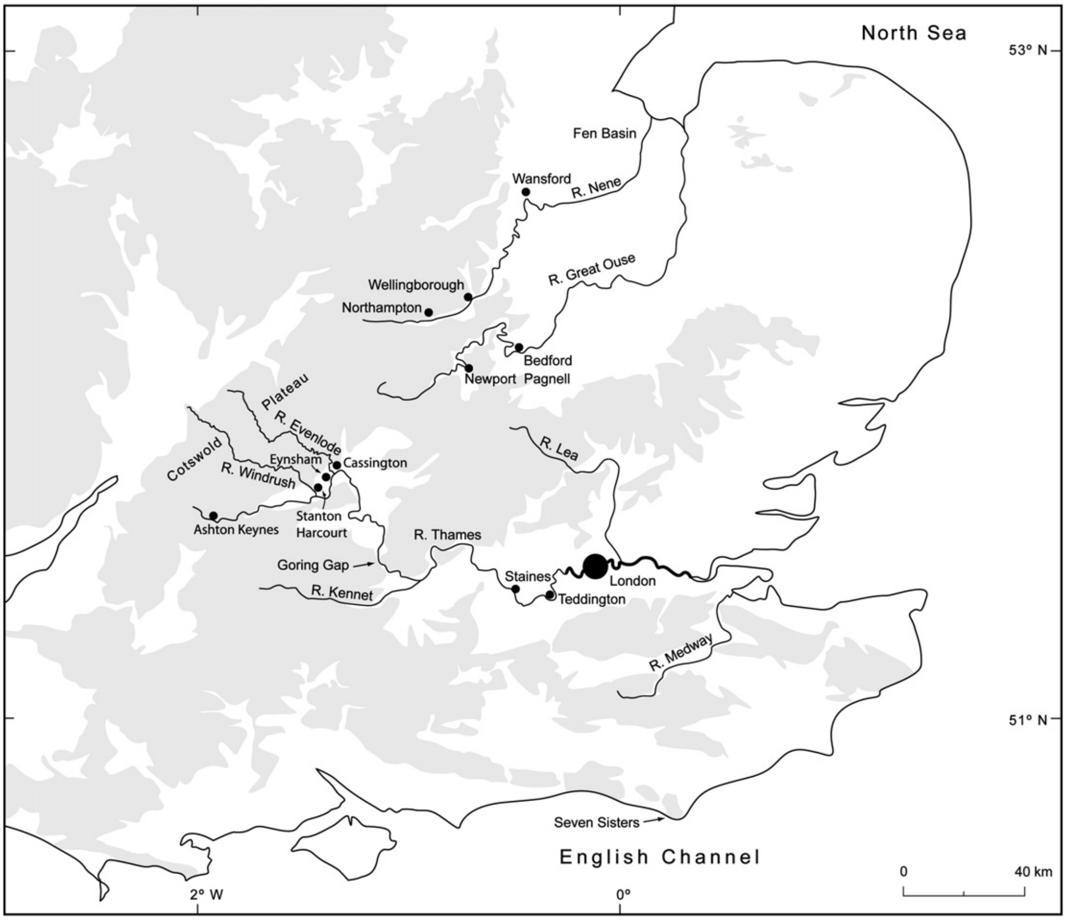

Figure 2. Map of southern England showing localities and rivers discussed in text. Shading indicates areas with elevations >100 m above sea level.

River flow capable of moving gravel in flattish wide valleys with very low gradients can occur, in theory, under two strongly contrasting scenarios.

Scenario 1. Rapidly fluctuating moderate discharges flowing over a wide irregular gravel surface (localised supercritical flow)

Moderate discharges are unable to remove gravel from the wide valley floor. From Manning's equation, slope contributes very little to the flow velocity in low-gradient rivers:

$$v = \frac{kR^{2/3}s^{1/2}}{n} \quad (1)$$

where v is the velocity ($m\ s^{-1}$), k is a coefficient (1 for the SI units), R is the hydraulic radius (mean depth (m) in wide channels), s is the channel slope ($m\ m^{-1}$) and n is the Manning roughness coefficient (a general measure of channel resistance).

The necessary increase in the discharge needed to evacuate floodwater results only from an increase in hydraulic radius (depth), which automatically causes an increase in velocity. The change in the velocity from the bed to the surface in such flow is very gentle, failing to produce sufficient lift and drag at the bed to move anything larger than fine gravel.

Material supplied by slope processes on the valley sides can produce an uneven surface on the valley floor and stream bed. Once initiated, the uneven stream bed allows the development of highly localised and erosive supercritical flows (Froude number>1) that cut shallow scour hollows in the material, winnowing out the fines and leaving behind the coarser particles. The enhanced gradient migrates upstream, allowing the scours to fill, forming cut-and-fill structures. Channels fill with fining-upward beds as the flow rapidly wanes, aiding the abandonment of

