Relation between alternations of uplift and subsidence revealed by Late Cenozoic fluvial sequences and physical properties of the continental crust

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Reversals in vertical crustal motion, alternations between uplift and subsidence over timescales of hundreds of thousands of years or more, have been identified in Late Cenozoic fluvial sequences in many regions worldwide. They form a class of fluvial archive that is distinct from the extreme stability observed in Archaean cratons and the monatonic uplift or subsidence that is widely observed in other regions. Such alternations between uplift and subsidence are characteristic of regions of Early or Middle Proterozoic crust, where the initial crustal consolidation included the development of a thick 'root' of mafic material at the base of the crust; the present study focuses on localities with this

crustal type in the USA and eastern Europe. It has previously been suggested on the basis of uplift modelling that this style of crustal behaviour occurs only in regions where the mobile lower-crustal layer is relatively thin. This study supports this conclusion on the basis of independent geothermal calculations, which indicate that such alternations between uplift and subsidence occur where the mobile lower-crustal layer is ≤~7 km thick. An understanding of this phenomenon, in relation to the understanding of vertical crustal motions induced by surface processes (and thus by climate change) in general, therefore requires analysis of the properties and dynamics of the mobile lower-crustal layer; detailed analysis of fluvial sequences thus contributes unique information in this area.

Key words: Uplift, incision, topography, isostasy, Late Cenozoic, fluvial sequences

Long-timescale fluvial sequences, typically river terraces, are valuable data-sources for investigating Late Cenozoic ‘epeirogenic’ vertical crustal motions within continental interiors (e.g., Bridgland & Westaway, 2008a,b; Westaway *et al.*, 2009a). Three characteristic patterns are indicated. First, Archaean cratons are highly stable, with fluvial sediment-body preservation that implies minimal Quaternary vertical motions. Second, most other regions have experienced vertical motions in a constant sense, consistent with overall increases in relief; typically, uplands have uplifted whereas offshore regions and terrestrial depocentres have subsided. This pattern is characteristic of western and central Europe, where many of the world’s best-documented long-timescale Late Cenozoic fluvial sequences are located. In some regions these records extend back in time for many millions of years (e.g., Matoshko *et al*., 2004; Demir *et al*., 2012). Rates of this vertical motion have varied over time, in phases that correlate with climate change, suggesting variation in phase with fluctuating rates of surface processes that have been climatically-induced. A widespread effect, regarded as an important positive feedback mechanism (e.g., Westaway, 2002a, 2007a; Westaway *et al.*, 2006a; Morley & Westaway, 2006; Westaway & Bridgland, 2007), is for increased rates of erosion from upland regions to cause more rapid sedimentation in adjacent depocentres, thus accelerating the displacement of lower-crustal material from beneath the depocentre to beneath the sediment-source. A third pattern has also been observed, involving alternation between uplift and subsidence on timescales of hundreds of thousands of years or more (e.g., Bridgland & Westaway, 2008a; Westaway *et al*, 2009a). This study will review a selection of Late Cenozoic fluvial sequences that exhibit such reversals in vertical crustal motion or provide evidence for behaviour near the threshold for such reversals to develop, and will use this evidence to infer a relation to physical properties of the crust. This selection is chosen, from the larger global dataset demonstrating similar behaviour (e.g., Bridgland & Westaway, 2008a; Westaway et al., 2009a), to include localities in the USA and Eastern Europe where the fluvial sequences and crustal properties are sufficiently well documented for comparison to be made; it includes localities in the USA where the nature and causal mechanism of the observed vertical crustal motions are currently subject to lively debate. It is important to note that this paper deals with epeirogenic uplift in intraplate regions where the only viable mechanism for driving vertical crustal motions is the effect of climate on surface processes; in plate boundary zones the active faulting that accommodates the plate motions can also give rise to localized vertical crustal motions, as is apparent from the interactions between the different forcing mechanisms in such regions (e.g., Seyrek *et al*., 2008; Abou Romieh *et al*., 2008, 2012).

The ‘jelly sandwich’ model for the continental lithosphere (e.g., Chen & Molnar, 1983), which holds that the lower crust is weaker than either the upper crust or the mantle lithosphere, is well established (see recent discussion of this model, for example by Bürgmann & Dresen, 2008, and Westaway, 2012). The transition from upper crust, which is elastic and can deform by brittle fracture (i.e., by faulting), to lower crust, which is plastic and deforms by flow, occurs in normal felsic continental basement rocks at ~350 °C (T_b; after Sibson, 1983, and others). Conversely, mafic lithologies, representing the mantle lithosphere or mafic underplating at the base of the continental crust, have a temperature threshold for plastic behaviour of ~600 °C (e.g., Jackson *et al.*, 2008). Thus, in many continental regions the base of the mobile layer is the top of the mafic underplating, not the Moho. Flow in this mobile layer is thus regarded by some (e.g., many contributors to the Law *et al*, 2006, compilation) as 'mid-crustal flow'; however, the present authors prefer the term 'lower-crustal flow', irrespective of whether a mafic basal layer is present. There is indeed usually a thick mafic basal layer in crust that consolidated in the Early or Mid Proterozoic, having apparently formed by magmatic differentiation processes at that time (see below); elsewhere, magmatism associated with mantle plume activity may have emplaced a basal mafic layer during or after crustal consolidation.

One physical property that is key to the present analysis is W_i, a measure of the thickness of the mobile lower-crust (e.g., Westaway, 2001; Westaway *et al.*, 2002), defined as equal to z_i-z_b, z_i being the depth where the lower-crustal flow is concentrated and z_b the depth of the base of the brittle upper-crust (at temperature T_b=~350 °C; see above). If the mobile layer has a temperature-dependent linear viscous rheology and a uniform steady-state geothermal gradient, then z_i-z_b is ~90% of z_u-z_b, z_u being the depth of the base of the mobile layer (Westaway, 1998). If z_u is known, say from seismic profiling studies, one may thus estimate z_b from uplift-modelling as z_u–10×W_i/9; in principle, this estimate can be compared with independent geothermal evidence as a test. W_i also governs the characteristic timescale of the uplift-response during a phase of lower-crustal flow forcing (LCFF), the time taken to attain the resulting maximum uplift-rate being proportional to W_i^2, estimated (Westaway, 2012) as ~0.1, ~0.4, ~1.0 or ~1.8 million years for W_i=5, 10, 15 or 20 km, respectively. Modelling of uplift histories revealed by fluvial sequences can thus determine W_i and thereby reveal the thickness of the mobile lower-crust. Such modelling envisages the uplift as an isostatic consequence of changing rates of surface processes, governed by climate change (see below). A likely cause is the loss of vegetation cover caused by increased aridity or reduced temperature, resulting in the aforementioned enhancement of erosion, which can be shown to have occurred at several characteristic points within the Late Cenozoic record and given rise to phases of vertical crustal motion. No new modelling of this type is presented here; readers are instead referred to previous studies (e.g., Bridgland & Westaway, 2008a).

This paper will discuss evidence from fluvial sequences for reversals in vertical crustal motion, drawing upon datasets from Eastern Europe and the USA. In each case geothermal evidence is utilized to estimate the thickness of the mobile lower crust. The wider implications of this analysis, for example for resolving between contradictory interpretations of equivalent evidence, will then be discussed.

Eastern Europe

The Törnqvist–Teisseyre zone (TTZ) (or Trans-European Suture Zone), running NW-SE across Poland and western Ukraine, is one of the principal geological boundaries in Europe, separating relatively young (Variscan and younger) crust to the southwest from more ancient (Proterozoic) crust of the East European Platform to the northeast (Fig. 1). This boundary marks a significant decrease in heat-flow (Fig. 2), and thus lower temperatures at mantle lithospheric depths, consistent with seismic tomography (e.g., Goes *et al*, 1999). Localities west of this boundary invariably show monotonic Late Cenozoic vertical crustal motions (e.g., Westaway, 2002b; Bridgland & Westaway, 2008a,b), typically indicative of $W_i \approx 10$km. This is consistent with what is known independently of the thermal state of the crust, which typically consists of ~20 km of upper crust, with the Moho usually slightly deeper than 30 km; in some localities kilometres of mafic crustal underplating are also present (e.g. Westaway et al. 2009b). The River Dniester, which flows NW–SE near the TTZ (Fig. 1), in the region where the thermal state of the crust tapers laterally (Fig. 2B), likewise has a terrace staircase indicating monotonic uplift (Fig. 3a) and consistent with $W_{i=}$~8 km (Bridgland & Westaway, 2008a). However, as noted previously (e.g., Matoshko *et al.*, 2004), the major rivers farther inside the East European Platform, the Dnieper, Don and Volga, have recorded alternations in the sense of long-timescale vertical crustal motions (as in Fig. 3B,C).

These eastern European fluvial sequences have been extensively studied, the work being documented in detail in the local literature, of which Matoshko *et al*. (2004) have provided a summary that is accessible to international readers. The chronologies are constrained by multiple techniques, including magnetostratigraphy and mammalian and molluscan biostratigraphy. As a result, uncertainties in ages may reach many hundreds of thousands of years for some of the older sites from the Miocene or Pliocene; such uncertainties affect rates of vertical crustal motion but have no bearing on the existence of reversals in the sense of this motion.

Details mentioned in this section are, unless otherwise stated, from Shchipansky & Bogdanova (1996), who reviewed the age-control evidence and standardized the nomenclature for crustal domains in the southern part of the East European Platform. The oldest crust in that part of the East European Platform under consideration here, the Sumy-Dnieper domain, extends south from the vicinity of Kursk (Fig. 1) to northern Crimea. This crust formed in the Archaean, during ~3700–2800 Ma; it is thought to have become fully consolidated by the late Archaean and was unaffected by the pervasive Early Proterozoic deformation in surrounding regions. In the lower Dnieper, which passes through this crustal domain, Matoshko *et al.* (2002) noted the close spacing of river terraces, which evidently reflects the relative stability of this Archaean crust. The cross-section in Fig. 3C is located within the Sevsk-Ingulets crustal domain. This region has yielded some Archaean dates (~3100–2800 Ma) but the crust was clearly not consolidated then, as widespread Early Proterozoic deformation and intrusive igneous activity followed, during ~2200–2000 Ma.