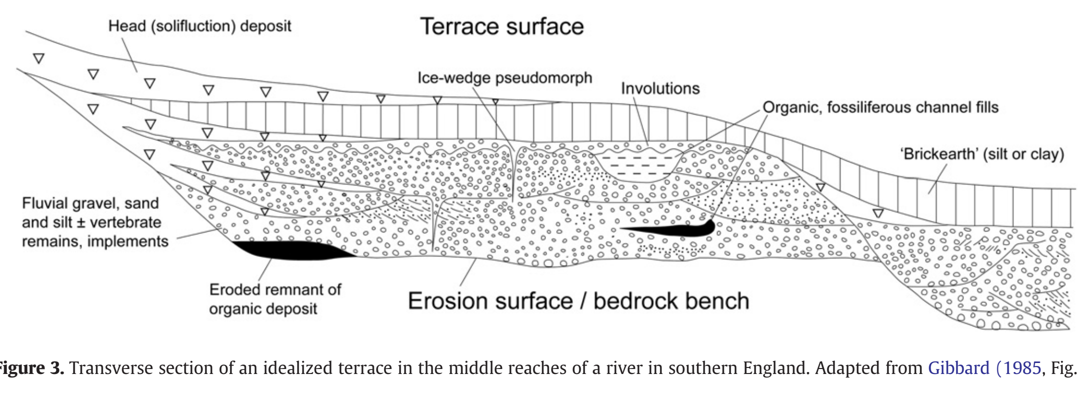

Figure 3. Transverse section of an idealized terrace in the middle reaches of a river in southern England. Adapted from Gibbard (1985, Fig. 3).

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Figure 4. Brecciated Oxford Clay bedrock, exposed in vertical section ~0.4–0.6 m below the base of fluvial gravels at Cleveland Farm Quarry, Ashton Keynes, in the Upper Thames basin; 2 June 1995. 12-cm-long pen for scale.

existing channels in the next flood event and resulting in the spread of thin sheets of gravel into topographic lows. While fine materials are washed away, the larger material moves infrequently, covering very short distances. Over time the process leads to the development of broad aggradational braidplains. Organic material may accumulate in the topographic lows and abandoned channels. The resulting sedimentary architecture is composed of thin beds of sands and gravels interrupted by small channels with cut-and-fill structures, while organic beds may occur randomly in the sequence. Such accumulations often total 3 to 5 m in thickness. The resulting architecture is similar to the finer sequences of Miall's Donjek model for distal braided gravels (Miall, 1977, 1978), but the scarcity of thick deposits of massive clast-supported gravels in the terrace deposits of the present study precludes an exact match with the proposed model.

The processes in this scenario are capable of producing the aggradational sequences of the terraces of the Upper and Lower Thames (Fig. 5) but they cannot account for the broad and flattish erosion surfaces at the base of the deposits discussed in this paper.

Scenario 2. Extremely high sustained discharges that fill the whole width of the valley (widespread supercritical flow)

On broad floodplains, shallow floods will be subcritical (Froude number < 1) and flow will be tranquil. But as flow depth increases, the downstream velocity of the flow increases more rapidly than the rate at which gravity waves (surface ripples) can propagate upstream, causing the Froude number to increase slowly with depth:

$$Fr = \frac{\overline{v}}{\sqrt{(gd)}} \qquad (2)$$

Figure 5. Ice-wedge pseudomorph, exposed in vertical section through fluvial gravel and sand of the Upper Thames basin at Cassington; 22 November 1996. Section is ~3 m high.

where Fr is the Froude number, $\overline{v}$ is the mean flow velocity (see Eq. (1)), g is the acceleration due to gravity and d is the depth of flow.

Should the flow approach critical (Froude number = 1), then bed material would be mobilised across the whole bed, allowing incision into bedrock and the production of a sedimentary architecture dominated by massive clast-supported gravels. On a gradient of 0.3% the mobilisation of bedload would occur once the flow depth exceeded about 5 m—probably the order of magnitude of flow depth proposed by Stemerdink et al. (2010). However, on the very low gradients of the Upper and Lower Thames (0.03%) the water depth would need to exceed ~100 m to begin to entrain bedload, while even greater sustained depths would be needed to start to erode channels in the bedrock.

This scenario requires such impossibly high discharges that it can be discounted as a mechanism for terrace formation in the Thames Basin.

Climatic conditions

We now consider how gravel movement and incision might occur by purely fluvial processes in three types of climatic conditions relevant to the middle and late Pleistocene of southern England, or transitions between them.

1. Humid interglacial conditions

Due to the hydrological conditions imposed by dense vegetation cover and soil development, rivers adopt a single meandering channel mode with cohesive banks developed in overbank sediments. Any increase in discharge causes the flood to spill over the floodplain and inhibits any effective increase in either the flow velocity or the depth of water in the channel. Thus, high discharges in low-gradient rivers cause extensive flooding with overbank sedimentation but little or no erosion of the bed. For example, in June 2007 exceptionally heavy rainfall in the southern Midlands caused extensive flooding in the valleys of the upper Thames, Windrush, Evenlode, Nene and Great Ouse, but with negligible effect on the channel beds. As human activity has led to increased and more peaky runoff than would have occurred in the past, it is unlikely that flooding in previous interglacials was capable of initiating valley incision. Rather, as a result of overbank flooding, the normal long-term fate of low-gradient river floodplains is to be submerged in successive loads of fine-grained overbank sediment as isostatic depression and sediment compaction lower the crust, as is occurring at New Orleans, on the Mississippi Delta. This general condition of sediment erosion, transport and deposition—as schematized by Church (2002) and shown in Figure 6—results in long-term storage of fine-grained sediments in floodplain valleys. In conclusion, we discount significant gravel movement and incision during humid interglacial conditions, in agreement with Gibbard and Lewin (2002, 2009).