The Ukrainian Shield is characterized by surface heat-flow of ~30–50 mW m^{-2} (e.g., Kutas *et al*, 1979; Kutas, 1979). In a relatively high-latitude region such as this, surface heat-flow can be significantly affected by the variations in the surface temperature T_o during the Pleistocene and Holocene. In particular, the warming at the Pleistocene-Holocene transition has reduced the geothermal gradient, and thus the heat-flow, in the shallow crust. However, such effects are only significant within ~1 km of the Earth's surface (e.g., Majorowicz & Wybraniec, 2001; Westaway & Younger, 2013), whereas the borehole heat-flow measurements in the Ukrainian Shield have typically been made at ~2 km (Kutas, 1979). Kutas (1979) envisaged that the regional surface heat-flow consists of a basal component of ~25 mW m^{-2} combined with variable contributions from radioactive heat-production in the upper crust. His calculations thus place the 350 °C geotherm at typical depths of ~30-35 km. Extrapolation of the geotherm for ~350 °C at ~30 km depth assuming a basal heat-flow of ~30 mW m^{-2} would indicate a lithospheric thickness (the depth where the temperature reaches ~1400 °C) of ~135 km.

We have verified these calculations, using standard theory by Lachenbruch (1970) (discussed in more detail below), which is applicable for uniform radioactive heat production Y at depths shallower than D, with no radioactive heat production elsewhere, in rock of thermal conductivity k, with a basal contribution to heat-flow q_o and with T_o ~10 °C. Kutas (1979) associated surface heat-flow of ~40 mW m^{-2} with Y=1.6 μW m^{-3} and D ~7 km, which would indicate q_o ~28.8 mW m^{-2}, and, with k=3 W m^{-1} $°C^{-1}$ (a representative value for the metamorphic basement lithologies in the region; cf. Kutas, 1979), 350 °C is predicted at ~34.1 km depth. Likewise, Kutas associated surface heat-flow of ~50 mW m^{-2} with Y=3.35 μW m^{-3} and D ~5 km. This would indicate q_o ~33.3 mW m^{-2} so, again with k=3 W m^{-1} $°C^{-1}$, 350 °C is now predicted at ~29.4 km. These calculations are in close agreement with those by Kutas, which assumed more complex distributions of radioactive heating, thus confirming his essential conclusions. Seismic profiling indicates that the Moho is at ~40-50 km depth beneath the Ukrainian Shield and is overlain by a crustal layer with P-wave velocities of ~6.8-7.2 km s^{-1}, as expected for mafic underplating, its top at ~32-38 km depth (e.g., Ilchenko, 1996). Comparison of these predicted depths for the 350 °C geotherm with depths of the top of the mafic underplating indicates that no mobile lower-crustal layer exists beneath some parts of the Ukrainian Shield, whereas elsewhere a mobile layer of a few kilometres thickness is present.

The Dniester

The River Dniester flows parallel to and ~50-200 km NE of the TTZ, through the SW part of the Dniester-Bug crustal domain. There is evidence of Archaean material here as old as ~3400 Ma, but the crust experienced regional metamorphism and igneous intrusion much later, around 2300-2100 Ma, so (in common with elsewhere in the East European Platform) was not fully consolidated (or cratonized) during the Archaean. The Dniester sequence is summarized in Fig. 3A. Deposition of a thick stacked fluvial sequence (the Balta Series or Balta Group) in the Late Miocene was superseded by fluvial incision and the development of a river terrace staircase reaching ~250 m above the present river. The Balta Group accumulated as the ‘coastline’ of the landlocked ‘Paratethys Sea’ retreated south towards the present Black Sea. The ancestral Dniester thus aggraded to maintain its gradient, forming a stacked deposition that does not, therefore, imply subsidence; on the contrary, the progressive coastal retreat suggests that regional uplift was already occurring.

Modelling of the uplift history revealed by the Dniester terrace staircase, by Bridgland & Westaway (2008a), indicates W_i=8 km, making the mobile lower-crust ~9 km thick. An early phase of LCFF, to account for the uplift during deposition of the Balta Group, and four more phases are required, starting at 6, 3.1, 2 and 0.9 Ma. The first of these, marking the initial incision into the Balta Group and the formation of the Stolnichen and Kuchurgan terraces (Fig. 3A), was presumably caused by the Pontian desiccation of the Black Sea region. This event, roughly contemporaneous with the Messinian salinity crisis in the Mediterranean, is thought to have resulted from the temporary capture of the Danube catchment by the Mediterranean basin, due to its lowered base-level (e.g., Kvasov, 1983; Clauzon *et al.*, 2005; Bache *et al.*, 2012; cf., Sakınç & Yaltırak, 2005). Increased erosion is apparent from the significantly enhanced sedimentation rate observed at ~6 Ma in orbitally-tuned sedimentary records (e.g., Vasiliev *et al.*, 2004), and by a significant erosional unconformity in other regions adjoining the Black Sea (e.g., Karlov, 1961; Clauzon *et al.*, 2005). The subsequent phases of LCFF, starting at ~3.1, ~2.0 and ~0.9 Ma, likewise correlate with times of climate change, as is discussed below.

The relatively high value of W_i deduced from the modelling of the Dniester succession also warrants comment, in the light of calculations, discussed below, that lower W_i characterizes the East European Platform and other ancient crustal provinces in general. Figure 2 shows the thermal state of the crust visibly perturbed up to ~150 km from the TTZ, the crust being hotter than at sites farther inside the East European Platform; this warming effect is due to the insulating effect of the sedimentary fill of the Carpathian Foredeep, along which oceanic lithosphere attached to the East European Platform has subducted beneath the region to the southwest during the Late Cenozoic. In the vicinity of the lower Dniester, where much of the fluvial terrace evidence originates, the surface heat-flow is ~45 mW m^{-2} and the sedimentary fill is up to ~2 km thick, with k ~1.75 W m^{-1} °C^{-1} (Kutas *et al.*, 1979). This surface heat-flow is regarded as a basal component of 29 mW m^{-2} and a radiogenic component in the uppermost basement characterized by D=8 km and Y=2 μW m^{-3}. If so, the temperature is ~61.4 °C at the base of the sediment and ~160.0 °C at the base of the radiogenic basement layer, so z_b (the depth for T_b=350 °C) is estimated as 29.6 km below the Earth's surface, or 27.6 km below the top of the basement. Without the sedimentary cover, but with the same inferred distribution of radioactive heat production, z_b would be 33.0 km, so the ~2 km of sediment has reduced the depth of the brittle-plastic transition, relative to the top of the basement, and has thus increased the thickness of the mobile layer, by an estimated ~5.4 km. We are not aware of any measurement of the radioactive heat production in the basement in this region or of any determination of the depth of the top of the mafic underplating. We presume, by analogy with the Ukrainian Shield, that the top of the underplating is ~35±3 km below the top of the basement (Ilchenko, 1996; see above). The thickness of the mobile layer in this region can therefore be estimated as up to ~10 km, roughly consistent with the 8 km value of W_i determined from the uplift modelling, this value being greater than elsewhere in the East European Platform due to the thermal insulating effect of the sediment.

An additional factor, which may also affect the thermal state of the lithosphere within the Dniester catchment, is the horizontal component of heat-flow that will result from the juxtaposition of the East European Platform against much hotter crust across the TTZ (Fig. 2B). Thus, at ~30 km depth the temperature is ~700 °C in the latter region and (as confirmed by the calculations above) ~350 °C in the former. Kutas *et al.* (1979) did not incorporate this horizontal heat-flow into the calculated geotherms depicted in Fig. 2C, but estimated the upper bound to the associated horizontal geothermal gradient at lower crustal/upper mantle depths as ~3° C km^{-1}, which is substantial compared with the unperturbed vertical geothermal gradient at such depths within the East European Platform: ~10 °C km^{-1}. Temperatures at depth beneath the Dniester catchment are thus expected to be somewhat higher than the predictions illustrated in Fig. 2B. The relatively high value of W_i for the Dniester catchment relative to sites (to be discussed below) in other crustal ancient provinces is evidently a consequence of both insulation of the basement by low-conductivity sediment and horizontal conduction of heat; this region is therefore an exception to the general rule that W_i is lower in ancient crustal provinces.

The Don

In the Voronezh area of SW Russia the River Don flows through Voronezh Shield or the Lipetsk-Losev crustal domain, where the crust is Early Proterozoic, dating from ~2300-1900Ma, with no evidence of Archaean material. The Don sequence dates back to the late Middle Miocene (the Sarmatian stage of the Paratethys realm) (Fig. 3B). Around Voronezh, an initial terrace staircase, spanning the late Middle Miocene to the latest Pliocene (the Fomenkovo to Belaya Gora terraces, inclusive; Fig. 3B) is explained by three phases of LCFF, starting at 12, 6 and 3.1 Ma. This early terrace staircase is buried beneath younger deposits (Fig. 3B). To explain the higher altitudes of the Goryanka and Petropavlovka terraces using the adopted modelling technique, another phase starting at 2 Ma is required, but with the opposite sense (i.e., driving net outflow, not net inflow, of lower crust), thus causing a reversal in the uplift history (i.e., subsidence). Renewed uplift, indicated by the incision from the level of the Petropavlovka terrace to that of the Yuzhnovoronezh I and II deposits (Fig. 3B), requires another phase starting at 1.2 Ma. The subsequent development of a stacked fluvial succession, interspersed with loess layers, temperate-stage palaeosols, and deposits of the MIS 16 Don glaciation, lasted until the aggradation of the Fourth terrace of the Don, dated to MIS 8, and requires another phase of LCFF with the opposite sense, starting at 870 ka. The final phase, starting in MIS 8 (250 ka), accounts for the incision below this Fourth terrace.