2. Warm periglacial conditions

Very peaky discharges and a plentiful supply of sediment characterize warm periglacial conditions with discontinuous permafrost or deep seasonal frost. Peaky discharges arise from rapid spring snowmelt, the breaking of ice dams and summer rainstorms. The sedimentary architecture attributed to these periods (e.g., Maddy et al., 1998) indicates that aggradation occurred under the local conditions for supercritical flow outlined in scenario 1 above, involving mainly fluvial processes and forming cut-and-fill structures. Such processes may have destroyed or reworked fine overbank sediments formed in temperate periods, but cannot have eroded the broad and flattish bedrock surface that underlies the braidplain deposits. Thus, we discount valley incision during warm periglacial conditions.

3. Cold permafrost conditions

As discussed below, cold (continuous) permafrost conditions in the Pleistocene of southern England were generally associated with

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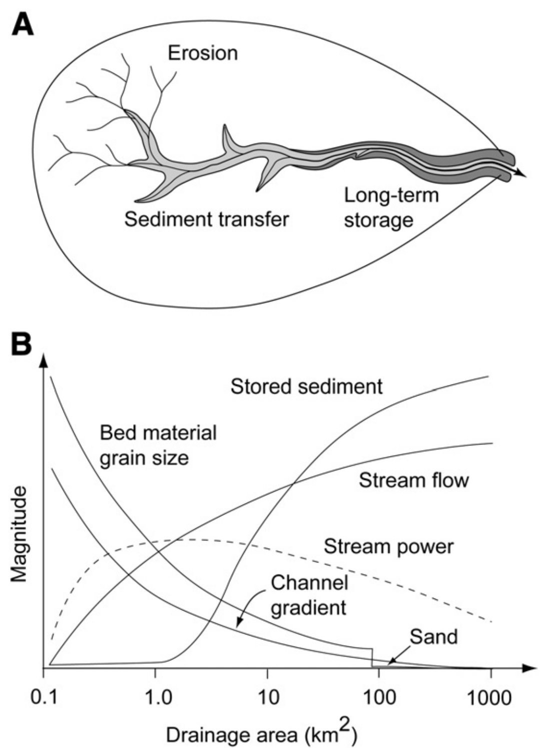

Figure 6. Relationship between stream discharge, gradient, and caliber of gravel moved. Adapted from Church (2002, Figs. 1a, b).

arid environmental conditions (Briant et al., 2004, 2005), with the result that fluvial processes were generally inactive at such times. However, the occasional channel incision into the basal clays associated with multi-stage ice-wedge pseudomorphs (Seddon and Holyoak, 1985) arises from local thawing of deeply frozen gravels by streamflow. A major contribution to a braided habit is the normal lack of cohesion in coarse clastic material. Frozen gravels, however, allow the development of a potentially wandering single-thread channel with near-vertical cohesive banks, making the flow velocity more critical compared with channels in non-cohesive gravels. In effect, the frozen banks transform the channel into a steep-sided flume. The lifespan and attainable depth of such single thread channels would be limited by the payoff between thawing giving increased runoff but causing the collapse of the banks and the obstruction of the channel. Evidence for such lateral migration of channels is found in the occasional truncated pseudomorphs and bank collapse structures (Seddon and Holyoak, 1985; Maddy et al., 1998). Any such incisions into the basal clay would form a narrow sharply bounded channel with an irregular floor rather than the very broad planar surface that underlies the terrace deposits of the Upper and Lower Thames. Such limited fluvial processes within frozen gravels are incapable of effective incision and the production of a wide basal surface in bedrock that has been protected from brecciation in the contemporary period by the overlying gravels.

4. Climate transitions

The prevailing terrace model for southern England proposes that substantial gravel movement and channel incision occurred during climate transitions between periglacial and interglacial conditions (Fig. 1; Bridgland, 2006), in line with the broader conceptual model proposed by Vandenberghe (2008) of fluvial incision at cold–warm–cold transitions in lowland regions. But the fundamental problem remains that the channel gradients reported above from southern England are insufficient to remove coarse gravel from the wide valley floor where the fluvial system is operating within the constraints of scenario 1. To drive gravel removal and valley incision, regional uplift of the landscape would have to cause substantial tilting and channel steepening. However, Westaway et al.'s (2002) analysis of uplift of the Thames basin does not suggest the development throughout the system of a gradient necessary to move coarse gravel along a riverbed. Sea-level change may have affected the development of terraces in the lowest part of the Thames system, but there is no evidence for the upstream progression of knickpoints into areas such as the Upper Thames basin (Stemerdink et al., 2010). Neither is there any evidence on the valley floor of the Upper Thames of the massive clast-supported gravel bar deposits that would have been left by the waning stages of any fluvial event large enough to have penetrated the floodplain gravels to incise a new valley.

In conclusion, we consider it unlikely that substantial movement of coarse gravel and valley incision took place by purely fluvial processes in any of the four situations outlined above. Instead, it is essential to consider periglacial conditions and processes, and their interaction with fluvial systems. Specifically, to explain the origin of the brecciated surfaces that underlie terrace gravels we attribute incision and planation to cold and arid permafrost conditions that favored the processes of fracture and thermal erosion directly to the bedrock, and suppressed the supply of coarse sediment from the valley sides by solifluction.

Periglacial conditions and processes

Bedrock fracture by ice segregation

The pervasive fractures (brecciation) towards the top of the bedrock beneath terrace deposits suggest that segregated ice was originally an important component of bedrock close to the contemporary ground surface. The finer brecciation towards the top of the bedrock suggests a higher concentration of segregated ice here than at greater depth. This vertical distribution of ground ice is commonly observed in contemporary cold regions, both in seasonally frozen ground and in the ice-rich upper ~0.5–1 m of permafrost developed in frost-susceptible Arctic soils and bedrocks (Fig. 7; French et al., 1986). The ice-rich layer in regions of cold (continuous) permafrost represents the bedrock 'ice-rind' identified by Büdel (1982) and the 'transient layer' in soils discussed by Shur et al. (2005).

Climate cooling in the Pliocene probably led first to seasonal freezing and then, in the early Pleistocene glacial periods, to perennial freezing of the low-relief terrains shaped by Tertiary erosion. Wherever moist, frost-susceptible materials froze, the process of ice segregation occurred within the upper meters of ground, causing fissuring of soils, fracturing of bedrock and heaving of the ground surface (Murton, 1996; Murton et al., 2006). Landscape evolution can be sensitive to freezing conditions because ice segregation and bedrock fracture in permafrost create an ice-rich regolith that can be transported downslope efficiently by solifluction, and because ice-cemented substrates are vulnerable to fluvial incision and planation.

Solifluction

Fractured bedrocks beneath hillslopes in southern England have been widely reworked by Pleistocene periglacial slope processes to form diamictons locally known as 'head deposits' (Ballantyne and Harris, 1994). The head deposits often interdigitate with fluvial gravels (Fig. 3), the former typically flanking the valley sides and the latter occupying valley centers (Green and McGregor, 1980; Shakesby and Stephens, 1984; Gibbard, 1985, pp. 4–6). Many head deposits have been attributed to solifluction.

Measured rates of downslope sediment transport are substantially greater in relatively wet continuous permafrost regions, such as those in the Arctic, than in either warm (discontinuous or sporadic) permafrost or in non-permafrost regions (Matsuoka, 2001). Solifluction is favored by melt of large volumes of segregated ice, particularly

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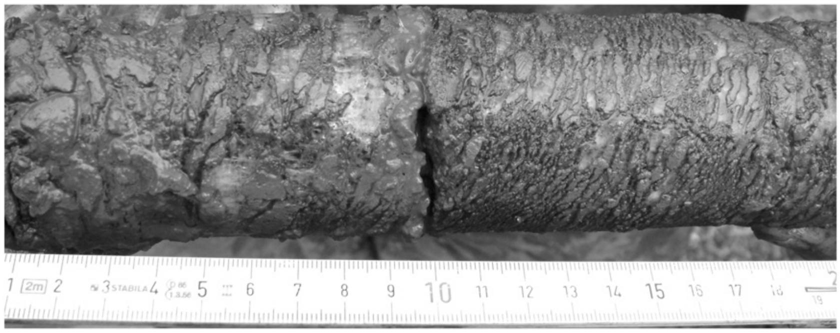

Figure 7. Core extracted from ice-rich layer in silt–clay in upper 1 m of permafrost, Endalen, Spitsbergen. Segregated ice is white or dark gray, and silt–clay is mid-gray; August 2005 (Courtesy of Charles Harris); 1-cm units marked on ruler.

by complete thaw of the ice-rich layer at the top of the permafrost, for example during periods of climate warming which cause permafrost degradation. Conversely, ice segregation and solifluction are limited under arid conditions. In short, sediment delivery to valleys is favored by moist and cold permafrost conditions, and restricted by arid ones.

Thermal erosion and valley incision by rivers

Thermal erosion occurs where flowing water melts ground ice by the combined effects of heat conduction and convection, and then mechanically erodes newly released sediment or rock fragments. Thermal erosion along riverbanks through ice-rich unconsolidated sediments causes undercutting and rapid bank retreat (Church and Miles, 1982; Lawson, 1983). Undercutting by currents excavates a horizontal cleft (thermo-erosional niche) that may extend 10 m or more laterally into the bank, at about water level. Above the niche, the undermined permafrost episodically collapses in large blocks, often along ice wedges. Retreat rates show high spatial and interannual variability. For example, on the Lena River, northern Siberia, retreat reaches 19–24 m yr^{-1}, or even, exceptionally, 40 m yr^{-1} (Are, 1983). On the Colville River, northern Alaska, the long-term retreat rates rarely exceed 3 m yr^{-1}, although block collapse can generate an almost instantaneous retreat as much as 12 m, protecting the bank from further retreat for periods of up to a few years (Walker et al., 1987). Numerical analysis, experimental simulation and field observations of fluvial thermal erosion suggest that the rate of erosion is very sensitive to increases in water temperature and less sensitive to increases in discharge (Costard et al., 2003, 2007).

Given the likely operation of these periglacial processes during periglacial or permafrost conditions in Pleistocene southern England, we now consider the probable timing of valley incision.