Bridgland & Westaway (2008a) modelled the Don sequence, showing it consistent with W_i=5 km. Seismic profiling by Tarkov & Basula (1983) indicates that the Voronezh Shield crust has a basal layer with a P-wave velocity of 6.95–7.8 km s^{-1}, consistent with mafic underplating. Along the profile closest to the city of Voronezh this layer occurs between 33.8 km depth and the Moho at 44.8 km. The surface heat-flow in this region is ~40 mW m^{-2} (Kutas *et al.*, 1979). If this arises as a result of the same distribution of crustal radioactivity as for the Ukrainian Shield, then the brittle upper-crust would, likewise, be ~35 km thick, so no mobile lower-crustal layer would be expected. On the other hand, if there were no upper-crustal radioactivity, then the 350 °C isotherm would occur at ~25.5 km and a mobile layer ~8.3 km thick would be present, requiring W_i=~7.5 km. Conversely, if Y and D for the Voronezh Shield are taken as ~0.6 μW m^{-3} and 10km, values inferred in the least radiogenic parts of the Ukrainian Shield by Kutas (1979), then the 350 °C isotherm would occur at ~29.1 km depth and a mobile layer ~4.7 km thick would be present. Notwithstanding the uncertainties in these calculations, it is probable that a thin mobile layer is present in this region, in approximate agreement with the value of W_i determined by Bridgland & Westaway (2008a).

The Dnieper

Figure 3C summarizes the Late Cenozoic sequence of the Dnieper east of Kiev, in central Ukraine, after Matoshko *et al.* (2004). This section is ~240 km long, its NE end located near the Ukraine-Russia border and the NE margin of the Palaeozoic Dnieper Basin, whereas its SW end is on the modern river, ~100 km downstream of Kiev, near the SW margin of this basin and the NE margin of the Ukrainian Shield (Fig. 1). The altitude range of the deposits is even more restricted than in the Don (Fig. 3B). The succession is also much less complete; notably, no deposits are known between the earliest Pliocene (Parafiivka Formation) and the early Middle Pleistocene (Traktemyriv Formation). Elsewhere, however, notably in the Chornobyl district (~200 km upstream of Fig. 3C), a

~35 m stacked fluvial succesion (the Chornobyl Formation) overlies the Parafiivka Formation and is thought to represent the late Early Pliocene and Middle-Late Pliocene (Matoshko *et al.*, 2004). That area, however, also lacks any record of the Early Pleistocene.

From the Kiev area, the Dnieper flows for ~400 km ESE along the Dnieper Basin, then turns abruptly into ~300km long Lower Dnieper, flowing SW to the Black Sea just west of Crimea (Fig. 1). As was noted above, Matoshko *et al.* (2002) described the river terraces as more closely spaced in the Lower Dnieper than in the reach farther upstream (cf. Fig. 3C, suggesting greater crustal stability.

An uplift modelling solution for the Dnieper has been prepared and is consistent with W_i=5 km, as for the Don. However, this uses a composite dataset for the region as a whole (i.e., for the area in Fig. 3C and the Chornobyl area), and the modelling takes no account of any lateral variations in crustal rheology between the Dnieper Basin and the Ukrainian Shield (see below); thus the resulting determination of W_i only provides a general indication of the thickness of the mobile lower-crust over a large region, rather than a value representative of any locality in particular (the modelling solution is therefore not particularly useful, so is omitted here). In this solution phases of LCFF starting at 8.0 and 6.0 Ma are used to model the net incision down through the levels of the Shostka and Pyryatin Formations to that of the base of the Parafiivka Formation, this deep incision reflecting erosion during the Pontian regression of the Black Sea (Matoshko *et al.*, 2004). Two further phases of LCFF, in the opposite sense, starting at 5.3 and 3.1 Ma are then used to model the subsidence accompanying deposition of the Chornobyl Formation. The next phase, starting at 2.0 Ma, created the incision down to the base of the Traktemyriv Formation. The final two phases, assumed (like in the Don) to start at 870 ka and 250 ka, account for the subsidence indicated by the stacked early Middle Pleistocene succession and the subsequent fluvial incision.

The Ukrainian and Voronezh shields were separated by rifting in the Devonian, creating the NW–SE- or WNW–ESE-trending Dnieper Basin in the Kiev area and its along-strike continuations (Fig. 1), the Pripyat Trough and Donets Basin (e.g., de Boorder *et al.*, 1996; Juhlin *et al.*, 1996). This rifting was associated with widespread basaltic magmatism, suggesting the presence of a mantle plume (e.g., Chekunov *et al.*, 1992; Wilson & Lyashkevich, 1996). Prolonged post-rift deposition followed, the maximum sediment thickness reaching ~10 km in the Dnieper Basin and ~20 km in the Donets Basin (e.g., Ilchenko, 1996; Maystrenko *et al.*, 2003). Along a SSW–NNE profile located ~100 km east of Kiev (thus roughly coinciding with Fig. 3C) this maximum thickness reaches ~7 km (Ilchenko, 1996) or ~8 km (Kusznir *et al.*, 1996). The extension that created this basin was magmatic, as the mafic underplating beneath its axis is much thicker than elsewhere in the East European Platform, indicating that the Palaeozoic mantle plume added to the material already present as a result of the region's earlier tectonic history (e.g., de Boorder *et al.*, 1996; Ilchenko, 1996; Juhlin *et al.*, 1996; Maystrenko *et al.*, 2003). The top of material with P-wave velocity 7.2 km s^{-1} is at a depth of ~18 km beneath a ~40 km wide zone along the basin axis (Ilchenko, 1996), thus only ~11 km below the base of the sediment. Similar results have been obtained from gravity modelling by Lobkovsky *et al.* (1996). This ~40 km wide axial zone of the basin coincides with a heat-flow high of ~50 mW m^{-2} of similar lateral extent (Kutas *et al.*, 1979). Again using 1.75 W m^{-1} $°C^{-1}$ as a representative thermal conductivity for sediments overlying the East European Platform, from Kutas (1979), and assuming a constant geothermal gradient, the temperature at the base of the sediment, at 7 km depth, can be estimated (given T_o=10 °C) as ~210 °C. Assuming no radioactivity in the underlying basement, but a thermal conductivity of 3 W m^{-1} °C^{-1}, the 350 °C isotherm is predicted at a depth of 15.4 km, so the mobile lower crustal layer can be expected to be no more than 2.6 km thick in this locality. However, with even a small contribution to heat-flow from radioactive heat production in the uppermost basement, the temperature at the top of the underplating will be lower, so no mobile layer may well exist locally.

On the other hand, at a site away from the rift axis, where the top of the underplating is much deeper (~30 km; Ilchenko, 1996), the sediment will be somewhat thinner (say, 6 km thick), and heat-flow lower (surface heat-flow, say, 45 mW m^{-2}, made up of a basal component of 29 mW m^{-2} and a radiogenic component in the uppermost basement characterized by D=8 km and Y=2 μW m^{-3}). The temperature can thus be estimated as ~164.3 °C at the base of the sediment and ~240.3 °C at the base of the radiogenic basement, making z_b 23.4 km. The mobile layer is thus ~6.6 km thick, in approximate agreement with the 5 km value of W_i deduced from the uplift modelling. The presence of thick sediment with low thermal conductivity in the Dnieper Basin thus helps to maintain relatively high temperatures at lower-crustal depths, thereby sustaining (at sites where the underplating is not too thick) a mobile layer of modest thickness.

When the SW-flowing Lower Dnieper exits this sedimentary basin it enters the region of Archaean crust with lower heat-flow (~40 mW m^{-2}; Kutas, 1979); in accordance with earlier calculations, the combination of modest radioactive heat production in the upper crust, thick underplating, and apparently thick lithosphere resulting in low basal heat-flow, evidently constricts the mobile layer (maybe eliminating it completely in some places; cf. Westaway *et al.*, 2003), increasing the crustal stability and explaining the more closely spaced fluvial terraces reported by Matoshko *et al.* (2002).

The USA

Further evidence for the dependence of Late Cenozoic uplift on crustal properties comes from fluvial sequences in the USA. The Archaean (~3200–2700 Ma) Superior crustal province underlies the north-central USA and much of central/eastern Canada (e.g, Kendall *et al.*, 2002). This is bounded to the south by the Penokean province in the central USA, which consolidated in the Early Proterozoic (~1900–1800 Ma) (e.g., Zhao *et al.*, 2002). The Penokean province is bounded to the west and south by the Yavapai and Mazatzal provinces, which persist west to the Rocky Mountains and Colorado Plateau and consolidated in the Early–Middle Proterozoic (by ~1600 Ma; e.g., Whitmeyer & Karlstrom, 2007). This ancient core of North America is bounded to the east by Late Proterozoic ('Grenville') and Palaeozoic crust, which underlies the Appalachian Mountains and Atlantic coastline (e.g., Ingle *et al.*, 2003; Hughes *et al.*, 2004; Ownby *et al.*, 2004). Tomography (van der Lee & Nolet, 1997; Goes & van der Lee, 2002) indicates that the thermal state of this lithosphere is largely determined by the age of the crust, being hotter where younger.

In the eastern United States, rivers in the relatively young crust of the Appalachian Mountains (up to ~45km thick, with no basal mafic layer; Durrheim & Mooney, 1994) have 'conventional' terrace staircases, examples being the Susquehanna (e.g., Pazzaglia & Gardner, 1993) and upper Ohio (e.g., Jacobson *et al.*, 1988). Modelling indicates high values of W_i, ~18–20 km, for these river terrace staircases (Westaway, 2007b). The surface heat-flow within the Appalachians is quite low, reaching a reported minimum of ~40 mW m^{-2} (e.g., Wisian *et al.*, 1999). Nonetheless, assuming no radioactive heating in the upper crust, with k=3 W m^{-1} $°C^{-1}$ and T_o=10 °C, the 350 °C isotherm is at ~25.5 km, broadly consistent with the observed value of W_i. However, like elsewhere in the USA, it is apparent from the original data source (http://smu.edu/geothermal/georesou/usa.htm) that many U.S. heat-flow measurements are from shallow boreholes and have not been corrected for palaeoclimate. This suggests (see earlier discussion; cf. Majorowicz & Wybraniec, 2011; Westaway & Younger, 2013) that the true heat-flow may have been underestimated. A similar temperature at mid-crustal depths may thus result, in each case, from a higher surface heat-flow but with some decrease in heat-flow with depth due to radioactive heating in the upper-crust.