Timing of valley incision

The attribution of major valley incision phases to periods of periglacial-to-interglacial warming is problematic. It arose from the stratigraphic position of interglacial deposits that lie at the base of individual terrace sediment sequences (Fig. 3). This stratigraphy, however, indicates that valley incision *preceded* interglacial valley filling; it does not distinguish whether incision occurred during the climate warming or during the prior periglacial conditions. For a number of reasons, the latter hypothesis is more reasonable than the former.

First, there was insufficient time for both the main incision period and for substantial aggradation to occur during climate warming from periglacial to interglacial conditions, as proposed for phases 1 and 2 of Bridgland's model (Fig. 1). The Greenland ice-core record indicates that glacial-to-interglacial climate warming here was in the order of a few years to several decades (Steffensen et al., 2008). By implication, North Atlantic regions downwind of Greenland would also have warmed rapidly, much too rapidly for a large river like the Thames to change abruptly from major incision to aggradation, with both coinciding with the warming phase.

Second, stratigraphic studies in Alaska, northern Canada and Siberia indicate that global warming during the last glacial-to-interglacial transition triggered a major phase of thermokarst activity and associated mass wasting where ice-rich permafrost degraded (Murton, 2009). In turn, the mass wasting (e.g., debris flows and solifluction) resulted in deposition of abundant sediment in lowlands (Murton, 2001). Although the ice content of Pleistocene permafrost in the UK was probably much lower than that in many Arctic and sub-Arctic lowlands, limited amounts of excess ice are still inferred from the widespread brecciation of near-surface bedrocks (Murton, 1996). Degradation of such permafrost during major phases of climate warming must have enhanced mass wasting and sediment supply to valley bottoms adjacent to icy hillslopes (cf. Bridgland and Westaway, 2008, p. 293). Such deposition equates with aggradational phase 2 (Fig. 1).

Third, the large dimensions of the valley floors suggest that a reasonably long period of time was necessary for their formation (perhaps 10^2–10^3 yr).

If incision did not take place during climate warming and permafrost thaw, then it must have occurred during the previous periglacial conditions. We discount widespread incision during warm and moist periglacial conditions characteristic of Marine Isotope Stage (MIS) 3 because the sediment stratigraphy and geochronology indicate that this was generally a time of fluvial aggradation in southern England and the northwest European lowlands (e.g., Van Huissteden et al., 2001). A third hypothesis is that incision coincided with very cold and arid permafrost conditions. We now set out conceptually a process-based model for valley incision and terrace formation that is consistent with this hypothesis.

A conceptual model of valley incision and terrace formation

Constraints

The model has a number of constraints:

1. It applies only for those terraces with sedimentary architecture of scenario 1.
2. The presence of a lag gravel of local origin that commonly contains temperate organic material suggests that the formation of the erosion surface and the accumulation of the lag gravel above it are temporally separate events.
3. The low longitudinal gradients of the valleys, combined with moderate peaky discharges, permit only winnowing by localised instability of the gravel introduced into the valleys mainly from hillslopes (scenario 1) and not the wholesale movement of gravel in river channels. Thus only particles finer than fine gravel move far

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downvalley. As a result, the gravel tends to accumulate as braidplains within the valleys, bearing the sedimentary features characteristic of scenario 1 (e.g., Maddy et al., 1998; Briant et al., 2004). Hence, the valleys provide accommodation space for the long-term storage of gravel.

4. The smooth clay surface manifest by the surface of the basal clay has a very low flow resistance (Manning $n \approx 0.01$; similar to concrete) compared to a gravel bed ($n \approx 0.035$; Selby, 1985, pp. 240–242). This favors high-velocity shallow flow (Wang et al., 2009) capable of rolling clasts up to cobble size (including bones and human artefacts) until they come to rest in some hydrodynamically sheltered position. By contrast, large boulders eroded from the bedrock (Collins et al., 1996; Maddy et al., 1998) remain in situ and become the focus for a discontinuous lag deposit.
5. Interstitial ice renders granular material cohesive. In this way, frozen sand and gravel on the pre-existing floodplain support steep or even overhanging banks subject to basal scour, as is common in contemporary Arctic rivers in permafrost regions (Church and Miles, 1982).

Model

The model envisages shallow incision and wide planation. During times of incision, the channel bed and lower part of the banks consisted of smooth clay or limestone, and the higher parts ice-cemented gravel of the former floodplain. The cohesion imparted by ice within the clay and interstitial ice in the gravel favored development of a single thread channel. Within the channel, water sometimes flowed at high velocity, thermally eroding the ice-rich brecciated bedrock. The geometry of this icy layer—a widespread horizon ~0.5–1 m thick (Büdel, 1982; Shur et al., 2005)—favored erosion that was shallow but laterally extensive: in effect, fluvial planation (Castleden, 1977, 1980). Planation frequently occurred through cutting of a thermo-erosional niche under the river bank. Over time, niche-cutting and channel migration planated a wide surface across the clay. As the channel was eroded, myriad fine clay or limestone fragments released from the bed and banks were transported downstream and out of the basin, along with limited amounts of fine gravel collapsing from undercut banks and sediment dislodged by riverbank freeze–thaw processes (Yumoto et al., 2006).

In winter, flow slowed or ceased in the shallow rivers, allowing ice segregation to brecciate underlying and adjacent bedrock. During low-stage winter conditions, rivers shallower than ~2 m (the likely maximum thickness of winter river ice) froze to the bottom, and segregated ice formed in the saturated river bed, as occurs in fine sediments beneath shallow Arctic lakes (Burn, 1990). Ice segregation was favored beneath river beds because these were the wettest elements of the otherwise dry landscape. This is consistent with (1) deeper brecciation beneath many chalkland valleys relative to adjacent interfluves, as seen in coastal cliffs through the Seven Sisters of Sussex (Fig. 2); and (2) ice segregation preferentially fracturing the wetter rocks used in physical modelling experiments (Murton et al., 2006). On dry parts of the floodplain, some disaggregated silt and clay was locally redistributed by aeolian activity. Loess accumulation started in many sites well before outwash plains from Devensian glaciers were an available source (Rose et al., 2000; Kasse et al., 2007). Fine sediments washed down the rivers and deposited in the emergent southern North Sea basin and English Channel palaeovalleys would also have been subject to deflation, contributing silt to the loess of southern England and northern France (Antoine et al., 2003).

Incision commenced when aridity suppressed the sediment supply by solifluction from hillslopes onto valley floors. Thermal erosion was initiated where flowing water came into contact with ice-rich brecciated bedrock close to the surface of the contemporary floodplain. This often occurred at the edge of the floodplain where water became concentrated from snow melting on the adjacent unterraced hillslope.

Incision ceased when increased soil moisture promoted ice segregation and solifluction on hillslopes. This enhanced the transport of coarse sediment into the river valleys (see layers of soliflucted sediment in Fig. 3), clogging them with sediment of varied grain size and so replacing incision with aggradation. Nival floods swelled by increased snow in the catchments winnowed away the fine sediments, leaving a thick gravel deposit overlying any previous deposits, as Van Huissteden et al. (2001) discussed for MIS 3. Hence, the valley floor became buried by gravel.

The accumulation of gravel protected the bedrock and preserved the terrace staircase. The bedrock beneath the old valley-floor gravels >2–3 m thick—now new terraces—was protected from significant ice segregation in the ice-rich layer at the top of permafrost because the active layer and near-surface permafrost were generally confined to the gravel. In other words, the underlying clay was below the depth favoring build-up of abundant segregated ice in the upper layer of permafrost. A useful analogy is the laying of thick gravel pads, beneath Arctic roads and railways, to serve as stable, non-frost-susceptible substrates to contain the depth of seasonal freezing and thawing (Ferrians et al., 1969). In this way, the terrace gravels rendered the clay resistant to subsequent cycles of incision (cf. Clarke and Fisher, 1983). The excellent preservation of primary sedimentary structures in the gravel discounts the possibility of significant post-depositional disturbance by frost heave or thaw settlement in the underlying clay. Lewis et al. (2001) and Bridgland et al. (2001) cited the deformation of clays under gravels <2 m thick into diapirs and diapiric ridges, but there appears to have been a limited loss of material as generally the sedimentary architecture is well preserved. The protected clay at the edge of the newly formed terrace would become vulnerable to undercutting, but the gravel that would be delivered to the valley floor would delay further incision on that side of the valley, leading to preservation of the material and the staircase effect.

The terraces above the active floodplain lost most of their surface drainage to the newly incised valley floor, accounting for the absence of active channels in the terraces above the floodplain that was noted by Briant et al. (2004).

Summary

In summary, the model challenges the prevailing view of incision during climate transitions. It attributes incision and planation to very cold and arid permafrost conditions, when rivers had limited discharges and hillslopes supplied limited volumes of stony debris into valley bottoms. Where the floors and lower banks of rivers exposed brecciated bedrocks the wetted perimeter was smoother than that of gravel-bed streams, resulting in relatively high flow velocity in spite of the low channel gradient (see Eq. (1)). The rivers were therefore able to remove disaggregated bedrock fragments from the river basins, roll some large clasts of gravel downstream and—as the channels migrated across the floodplain—thermally erode a broad low-gradient erosion surface. When periglacial conditions became moister, enhanced ice segregation on the hillslopes increased the transport of coarse sediment by solifluction to the valley floor. This clogged the river channels with sediment, increasing the roughness of the channel floor, because incompetent tractive forces on the low gradient caused them to aggrade the gravel and winnow away most of the silt and clay, forming shallow braidplains on the broad valley floor. The gravel later served to protect the underlying bedrock surface from erosion during subsequent cycles of incision and planation.

Overall, during the long periods of Pleistocene periglacial conditions, the rivers alternated between incision of bedrock and aggradation of gravel. This bimodal behaviour allows us to explore how the landscape developed through periglacial–interglacial cycles.

Landscape development through periglacial–interglacial cycles

Development of the fluvial landscape of southern England through periglacial–interglacial cycles can be broadly related to three environmental regimes separated by transitions resulting from climate warming and cooling.