The tomographic evidence (van der Lee & Nolet, 1997) indicates that the temperature at 110 km depth beneath the part of the Appalachians with the lowest heat-flow is ~1200 °C. This is much less than would be predicted from the aforementioned extrapolation but could be accommodated instead with a more complex crustal model. For example, if the uppermost 5km of the crust is assumed to be sediments with k=2 W m^{-1} $°C^{-1}$, and the uppermost basement has Y=1 μW m^{-3} over D=10 km, then the 350 °C isotherm would occur at 33.5 km depth, making the temperature at 110 km ~1180 °C, in rough agreement with the tomography. This would imply a mobile layer ~11.5 km thick, albeit thinner than that estimated by Westaway (2007b).

Green River, Kentucky/Mammoth Caves karstic levels

An example of the contrasting pattern, with reversals in vertical crustal motion, observed within the USA, is provided by the record from the karstic region of western Kentucky, drained by the Green River, a tributary of the lower Ohio (e.g., Granger *et al.*, 2001) (Fig. 4). In karstic systems such as this, the height of the water table is controlled by the depth of fluvial entrenchment in the adjacent river valley. Dating of cave development, in this case cosmogenic dating of coarse sediments washed into the cave system and deposited at each cave level, thus reveals histories of fluvial incision and aggradation. According to Fig. 21 of Braile (1989), the crust is locally 52 km thick, the basal 14 km being interpreted (e.g., by Durrheim & Mooney, 1994) as mafic underplating. The tomography (van der Lee & Nolet, 1997) indicates that the temperature is ~900 °C at 110 km depth. Modelling of the local history of vertical crustal motion requires W_i only ~4 km (Westaway, 2007b), again consistent with the relatively cold crust.

The surface heat-flow is locally ~50 mW m^{-2} (e.g., Wisian *et al.*, 1999). Assuming k=3 W m^{-1} $°C^{-1}$ for the crustal basement, as before, then a temperature of 350 °C would be expected at ~20.4 km depth if there is no radioactive heating in the upper crust, raising the possibility of a mobile layer of substantial thickness. However, such a temperature distribution would extrapolate to ~1850 °C at 110 km depth, which is much too hot given the tomographic evidence. If Y=2 μW m^{-3} and D=10 km, then the predictions would be ~30.7 km for z_b and ~7.3 km for the thickness of the underlying mobile layer. The resulting basal geothermal gradient of ~10 °C km^{-1} would give a temperature of ~1150 °C at 110 km, which is still too high. Nonetheless, with Y=2.5 μW m^{-3} and D=10 km, the predictions would adjust to ~35.8 km for z_b and ~2.2 km for the thickness of the underlying mobile layer. The resulting basal geothermal gradient of ~8.3 °C km^{-1} would give a temperature of ~970 °C at 110 km, much closer to the ~900 °C value from the tomography. The further refinement of incorporating a low-conductivity sedimentary layer above the basement and a modest increase in the radioactive heating would increase the predicted thickness of the mobile layer while decreasing still further the predicted temperature at 110 km depth; for example, with a 2 km thick surface layer with k=2 W m^{-1} $°C^{-1}$ and with Y=2.7 μW m^{-3} and D=10 km in the underlying basement, z_b would be 34.0 km, making the mobile layer ~4 km thick, with a temperature of ~930 °C at 110 km. Significant radioactive heat production in the uppermost crustal basement is thus required to account for the low value of W_i deduced from uplift modelling jointly with the other evidence from this region.

The Platte River catchment

The terraces of rivers flowing eastward from the Rocky Mountains to the Mississippi/Missouri indicate a more gradual transition from monotonic uplift to reversals in vertical motion, an example being the Platte (Figs. 5, 6). Near Denver, Colorado, just east of the Rockies, the South Platte has a 'conventional' terrace staircase (Fig. 5), for which modelling (Bridgland & Westaway, 2008a) indicates W_i ~8-10 km, making the mobile layer ~9-11 km thick. The crust beneath Denver was long regarded as ~50 km thick (e.g., Allenby & Schnetzler, 1983; Braile, 1989), with mafic underplating accounting for the lower part of the crust, at >30 km depth (e.g. Durrheim & Mooney, 1994; Karlstrom *et al.*, 2002). More recent studies (e.g., Gilbert & Sheehan, 2004; Karlstrom *et al.*, 2012) estimate ≥45 km thick crust in this area, implying ~15 km of underplating instead of the ~20 km thickness formerly envisaged. The Gilbert & Sheehan

(2004) dataset indeed illustrates a seismic phase that might well result from the acoustic impedance contrast at the top of this layer of underplating at ~30 km depth. The tomography (van der Lee & Nolet, 1997) indicates a temperature of ~1400 °C at 110 km depth, which thus roughly marks the base of the lithosphere. The best local indication of heat-flow is from the Rocky Mountain Arsenal borehole (Sass *et al.*, 1971), which reached ~3600 m depth, the deepest ~600 m being in Proterozoic basement beneath sedimentary cover. The bottom-hole temperature was found to be ~112 °C and the heat-flow in the basement (with k=3.2 W m^{-1} $°C^{-1}$) was determined, after topographic correction, as 79 mW m^{-2}. The top of the basement is thus at ~97 °C; taking Y=4.2 μW m^{-3} and D=10 km in the basement, a basal heat-flow of ~37 mW m^{-2} is indicated, giving a temperature of ~1400 °C at 110 km depth, as suggested by the tomography. The 350 °C isotherm is thus expected at ~19.2 km, making the mobile layer ~10.8 km thick, in good agreement with the estimate of W_i from Bridgland & Westaway (2008a). Reiter (2008) reported a revised heat flow of 83 mW m^{-2} in this borehole (based on Decker, 1995, and Decker et al., 1988) but without explaining the revision; on this basis a slightly thicker mobile lower-crustal layer would be predicted. We are not aware of any measurement of the heat production in this particular borehole; however the value adopted is similar to others measured in the surrounding region (e.g., Edwards *et al*., 1978) and the resulting basal heat flow is within the ~34±4 mW m^{-2} range long considered characteristic of the Great Plains (e.g., Roy *et al*., 1968; Decker & Smithson, 1975).

The South Platte terrace staircase indicates ~500 m of incision near Denver (Fig. 5), less farther downstream (Fig. 6(a)). The North Platte flows through the High Plains of eastern Wyoming and western Nebraska, where it has formed a similar terrace staircase (Fig. 6(b)). The land surface, locally known as the Ogallala surface, is at the top of a stacked fluvial sequence, the Ogallala Group, derived from erosion of the Rocky Mountains during the Middle–Late Miocene and dated using biostratigraphy and tephrochronology (e.g. McMillan *et al.*, 2002; Heller *et al.*, 2003; Condon, 2005; Wobus *et al.*, 2010). These deposits are inset by younger fluvial deposits, of the Broadwater Formation, which provide a record of fluctuations in climate and palaeo-hydrology in and around the Mid-Pliocene climatic optimum (Duller *et al*., 2012). In the locality depicted in Fig. 6B, the river incised by ~90m into the Ogallala Group before aggradation, during deposition of the Broadwater Formation, back to almost exactly the same level (see Fig. 1(c) of Duller *et al*., 2012). The local relief thus represents the incision since the Mid-Pliocene climatic optimum and began at ~2.5 Ma according to Duller *et al*. (2012). Analysis of palaeo-hydrology by McMillan *et al.* (2002) and Heller *et al.* (2003) established that the Ogallala Group sediments were deposited by rivers with gradients that decreased downstream from ~1.5 m km^{-1} in eastern Wyoming and western Nebraska to ~1 m km^{-1} in central Nebraska, similar to the modern Platte system (see, however, Wobus *et al.*, 2010; see below). In contrast, the Ogallala surface slopes more steeply eastward, decreasing from ~8m km^{-1} in central Wyoming to ~2 m km^{-1} in west-central Nebraska. For instance, near the town of Ogallala (Fig. 6A), the North Platte descends from ~1000 to ~910 m a.s.l. over ~75 km at ~1.2 m km^{-1}. In the same area, between longitudes 103° and 101°W, the Ogallala surface descends from ~1260 to ~880 m a.s.l. (comparing fluvial data from Eaton, 1987, with local topographic data), a gradient of ~2.3 m km^{-1}. The fact that the Ogallala surface is steeper than the modern river gradients

(McMillan *et al.*, 2002; Heller *et al.*, 2003), corresponding with a west–east decrease in fluvial incision, is clear evidence of post-Early Pliocene uplift by net amounts that decrease eastward (contra Wobus *et al.*, 2010; see below). Furthermore, the evidence that river gradients have not changed significantly on this timescale means that the observed fluvial incision is a good proxy for regional uplift.

Nereson *et al*. (2013) have recently proposed similar development of topography farther south in SE Colorado and NE New Mexico, where dated basalt flows and sediments of the Ogallala Group in interfluve locations flanking the upper reaches of the Canadian River indicate post-Early Pliocene incision decreasing eastward from ~600 m near the front of the Rocky Mountains (the Sangre de Cristo Range) to ~350 m over a distance of ~200 km. Here, too, there is no reason why the longitudinal gradient of the river should change over time, so this incision can likewise serve as a proxy for uplift.

The total incision (and thus uplift) decreases from ~500 m at Denver to ~150 m in western Nebraska (Fig. 6B) and to a few dozen metres in central Nebraska (Fig. 6C). Farther east there are more complex fluvial records that indicate reversals in the pattern of vertical crustal motions (Fig. 6D and 6E). Finally, near the Missouri confluence vertical motions have been minimal, indicating ultra-stability of the crust (Figs. 6F and 6G). Given the uplift modelling solutions discussed earlier, these observations suggest an eastward decrease in W_i from ~10 km in the Denver area to perhaps 4–6 km in western and central Nebraska and then to zero in eastern Nebraska.

Moving eastward along this traverse, the surface heat-flow decreases from the aforementioned higher values in the Denver area to ~60 mW m^{-2} across Nebraska (e.g., Wisian *et al.*, 1999). The crustal thickness decreases gradually to ~45 km, but with the top of the mafic layer at a roughly constant ~30 km (e.g., Durrheim & Mooney, 1994). The temperature at 110 km depth decreases eastward from ~1400 °C around Denver to ~700 °C in and around eastern Nebraska (van der Lee & Nolet, 1997), indicating a significant eastward thickening of the continental lithosphere. The sedimentary cover overlying the basement also thins eastward; in and around eastern Nebraska Precambrian basement crops out or is in the shallow subsurface.