Regime 1. Cold (continuous) permafrost and arid conditions

Regime 1 resulted in valley incision when cold (continuous) permafrost underlay most of southeast England. As discussed above, very cold and arid conditions promoted fluvial incision because they restricted sediment supply from hillslopes into valley bottoms and thus allowed limited river discharges to incise bedrock-bottomed channels with ice-cemented banks. This regime is best known from the last glacial maximum (LGM), the most recent period (26.5 to 19 cal ka BP) when global ice sheets reached their maximum integrated volume (Clark et al., 2009). Continuous permafrost developed during the coldest phase (~23–19 cal ka BP) of the LGM in western Europe (Vandenberghe et al., 2004), but it is also known from parts of MIS 4, 6 and 12 (Rose et al., 1985; Worsley, 1987; Huijzer and Vandenberghe, 1998) and must have occurred during other cold periods. Mean temperatures of the coldest months of the LGM in western Europe have been estimated as between −25°C and −20°C, and mean annual temperatures about −9°C to −8°C; winter–summer temperature ranges were probably ~28°C to 33°C, indicating substantial continentality (Vandenberghe et al., 2004, Fig. 4). At this time thermal contraction cracking and ice-wedge formation in central and southern England were widespread (Ballantyne and Harris, 1994).

Arid conditions associated with continuous permafrost during the coldest time of the LGM are inferred from the widespread loess and coversands in the North Sea region, which indicate extensive aeolian activity (Koster, 2005). Aridity resulted in part from an extended sea-ice cover over the North Atlantic Ocean. The ice cover promoted surface cooling and development of a relatively high surface pressure, potentially blocking storms travelling towards Europe. In addition, the ice cover reduced the availability of moisture (latent heat release) needed for growth of mid-latitude storms (Renssen et al., 2007). Aridity also resulted from anticyclonic pressure systems developing above the Scandinavian and British ice sheets, restricting precipitation above the ice sheets and the permafrost terrain to their south (Briant et al., 2004). Geological evidence and climatic modelling both point to a dominant LGM winter wind regime that was broadly westerly and stronger than that of today, relating to the eastward relocation of the Icelandic Low and an enhanced pressure gradient over northwest Europe (Renssen et al., 2007). Consistent with such aridity, fluvial sediments dated to the LGM are rare in Britain, suggesting that normal river activity was very limited, "with low run-off rates and possibly significantly truncated flood seasons, few active river channels and probably extremely limited vegetation growth" (Briant et al., 2004, p. 493).

Regime 1 corresponds to phases 1 and 4 of the Bridgland model (Fig. 1), but the incision occurs to the side of the old floodplain rather than through it. The old floodplain experiences phase 6 of the Bridgland model.

Regime 2. Warm (discontinuous) permafrost and moist conditions

Regime 2 resulted in major gravel aggradation associated with warm (discontinuous) permafrost and moist conditions. The regime was dominant during MIS 3 (~60–28 ka) and the Younger Dryas. During this regime rivers in southern England were mainly of braided type and deposited gravel as a result of the supply of abundant coarse sediment from mass wasting on adjacent hillslopes (Van Huissteden et al., 2001). A highly variable nival discharge regime favored reworking of solifluction deposits during snowmelt floods, winnowing out the fine-grained sediment and leaving behind the gravel. Vegetation on or near to floodplains was treeless, herb-rich grassland. Regime 2 corresponds to phases 2 and 5 of the Bridgland model (Fig. 1).

Regime 3. Temperate and moist (interglacial) conditions

Regime 3 characterized interglacial periods, and led to the deposition of fine-grained overbank sediments on the floodplains of meandering rivers, and slightly coarser material in the channels. Incision of single-thread channels into bedrock is recorded at several sites where interglacial deposits are preserved below subsequent gravel deposits (Shotton et al., 1993; Bridgland et al., 2001; Keen et al., 2006). Interglacials that follow the deposition of gravels on the new valley floor (as in the Holocene following the Younger Dryas) generally had limited impact on the underlying bedrock and left little evidence. Regime 3 corresponds with phase 3 of the Bridgland model.

Climate warming

Climate warming favored gravel aggradation in river valleys when an ice-rich layer at the top of permafrost thawed beneath adjacent hillslopes. The warming caused active-layer deepening, thaw of the ice-rich layer and therefore enhanced sediment transport by solifluction to valley floors, as discussed above. Active-layer deepening and gravel aggradation probably occurred rapidly during abrupt climate warming (associated with ice-age terminations and interstadials during MIS 4 and 2; Dansgaard et al., 1993), but were slower during periods of gradual warming, as during Greenland stadial 2b (20.9 to ~17.4 ka; Lowe et al., 2008). In the absence of the ice-rich layer, the effects of warming on slope processes and river behaviour remain to be determined.

Climate cooling

Climate cooling had differing effects on river behaviour, according to the nature of cooling and the pre-existing conditions. Cooling from interglacial to warm periglacial conditions initiated solifluction, transporting substantial amounts of colluvial materials from hillslopes to valleys, and leading to valley aggradation (cf. Green and McGregor, 1980, stage ii; Bellamy, 1995). The increase in colluvial transport, we suggest, more than compensated for the hypothesized effects of lingering vegetation cover in reducing sediment supply to rivers and thus favoring incision (Vandenberghe, 1993, 1995, 2003). Vegetation would have limited solifluction through binding and thermal insulating effects (Matsuoka, 2001) but it would have not prevented it, otherwise landforms such as turf-banked solifluction lobes could never develop. Thus sediment transport from hillslopes to valleys increased substantially compared with transport mainly by soil creep in temperate interglacial conditions. In contrast, climate cooling from cold and moist conditions of regime 2 to very cold and dry conditions of regime 1 may have reduced ice segregation and solifluction, and thus caused valley aggradation to slow or cease.