This approximate halving of the temperature difference between 110 km depth and the Earth's surface implies a rough doubling of the lithospheric thickness and thus an approximate halving of the basal heat-flow, which can therefore be estimated for eastern Nebraska as ~20 mW m^{-2}. This implies that some 40 mW m^{-2} of the surface heat-flow in eastern Nebraska is derived from radioactive heat production in the upper crust. Estimating k=3 W m^{-1} $°C^{-1}$, D=10 km and Y=4 μW m^{-3}, similar values to the Denver area, one obtains an ~810 °C temperature at 110 km with z_b ~41.0 km. No mobile layer is thus present in this region, consistent with the observed crustal stability (cf. Figs 6F and 6G), and this would remain so if the parameters were adjusted (say, by increasing Y) to reduce the predicted temperature at 110 km to the value indicated by the tomography. As one moves westward, the increase in the basal heat-flow evidently results in a threshold being reached where a mobile layer is present, presumably causing the different pattern of Quaternary landscape development that is observed (Fig. 6). The combination of thinner lithosphere and a higher mean temperature in the crust, and its resulting thermal expansion, presumably contribute, given isostatic equilibrium, to the higher topography that is present.

The Colorado

A further example indicative of similar effects is provided by the Colorado River within the Colorado Plateau of the SW USA, another region with a thick mafic layer at the base of the crust, attributed to magmatic underplating during Early Proterozoic crustal consolidation (e.g., Wendlandt *et al.*, 1993). Figure 7 illustrates the Colorado sequence near Grand Junction in the western part of the state of Colorado, modified after Bridgland & Westaway (2008b). This composite transverse profile sought to integrate data from sites (up to ~250 km apart) adjoining both the main Colorado River and its tributary, the Gunnison; however, it is now clear (Karlstrom *et al*., 2012; Donahue *et al*., 2013) that rates of fluvial incision vary significantly across this substantial region. In the vicinity of Grand Junction the disposition of the dated Grand Mesa basalt indicates ~1500 m of incision since ~10 Ma at a time-averaged rate of ~0.15 mm a^{-1} (Karlstrom *et al*., 2012). Nearby river terrace deposits ~100 m above the modern valley floor, in which tephra from the ~600 ka Yellowstone eruption is preserved, indicate a similar incision rate; it has thus been argued on the basis of this sparse evidence that the incision rate has remained constant for ten million years (Karlstrom *et al*., 2012; Donahue *et al*., 2013). However, ~200 km upstream in the Glenwood Canyon area, heights of other dated basalt flows (including those at Spruce Ridge and Gobblers Knob in Fig. 7) 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). Aslan *et al*. (2010) proposed an *ad hoc* reinterpretation of these data in terms of a uniform incision rate, a suggestion that was dismissed by Polyak *et al*. (2013). Using speleothem data, Polyak *et al*. (2013) indeed resolved the local incision history into phases at ~0.3 mm a^{-1} between ~3 and ~0.9 Ma, decreasing to ~0.15 mm a^{-1} since 0.9 Ma. According to Scott *et al*. (2002), at Grand Junction the Colorado has a latest Pleistocene terrace 3-5 m above the modern river, then older terraces at heights of 24-37 m, 64-67 m, 80-100 m, and 163-175 m. The first three of these were assigned by Scott et al. (2002) to MIS 2, 12 and 16, the latter age assignment now being confirmed by recognition of tephra from the ~600 ka eruption of Yellowstone in a correlative terrace deposit of the Gunnison (a left-bank tributary that joins the Colorado at Grand Junction; Fig. 7) at nearby sites (Karlstrom *et al*., 2012). We thus infer that the highest terrace recognized dates from MIS 22 or thereabouts. The terrace heights thus indicate time-averaged incision rates of ~0.28 (MIS 6-2), ~0.15 (MIS 12-2), ~0.16 (MIS 16-2), and ~0.20 (MIS 22-2) mm a^{-1}. The MIS 6 terrace includes outwash from the MIS 6 Bull Lake glaciation (Scott *et al*., 2002); by analogy with other examples worldwide, associated glacial or glacio-isostatic perturbation of the fluvial system might have caused emplacement at a higher-than-expected relative height. However, in the reach of the Colorado upstream of Grand Junction, Yeend (1969) recognised pediments that grade to former river levels ~400 m and ~150 m above the modern floodplain. We suggest that these represent times of relative landscape stability, the higher in the Pliocene, as Yeend (1969) tentatively suggested (by which we mean >3 Ma), and the lower in the late Early Pleistocene, just before the start of the faster incision marked by the aforementioned fluvial terrace staircase. During this phase, the fastest incision, in the range 0.27-0.41 mm a^{-1}, is evident between MIS 22 and MIS 16, consistent with the expected uplift response to a phase of LCFF starting in MIS 22 (e.g., Westaway, 2001). Excluding sites in the vicinity of recent drainage diversions (Donahue *et al*., 2013), there is no reason why incision within this system should not serve as a proxy for uplift since the modern geometry of the Colorado catchment came into being at ~6 Ma (Bridgland & Westaway, 2012). Contrary to other recent studies (Karlstrom *et al*., 2012; Donahue *et al*., 2013), which have inferred constant fluvial incision rates for multi-million-year timescales in this region, we identify variable rates that correlate with the expected climatic forcing (see also below).

At Cactus Park (~70 km south of Grand Junction), fluvial gravels of the Uncompahgre palaeo-river (thought to be (?) late Early Pleistocene; Aslan *et al*., 2008, 2010), ~435 m above present river level, are overlain by ~40 m of lacustrine deposits (Fig. 7). Roughly contemporaneous diversion to form the modern Uncompahgre, which joins the Gunnison at Delta, ~100 km SE of Grand Junction, resulted in abandonment of this palaeo-valley (which led directly into the Colorado ~50 km downstream of Grand Junction), seemingly 'fossilising' the Early Pleistocene landscape. The height of the deposit relative to the modern valley-floor is largely a consequence of this diversion and thus no indication of uplift; the essential point relevant to the present study is that the infilling of this palaeo-valley with sediment of lithology indicative of transport by the Uncompahgre suggests a reversal from fluvial incision to subsidence, before fluvial incision later resumed (with the river now located elsewhere), analogous to the localities already discussed. Diverse explanations have hitherto been put forward to account for this deposit, the most recent being that the valley was blocked by a palaeo-landslide that was later completely removed by erosion (e.g., Aslan *et al*., 2008, 2010). A second example of a similar effect has been reported along the San Juan tributary in SE Utah (which joins the Colorado ~200 km downstream of Grand Junction), cosmogenic dating (Hanks & Finkel, 2005) suggests that a 31 m thick fluvial succession ~150 m above the modern river consists of a 27 m lower part that aggraded around 1.5 Ma, followed by erosion, then deposition of the upper 4 m at ~66 0ka (MIS 16). This evidence (which, admittedly, is disputed; cf. Wolkowinsky & Granger, 2004, 2005) that the river returned to the same level at different times suggests a reversal in vertical crustal motion; the aggradation of this very thick deposit itself suggests subsidence, rather than uplift, in the Early Pleistocene. The relatively rapid rates of change in uplift rates indicated in the regional dataset, and the apparent reversal of incision in the Early Pleistocene, favour low W_i, which Bridgland & Westaway (2008b) deduced as 6 km; revised modelling shows that the dataset favoured in the present study for the immediate vicinity of Grand Junction is consistent with a similar W_i value.

Diverse explanations have been proposed for the Late Cenozoic uplift of the Colorado Plateau and the resulting incision by the Colorado River, including the development of the Grand Canyon (the online supplement of Levander *et al.*, 2011, provides a list; see Bridgland & Westaway, 2012, for a recent review; see also below). Opinions have indeed differed regarding whether or not the incision rate of the Grand Canyon has varied over time. Karlstrom *et al.* (2008) have argued, for example, that the available data support a constant rate since the Late Miocene (~6 Ma) of 0.19 mm a^{-1} in the eastern Grand Canyon, with ~800–600 m of incision since ~4–3 Ma; Karlstrom *et al*. (2012) likewise advocated a constant ~0.15 mm a^{-1} since ~10 Ma at Grand Junction. Levander *et al.* (2011) have attributed this large amount of uplift to a unique combination of mantle processes beneath the region, which has resulted in delamination of the mantle lithosphere and the relatively dense mafic part of the lower crust; the resulting increase in buoyancy of the remaining crust has led to uplift. However, they inferred that these processes were restricted to the region beneath the Grand Canyon, whereas essentially the same uplift has occurred around Grand Junction (Fig. 7), some 400 km farther upstream. Moreover, the Karlstrom *et al.* (2008) dataset is relatively sparse; although it allows incision histories with uniform incision rates, the 'gaps' between the data would also permit more variable rates, as we are now also suggesting for the Grand Junction area.

A seismic profile ~150 km east of the Grand Junction area indicates ~60 km thick crust with ~20km of mafic underplating (Karlstrom *et al.*, 2002). Gilbert and Sheehan (2004) reported the crust in the Grand Junction area as ~45 km thick and noted that there is limited acoustic impedance contrast at this level, which is consistent with a layer of mafic underplating at the base of the crust. However, Bueller & Shearer (2010) reported the crustal thickness near Grand Junction as between ~35 and ~40 km (say ~37 km) using seismic tomography. We take this as an estimate of the depth of the base of the mobile lower-crustal layer, irrespective of whether it is underlain by a layer of mafic underplating rather than the Moho. The surface heat-flow around Grand Junction is known from several boreholes (e.g., Sass *et al.*, 1971) and was reported as 66–69 mW m^{-2} by Eggleston & Reiter (1984) and depicted as ~65 mW m^{-2} by Wisian *et al.* (1999). On the other hand, Reiter (2008) reported a best estimate for the heat flow of 60 mW m^{-2}, based on Decker (1995) and Decker *et al*. (1988). The tomographic imagery by van der Lee & Nolet (1997) indicates a ~1300 °C temperature at 110 km, whereas the seismic investigations reported by Levander *et al.* (2011) place the base of the lithosphere (marking the ~1400 °C isotherm) at ~140 km in this area.

Modelling of the dataset summarized above, with W_i ~6 km and z_u ~37 km, suggests that the brittle upper-crust is ~30 km thick. Surface heat-flow of 65 mW m^{-2} can be modelled as occurring in basement with k=3.3 W m^{-1} °C^{-1}, Y=3.3mW m^{-3} and D=10km, placing the 350 °C isotherm (marking z_b) at ~29.9 km and the 1400 °C isotherm, representing the base of the lithosphere, at ~138 km; surface heat-flow of 60 mW m^{-2} would require less radioactive heat production in the upper crust to achieve equivalent agreement. The thermal state of the crust in this region is thus consistent with the model prediction of W_i.

Discussion

Summary of results

As noted previously (Bridgland & Westaway, 2008a,b; Westaway *et al.*, 2009a), significant reversals in vertical crustal motion are evident only in regions where modelling indicates that W_i is ~5 km or less, such as the Don (Fig. 3B) and the Green River (Fig. 4). Where W_i is ~5.5 or ~6 km, as in the Colorado (Fig. 7), one observes phases of minor subsidence (or near-stability) superimposed onto predominant uplift, indicating proximity to the threshold of the observed effect for smaller W_i. With W_i>6 km, as in the Dniester (Fig. 3a), South Platte (Fig 5) or many other examples worldwide (Bridgland & Westaway, 2008a,b; Westaway *et al.*, 2009a), monotonic uplift has occurred.

Such alternations of uplift and subsidence, observed in the East European Platform and the sites in the USA, have occurred in continental crust that consolidated during the Early or Middle Proterozoic, in which mafic underplating at the base of the crust constricts the overlying mobile layer, limiting its thickness to a few kilometres at most. Similar behaviour is also observed in younger crustal provinces with thick mafic underplating, such as parts of South America and the Arabian Platform in Syria and SE Turkey (e.g, Bridgland & Westaway, 2008b; Westaway, 2012). This behaviour contrasts with Archaean cratons, in which the combination of thick lithosphere and low radioactive heat production results in the absence of any mobile layer, leading to great stability and, as a result, the absence of terrace staircases (e.g., Westaway *et al.*, 2003) or minimal vertical movement, as observed in the lower Dnieper (Matoshko *et al.*, 2002; see above). The principal factor in determining this Late Cenozoic ultra-stability, the great thickness (up to ~400 km) that typifies cratonic lithosphere, is arguably a consequence of Archaean plate tectonics. Because the Earth's interior was hotter in the Archaean, magmatism at spreading centres produced oceanic crust that was much thicker (>20 km) and more buoyant than in more recent times and could not be subducted, so there was overthrusting of its upper part at convergent plate-margins, forming the 'greenstone belts' that characterize many Archaean terranes (e.g., Moores, 2002). The associated accumulation of residual oceanic lithosphere, minus its upper crust, at the base of the pre-existing continental lithosphere is thought to be the main process responsible for accumulating the great lithospheric thickness of Archaean cratons (e.g., Abbott, 1991; Kusky, 1993; Kusky & Polat, 1999).

It is evident that continental crust has been underplated during the Phanerozoic as a part of magnatism associated with mantle plumes, due to decompression melting of the mantle lithosphere during crustal extension in the presence of higher-than-usual ambient temperatures (e.g., White & McKenzie, 1989). However, it is debateable whether the same process can explain the greater thickness of mafic rocks typically present at the base of Early to Middle Proterozoic crust, even with potential repetition during the considerable timespan represented. Some authors (e.g., Durrheim & Mooney, 1994) have attributed the thick underplating in crust of this age to contemporaneous magmatic processes, including subduction and anorogenic heating of the crust, as well as rifting, all leading to major addition of crustal volume. The consolidation of continental crust in the Early–Middle Proterozoic is also typically accompanied by the widespread emplacement of granite intrusions (e.g., Andersen *et al.*, 2004), which can perhaps account for the high radioactive heat production estimated for the examples analysed in the present study. It has been suggested that the granitic/basaltic magmatism during the Late Proterozoic Brasiliano orogeny was associated with similar lithospheric differentiation to that which occurred earlier in the Proterozoic elsewhere (e.g., Neves *et al.*, 2000) and is, likewise, responsible for the observed mafic underplating. There is an alternative view, however, that sees Proterozoic magmatism as mainly associated with reworking of older continental crust rather than the accretion of new crust (e.g., Dhuime *et al.*, 2012; cf. Dewey & Windley, 1981).

Returning to consideration of the Quaternary fluvial datasets, it is concluded that, in regions with monotonic uplift, the timings of all phases of LCFF required to drive the model solutions relate to long-timescale climate change. Thus, the phase starting at 0.9 Ma can be attributed to climate deterioration around MIS 22, at the start of the 100 ka Milankovitch forcing (e.g., Mudelsee & Schulz, 1997) or the ‘Mid-Pleistocene Revolution’ (MPR), and thus to faster erosion in response to the more severe climate. Increases in fluvial incision rates at this time, attributed to faster uplift, noticed in central Europe decades ago (Kukla, 1975, 1977, 1978), have since been documented in many regions worldwide (e.g., Bridgland & Westaway, 2008a,b; Westaway *et al.*, 2009c), and are evident in this study in the Green River (Fig. 4), Platte (Fig. 5) and Colorado around Grand Junction (Fig. 7). However, farther upstream, the Colorado in Glenwood Canyon indicates a decrease in uplift rates at the MPR, attributed (Polyak *et al*., 2013) to decreasing erosion rates at this time in response to increased local aridity, thus demonstrating a counter-example to the trend usually evident at temperate latitudes while nonetheless indicating an effect of climate on uplift rates. An earlier phase of LCFF starting at ~2.0 Ma is observed in several systems, including the Dniester and Green River. Its start corresponds to the Tiglian C Stage of NW Europe, another time of marked climate deterioration (e.g., Pross & Klotz, 2002), equivalent to the Late Akchagyl Stage of eastern Europe (e.g., Matoshko *et al.*, 2004) and the early 'Nebraskan' stage of North America (e.g., Roy *et al.*, 2004). Cooling circa 2 Ma is also evident from global oxygen-isotope records (cf. Lisiecki & Raymo, 2005). The preceding phase of LCFF starting at ~3.1 Ma, also apparent in most of the systems, has been explained in terms of climate deterioration at the end of the Mid-Pliocene climatic optimum, which marked the start of upland glaciation in Europe (e.g., Westaway, 2001, 2002b,c; Westaway *et al.*, 2002; Bridgland & Westaway, 2008b). The earlier phase, starting around 6 Ma, observed (Fig. 3) in the Dniester and Don (and elsewhere in the Mediterranean region; Bridgland & Westaway, 2008b), corresponds to the Messinian salinity crisis, or rather its Pontian counterpart in the Black Sea, as already discussed.

In contrast, the timings of the interspersed reversals in vertical crustal motion vary between localities, suggesting that they result from differences in local conditions. Westaway (2012) has suggested that these reversals represent phases of deformation of the mantle lithosphere that were induced by the preceding phases of LCFF and the resulting crustal thickening and associated loading of the mantle lithosphere. They are thus only directly observable in regions where the mobile lower-crust is relatively thin, so its response to a phase of LCFF is correspondingly rapid, and the mantle lithosphere is relatively thick and stiff, so its response to the same phase of LCFF occurs significantly more slowly than that of the mobile lower-crustal layer. Alternations of uplift and subsidence have also been observed in fluvial sequences from the northern Arabian Platform, a region of Late Proterozoic crust with thick mafic underplating, from the rivers Euphrates, Tigris, and Orontes, and support the deductions made in the present study regarding the relation between this effect, crustal and lithospheric properties, and climatic forcing (e.g., Demir *et al.*, 2007, 2008, 2012; Bridgland *et al.*, 2012; Westaway, 2012).

By undertaking more detailed calculations of the thermal state of the continental crust than have previously been attempted for any of the discussed study regions, our analysis confirms that the values of W_i deduced from previous uplift modelling studies reflect the true, or at any rate plausible, estimates of the thickness of the mobile layer in each case. This strengthens the argument that alternations of uplift and subsidence over the timescales indicated are, indeed, only observed in regions where this mobile layer is thin (≤7 km thick) but not absent.

Uncertainties in geothermal calculations

Notwithstanding the plausibility of the above findings, it is apparent that the geothermal calculations presented are subject to some uncertainty. In most cases, with the exception of the Ukrainian Shield, we have been unable to find published information regarding the radioactive heat-production in the uppermost crustal basement in our study localities. The depth distribution of this heat-production is also not directly constrained. We have therefore chosen values in each case that are consistent with the available evidence, which typically comprises measurements of surface heat-flow, the depth of the base of the mobile layer, and the lithospheric thickness. The observed surface heat-flow has thus been partitioned into 'basal' and 'radiogenic' components, the former in rough inverse proportion to the lithospheric thickness. In each case we have assumed that the radioactive heat production Y is constant down to a depth D below the top of the basement. An alternative to the latter assumption would be to assume that the heat-production decreases exponentially downwards, taking D to represent the range of depth z over which Y decreases to 1/e of its surface value, i.e.,

$$Y(z) = Y_o \exp(-z/D) \tag{1}$$

where Y_o=Y(z=0). This results in the following temperature distribution if the exponential decrease in heat production is assumed to persist indefinitely:

$$T(z) = T_o + \frac{q_o z}{k} + \frac{Y_o D^2}{k} (1 - \exp(-z/D)) \tag{2}$$

where q_o is the basal heat-flow and k the thermal conductivity (e.g., Lachenbruch, 1970). The surface heat-flow q_s is thus

$$q_s = q_o + Y_o D . \tag{3}$$

Alternatively, as in the present study, one may assume that the radioactive heat production is constant, at $Y=Y_o$, across a depth range from zero to D and zero at greater depths where the basal heat-flow is again q_o. If so, simple algebra indicates that

$$q(z) = q_o + Y_o (D–z) \qquad 0 \le z \le D , \tag{4}$$

so q_s is again given by equation (3). It follows from equation (4) that

$$T(z) = \frac{q_o\, z}{k} + \frac{Y_o\, D\, z}{k} - \frac{Y_o\, z^2}{2\, k} + T_o \quad 0 \leq z \leq D\,. \qquad (6)$$

The temperature at depth D is thus T_D where

$$T_D = \frac{q_o\, D}{k} + \frac{D^2\, Y_o}{2\, k} + T_o, \qquad (7)$$

the first two terms being the contributions from the basal heat-flow and the radioactive heating. At depths z>D, the basal heat-flow therefore causes the following temperature distribution:

$$T(z) = \frac{q_o\, (z-D)}{k} + T_D \qquad z>D\,. \qquad (8)$$

Equations (7) and (8) have been used for many temperature calculations throughout this study.