Discussion

The model presented above is idealized and oversimplified. The environmental regimes are illustrative of characteristic environmental conditions rather than exhaustive. The precise effects of glacial–interglacial changes as opposed to stadial–interstadial changes remain to be fully explored, as the spacing between intervals of climatic change influenced the development of the ice-rich layer at the top of permafrost and therefore the vulnerability of the landscape to geomorphic change.

The model favors lateral planation by thermal erosion rather than by piping of clay. Piping of clay fragments upward through gravel interstices—as occurs around some bridge foundations—would allow removal of clay from beneath a gravel sheet. But we discount this process on account of the large lateral and downstream extent of the erosion surfaces.

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Our model supports Gibbard and Lewin's (2009) model in restricting the timing of valley incision to periglacial conditions rather than climate transitions or interglacial periods. It disagrees, however, over some of the processes. The Gibbard and Lewin model envisages that cold-climate fluvial processes lead to deep coupled incision and planation, associated with a thin (few to several meters thick) 'working depth' of gravels—from scour depth to highest flood level—at any one time. The gravel represents transported but temporarily stored sediment. Repeated sweeps of laterally mobile rivers across floodplains, with shallow incision each time in scour pools at channel bends and in channel confluences, might plane a broad bedrock erosion surface with almost immediate 'lag' deposition above it. This could result in basal bedrock erosion by rivers on one part of the floodplain, and then thin deposition as the channel shifts to another, all taking place over the same time period. Although this process might apply to conditions of high discharges in humid periglacial times, we think it is unlikely to characterize low-flow arid conditions. More fundamentally, coupled incision and planation does not overcome the difficulties, discussed above, of moving gravel-sized clasts on very low gradients across rough channel bottoms, irrespective of the climatic conditions. Lastly, at Cassington, on the Upper Thames (Fig. 2), the dating suggests that the main gravel sheet was deposited tens of thousands of years after the basal fluvial deposits (Maddy et al., 1998).

Our model develops partly from Büdel's (1982) concept of 'excessive valley cutting'. Büdel recognized on Svalbard the importance of ground ice in fractured bedrock permafrost as preparatory to fluvial planation and incision by steep channels of gravel-bed braided streams through uplifted etchplains. We estimate that Stauffer Brook, Barents Island—where Büdel (1982, Fig. 16) exemplified his valley-cutting concept—has a gradient of ~3%. Our model, however, differs from Büdel's in being limited to frost-susceptible bedrocks in non-glaciated catchments with rivers of low valley gradients (i.e., one to two orders-of-magnitude lower than that of the Stauffer Brook). Thus, our model applies to a smaller geographical area than envisaged in Büdel's subpolar zone of excessive valley cutting (Büdel, 1982, Fig. 13).

Our model is developed specifically for valleys underlain by frost-susceptible bedrock covered by gravelly terrace staircases in southern England. It is exemplified by those in the Upper Thames basin and in the Thames downstream of London, where incision of bedrock and deposition of gravel each requires distinctly different geomorphic activity. The English deposits tends to be coarser grained than the dominantly sandy and silty terrace deposits of the northwest European lowlands (Van Huissteden et al., 2001. Thus, the model's applicability to these finer deposits and their specific substrates is not known.

The Middle Thames terraces are excluded from our model because their sedimentary architecture differs from that of the Upper and Lower Thames. In many cases the basal lag is missing and thick, massive clast-supported gravels lie directly on a planar clay base that shows evidence of channelling. The flow conditions necessary to produce this would be more appropriately described as scenario 2 above and will be discussed in a forthcoming paper.

Our model is appropriate only for the formation of aggradational terraces on frost-susceptible bedrock in permafrost areas experiencing uplift. Along the southern edge of the Fen Basin, late middle and late Pleistocene deposits are superimposed (Langford et al., 2007), implying either a lack of uplift in that area or that overdeepening by the Anglian ice sheet affected relative base levels. In areas subject to proximate glacial activity aggradational terraces can be formed by the disruption of drainage by the ice sheet itself or by the effects of glacial loading and unloading, without a preceding period of valley incision.

Conclusions

We have attempted to show that the existing model of terrace formation—which attributes valley incision to climatic transitions—is difficult to accept in the light of modern understanding of periglacial processes and gravel transport by rivers in low-gradient valleys. Instead, we suggest that the key driver of incision was the limited moisture supply for ice segregation and fluvial processes during times of arid and cold permafrost conditions.

Our model supports Te Punga's (1957, p. 410) conclusion that "Southern England is at present a typical relict periglacial landscape." Te Punga recognized the fundamental importance of periglacial mass wasting, especially solifluction, on hillslope morphology. We contend that thermal erosion by rivers during times of diminished solifluction was fundamental to the incision and planation of river valleys. The role of periglacial processes in the drainage development of southern England, as far back as the early Pleistocene, is clearly a promising area of research (Belshaw et al., 2006; Belshaw, 2007). The hypotheses used to construct our conceptual model now require rigorous testing in order to forge a robust process model and to determine if the model has wider applicability to Pleistocene river terraces in other regions of past permafrost.

The real challenge is to determine the age and length of formation of the valley floors, i.e. the timing of incision and planation events. A more tractable objective is to test the hypothesis that terraces in southern England occur in the valleys of low-gradient rivers experiencing uplift that flow over frost-susceptible materials, whereas valleys where these materials are absent experience little or no terrace formation.

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