It is thus apparent that, for the same parameter values, the two assumptions about the distribution of radioactive heat production predict identical values for the surface heat-flow; however, as Fig. 8A shows, they predict different geotherms. Because it places some of the heat production deeper within the crust (i.e., it locates a proportion of 1/e of it deeper than z=D, rather than none of it), the exponentially-decaying assumption results in higher geothermal gradients in much of the upper-crust and, therefore, significantly higher temperatures at mid-crustal depths. However, for the uppermost few kilometres (say, for $z<D/3$), roughly the depth-range directly accessible to borehole heat-flow measurements, the two assumptions predict almost identical geotherms (Fig. 8B).

For the data used to generate Fig. 8 (Q_o=40 mW m^{-2}, Y_o=5 μW m^{-3}, D=10 km, T_o=10 °C, and k=3 W m^{-1} °C^{-1}, values chosen to represent the crust of northern England (which has been intruded by highly radiogenic Palaeozoic granites) rather than any of the localities discussed in the present study, the assumption of constant heat production above depth D predicts z_b (the 350 °C isotherm) at 19.25 km and the base of the lithosphere (the 1400 °C isotherm) at 98 km. In contrast, the exponentially decaying assumption predicts these depths as 15.62 km and 91.75 km. These two assumptions can therefore be distinguished by their different predictions of the conditions at depth. Thus, if the lithospheric thickness is known to be ~98 km, and the geotherm were to be modelled with exponentially-decaying heat production, a different combination of values for q_o, Y_o, D, and k would have to be chosen, which would result in the predictions for z_b (and thus for the thickness of the lower crust) being close to the value calculated for the original choice of uniform radioactive heat production. The available constraints are, therefore, sufficient to enable reliable estimates of the thickness of the mobile layer to be made using geothermal evidence.

Alternative interpretations

The above-mentioned pattern, whereby reversals in vertical crustal motion occur in regions where the mobile lower-crust is thin, only became apparent following examination of the global long-timescale fluvial dataset (Bridgland & Westaway, 2008a,b); it had previously been overlooked due to a paucity of systematic study and lack of widespread comparison. These instances demonstrating reversals in vertical crustal motion were indeed previously each thought to be unique, and were thus explained in terms of local effects that each only work for particular localities. For instance, Granger *et al.* (2001) suggested that the alternation of base-level lowering and rising, evident in Fig. 4, has been caused by a succession of diversions of the Ohio River (of which the Green River is a tributary) by Pleistocene glaciations. The diverse past explanations of the uplift of the Colorado Plateau have already been noted (see above): buoyant 'hot mantle blobs' (e.g., Pazzaglia *et al.*, 1999; Moucha *et al.*, 2009) or rebound following removal of part of the mantle lithosphere (e.g., Spencer, 1996; Levander *et al.*, 2011). The scenario proposed by Levander *et al.* (2011), for example, attributes the uplift occurring in the Colorado Plateau to a unique combination of processes that have been made possible by this region's particular magmatic and tectonic history, rather than (as we are suggesting) a general process, namely erosional isostasy mediated by induced lower-crustal flow, that has been occurring in many regions worldwide. Others (e.g., Pederson *et al.*, 2002) have interpreted the uplift of the Colorado Plateau as the isostatic response to regional erosion, but have not recognized its connection with patterns of LCFF caused by climate change. Until recently, Late Cenozoic uplift in general was likewise interpreted using arbitrary explanations. For instance, it has traditionally been explained in some parts of Europe by plate motions and in others by mantle plumes. However, the similarity in uplift histories in all areas of dynamic crust suggests that a single general process predominates.

Diverse explanations have likewise been proposed for the Late Cenozoic uplift of the Colorado Plateau and the resulting incision by the Colorado River, including the development of the Grand Canyon (the online supplement of Levander *et al.*, 2011, provides a list; see Bridgland & Westaway, 2012, for a recent review; see also below). Opinions have indeed differed regarding whether or not the incision rate of the Grand Canyon has varied over time. Karlstrom *et al.* (2008) have argued, for example, that the available data support a constant rate since the Late Miocene (~6 Ma) of 0.19 mm a^{-1} in the eastern Grand Canyon, with ~800–600 m of incision since ~4–3 Ma; Karlstrom et al. (2012) likewise advocated a constant ~0.15 mm a^{-1} since ~10 Ma at Grand Junction. Levander *et al.* (2011) have attributed the uplift in the eastern Grand Canyon to a unique combination of mantle processes beneath the region, which has resulted in delamination of the mantle lithosphere and the relatively dense mafic part of the lower crust; the resulting increase in buoyancy of the remaining crust has led to uplift. However, they inferred that this combination of processes has been restricted to the region beneath the Grand Canyon, whereas Pederson *et al*. (2013) have shown by dating of Colorado terraces that the uplift is centred upstream of the Grand Canyon, in a region of readily erodible lithologies and no known complex mantle processes, thus emphasizing the role of erosional isostasy. Moreover, the Karlstrom *et al.* (2008) dataset for the Grand Canyon is relatively sparse; although it allows incision histories with uniform incision rates, the 'gaps' between the data would also permit more variable rates, as we are now suggesting for the Grand Junction area. Setting aside local effects of course changes, Karlstrom *et al.* (2012) and Donohue *et al*. (2013) attribute the uplift in the Grand Junction area to a combination of erosional isostasy and effects of mantle buoyancy, relatively hot, low-density sub-lithospheric mantle being indicated in this region by seismic tomography. However, they offer no evidence that the thermal state of the mantle beneath this region has changed over time, as would be necessary to generate any transient effect. Our own geothermal calculations thus assume steady-state thermal conditions, and although we have taken into account the lateral variations in lithospheric thickness indicated by the tomography we have made the approximation that the asthenosphere temperature does not vary laterally. Likewise, Karlstrom *et al.* (2012) and Donohue *et al*. (2013) have not considered the possibility that lower-crustal flow might be induced by the pressure variations at depth caused by the evident lateral variations in erosion, or that erosion rates in any locality have varied over time in response to climatic forcing. There is no modelling technique that can explore all the various effects simultaneously; i.e., including the complex perturbations that develop to the thermal state of the crust as a result of the induced lower-crustal flow, and which act as a regulating feedback mechanism (Westaway, 2002a, 2007a), and the complexity of the deformation to the mantle lithosphere that may result (Westaway, 2012), as well as effects of independent pre-existing lateral variations in the thickness and thermal state of the continental crust and of independent thermal and buoyancy effects within the sub-lithospheric mantle. We therefore draw the principal conclusion apparent from the empirical evidence, which is that the observed abrupt variations in uplift rates indicated by the fluvial incision, and the evidence for reversals between uplift and subsidence, are consistent with a thin mobile lower-crustal layer. We estimate this to be ~7 km thick, which is consistent with a global pattern reflecting other fluvial datasets.

There has also been reluctance on the part of some workers to accept river terraces as records of uplift. For example, Hancock & Anderson (2002) suggested that a fluvial terrace staircase might develop in the absence of uplift as a result of systematic changes to the hydrology brought about by climate change since the start of the Pleistocene, causing the river to develop a more strongly concave-upward longitudinal profile in its middle reach. Such behaviour can be predicted on theoretical grounds (see below), but no known pattern of climate change throughout the Pleistocene can achieve the required effect (the climate in any given region having been very similar during equivalent parts of successive Pleistocene climate cycles, rather than becoming systematically wetter from one climate cycle to the next as this theoretical mechanism would require). Others have argued on theoretical grounds that uplift can be recognized within fluvial systems by the development of knickpoints, formed in response to increasing uplift rates, which propagate upstream. This point of view, despite criticism and a lack of fit to the known behaviour of fluvial systems (cf. Bridgland & Westaway, 2012), is highly engrained; for example, Gallen *et al.* (2013) have argued that uplift is occurring in the Appalachian Mountains of the eastern USA because knickpoints are present in some valleys. We would support their conclusion, but not the underlying argument; in our view a much better argument for uplift of the Appalachians is that major rivers within this mountain belt, such as the Susquehanna and upper Ohio (discussed above; cf. Westaway, 2007b) have produced long-timescale terrace staircases, showing that erosional isostasy has been occurring here, with positive-feedback enhancement from lower-crustal flow. Wobus *et al.* (2010) have discussed the rivers in the Great Plains of the central USA (including the Platte, also considered above), in the light of the above-mentioned theoretical concepts, looking for evidence of systematic changes in longitudinal gradient: either increases, which (in their view) would indicate that uplift had caused the observed fluvial incision, or decreases, which would indicate that the incision has resulted from a significant change in hydrology. They favoured the latter explanation, contradicting McMillan *et al.* (2002), who noted evidence that the Ogallala Group sediments have become tilted eastward as a clear indication of laterally-varying uplift. Duller *et al.* (2012) argued that although the Ogallala Group has clearly been tilted eastward, the younger Broadwater Formation has not. However, in the vicinity of Fig. 6B they determined 1.5 m km^{-1} as the modern longitudinal gradients of the North Platte River and 1.9 m km^{-1} as the present-day slope of the base of the Broadwater Formation. Their palaeohydrological calculations indicated that the depositional slope of the latter was 1.8 m km^{-1}. However, the uncertainties in the latter calculation spanned 0.8 to 3.4 m km^{-1}; this large margin of uncertainty makes it impossible to establish on this basis whether the longitudinal gradient of the river has remained constant or not. If sustained over 200 km distance, this 0.4 m km^{-1} difference between the tilt of the sediments and the modern river gradient would result in an ~80 m difference in height of the Broadwater Formation relative to the river, roughly equivalent to the difference in post-0.9 Ma incision indicated between Figs 6B and 6C. In any case, Duller *et al.* (2012) reported no evidence of any change to the longitudinal gradient of the river on the timescale of development of the fluvial terrace staircase in Fig. 6; until such a change is demonstrated we consider it reasonable to assume no change, such that incision can be taken as a proxy for uplift. Wobus *et al.* (2010) also remarked that the fluvial incision in this region would have isostatic consequences, which their analysis did not consider, and would itself create a component of down-to-the-east tilt. We would indeed support this conclusion; explanation by erosional isostasy, with positive feedback from LCFF, obviates any need for either an independent ‘tectonic’ cause of the uplift (cf. McMillan *et al.*, 2002;

Nereson *et al*., 2013) or a hypothetical long-timescale change in the river gradient caused by a change in hydrology (cf. Wobus *et al.*, 2010), neither of which has any supporting evidence.

Conclusions

Reversals in Late Cenozoic vertical crustal motion, involving alternations between uplift and subsidence, have been identified in fluvial sequences worldwide. Examples include the Don and Dnieper in eastern Europe and tributaries of the Mississippi/Missouri in the central USA. It has previously been suggested on the basis of uplift modelling that this behaviour occurs in regions where the mobile lower-crustal layer is thin, with $W_i \leq 6$ km. The present analysis shows that geothermal calculations support this conclusion; the phenomenon occurs in regions where the mobile layer is ≤~7 km thick, consistent with $W_i \leq 6$km. Future development of understanding of this phenomenon should therefore concentrate on its link to the presence of a thin mobile layer and the associated stiffness and thickness of the underlying mafic underplating and mantle lithosphere. This study underlines how detailed analysis of fluvial archives can contribute to the understanding of properties of the continental crust, thereby explaining differences in landscape evolution in different regions. It also emphasizes the value of widespread comparison of data, as enabled by organizations such as FLAG, rather than localized studies of particular systems leading to results that are interpreted in isolation.

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Figure Captions

Figure 1. Map of crustal provinces in Eastern Europe, adapted from Fig. 1 of Kutas *et al.* (1979), with additional information from Nikishin *et al.* (1996). Dashed line indicates the border of the former Soviet Union. TTZ denotes the Tornqvist-Teisseyre Zone. A-B and C-D indicate the lines of the geothermal profiles across Poland and western Ukraine in Fig. 2, which cross the TTZ. E-F locates the section line through the Late Cenozoiic deposits of the River Dnieper discussed by Matoshko *et al.* (2004), which indicates reversals in vertical crustal motion (Fig. 3B). K and M denote the Karpinsky Swell and the Mangyshlak Uplift, regions of particularly intense Hercynian deformation within the Scythian and Turan platforms. R and W denote the Rostov-on-Don Anticlinorium and the West Siberian Platform. The Donbas Foldbelt originated as the SE continuation of the Devonian Pripyat-Dnieper-Donets rift, but experienced crustal shortening during the Variscan orogeny, when the most recent major deformation in the adjacent Rostov-on-Don Anticlinorium is also thought to have occurred. The Pre-Caspian Block represents crustal basement of unknown origin (e.g., Nikishin *et al.*, 1996), which is thought to have accreted aganst the SE margin of the East European Platform in the Late Proterozoic; drilling has reached latest Proterozoic (Vendian) sediments but not the underlying

basement. The crust in this region is clearly strong, since no significant deformation has occurred throughout the Phanerozoic. It has been suggested that it is a trapped fragment of oceanic lithosphere (cf. Şengör *et al.*, 1993); it may possibly be an aseismic ridge or oceanic plateau whose buoyancy has prevented subduction.

Figure 2. Profiles indicating the variations in crustal thickness and surface heat flow across the Tornqvist-Teysseire Zone. **A.** In Poland, from Fig. 6 of Majorowicz & Plewa (1979). **B.** In western Ukraine, from Fig. 5 of Kutas *et al.* (1979). Shading denotes sediment overlying crustal basement. See Fig. 1 for locations.

Figure 3 Idealized composite transverse profiles through fluvial sequences in Eastern Europe, showing MIS correlations (after Matoshko *et al.*, 2004). **A.** The middle to lower Dniester in the Ukraine-Moldova border region, representing the reach of the river from Serebriya in Fig. 1 southeastward to the Black Sea coast. **B.** The Don in the vicinity of Voronezh, Russia. **C.** The Dnieper in central Ukraine. A much more detailed cross-section, also showing locations of the boreholes on which part C is based, was published by Matoshko *et al.* (2002). Angled arrows in A and B denote the timings of increases in the rate, or reversals in the sense, of vertical crustal motion. These are recognised in the local literature (e.g., Matoshko *et al.*, 2004), and correspond with the starts of phases of LCFF in the uplift modelling of these fluvial sequences by Bridgland and Westaway (2008a). Shaded angled arrows indicate phases that are readily explicable in terms of variations in rates of lower-crustal flow induced by surface processes, in response to long-timescale climate change. Open arrows indicate phases that are not readily explicable in this manner, but can be tentatively explained (after Westaway, 2012) as consequences of relaxation of the mantle lithosphere in response to earlier phases of induced lower-crustal flow. Numbers in B indicate the sequence order of aggradation of the fluvial deposits.

Figure 4. Evidence of change in the level of the Green River in Kentucky, U.S.A., as determined by cosmogenic dating of fluvial sediments in the Mammoth Caves karstic system. Adapted from Fig. 6 of Granger *et al.* (2001), which shows more detail including the distribution of dates used to constrain the interpretation. Angled arrows have the same meaning as in Fig. 3; they are only shown for the most significant variations in rate or sense of vertical crustal motion, and correspond to the starts of phases of LCFF identified in uplift modelling by Westaway (2007b).

Figure 5. Representative transverse profile across the terrace staircase of the South Platte River near Denver, Colorado, U.S.A.. Modified from Osterkamp *et al.* (1987), with age constraints from many other sources, as discussed by Bridgland & Westaway (2008b). Angled arrows have the same meaning as in Fig. 3; they correspond to the starts of phases of LCFF at 3.1 and 0.9 Ma identified in uplift modelling by Bridgland & Westaway (2008a). In addition, another phase of uplift may have also occurred starting at ~2 Ma, but is not well-resolved; if present, it is minor compared with the other two phases depicted.

Figure 6. Variation in Late Cenozoic fluvial geomorphology within the Platte River system in Nebraska and adjoining states (after Reed *et al.*, 1965, as modified by Bridgland & Westaway, 2008a, who list other sources of information). **A.** Location map. **B-F.** A sequence of sections through the local fluvial sequences ordered generally from west to east. Deposits are labelled with the names and age assignments from Reed *et al.* (1965), although many of these are no longer used in the local literature, with some updating from the more recent literature (Condon, 2005; Duller *et al.*, 2012; and references cited therein). The principal age constraint is provided by the presence in the 'Kansan' Sappa Loess of ash that has been correlated with the Lava Creek B eruption of Yellowstone, now dated to 602±4 ka (MIS 15) in the vicinity of its inferred source (Gansecki *et al.*, 1998) or to 639±2 ka (MIS 16) by Lanphere *et al.* (2002). Balco *et al.* (2005) have obtained cosmogenic ages for what appears to be stratigraphically equivalent loess, of 580±120 ka (MIS 15 or early MIS 14), and for underlying till, presumed equivalent to the 'Kansan' Cedar Bluffs till, of 650±140 ka (MIS 16). The 'Pearlette Ash' in C and E has been confirmed as from the ~600 ka eruption of Yellowstone (Izett & Wilcox, 1982); that in D has not been, although this is probable as few occurrences of older Yellowstone tephras are known in Nebraska. The current classification scheme for the 'Nebraskan' sequence (Roy *et al.*, 2004) recognizes an 'R2' group of Early Matuyama reverse-magnetized tills that pre-date the Huckleberry Ridge eruption of Yellowstone, at 2003±14 ka (Gansecki *et al.*, 1998) or 2059±4 ka (Lanphere *et al.*, 2002), and include the Elk Creek till in D; and an 'R1' group of Late Matuyama reverse-magnetized tills that post-date the Mesa Falls eruption, at 1293±12 ka (Gansecki *et al.*, 1998) or 1285±4 ka (Lanphere *et al.*, 2002), and include the Iowa Point till in E and F.

Many of the 'Nebraskan' deposits in the region are thus inferred to be late Early Pleistocene, such that a phase of fluvial incision, inferred to have begun at ~0.9 Ma, can be identified in B and C. Angled arrows have the same meaning as in Fig. 3; they are only shown in B, where monotonic uplift since the Mid/Late Pliocene is apparent, and in C. Other localities farther east show more complex patterns, indicative of reversals in the vertical crustal motion on the same timescale.

Figure 7. Idealised composite transverse profile through the terrace staircase of the Colorado River and its major tributary, the Gunnison, in the vicinity of Grand Junction, Colorado. Modified from Fig. 11 of Bridgland & Westaway (2008b), which lists sources of data (the numerical ages shown being from Ar-Ar dating), ornamented to indicate sites that have contributed data. As is discussed in the text, these sites are now known to have experienced different histories of fluvial incision; the post-6 Ma incision histories in the Grand Junction and Glenwood Canyon areas can serve as proxies for uplift. Numbers in circles are interpreted MIS correlations, from Bridgland & Westaway (2008b), for the Gunnison terrace staircase in the vicinity of Delta, Colorado (cf. Donahue *et al*., 2013). Note phases of major incision interpreted in the Late Miocene (1), in the Late Pliocene (2) and starting in the early Middle Pleistocene (3). Angled arrows have the same meaning as in Fig. 3; they correspond to the starts of the principal phases of LCFF identified in uplift modelling by Bridgland & Westaway (2008b).

Figure 8. Predicted geotherms for exponentially-decaying and uniform distributions of heat production with depth, for the same crustal parameters. All graphs assume

q_o=40mW m^{-2}, Y_o=5μW m^{-3}, D=10km, T_o=10°C, and k=3W m^{-1} °C^{-1}, but each is calculated using the appropriate equations for the assumed distribution of heat production, as discussed in the text.

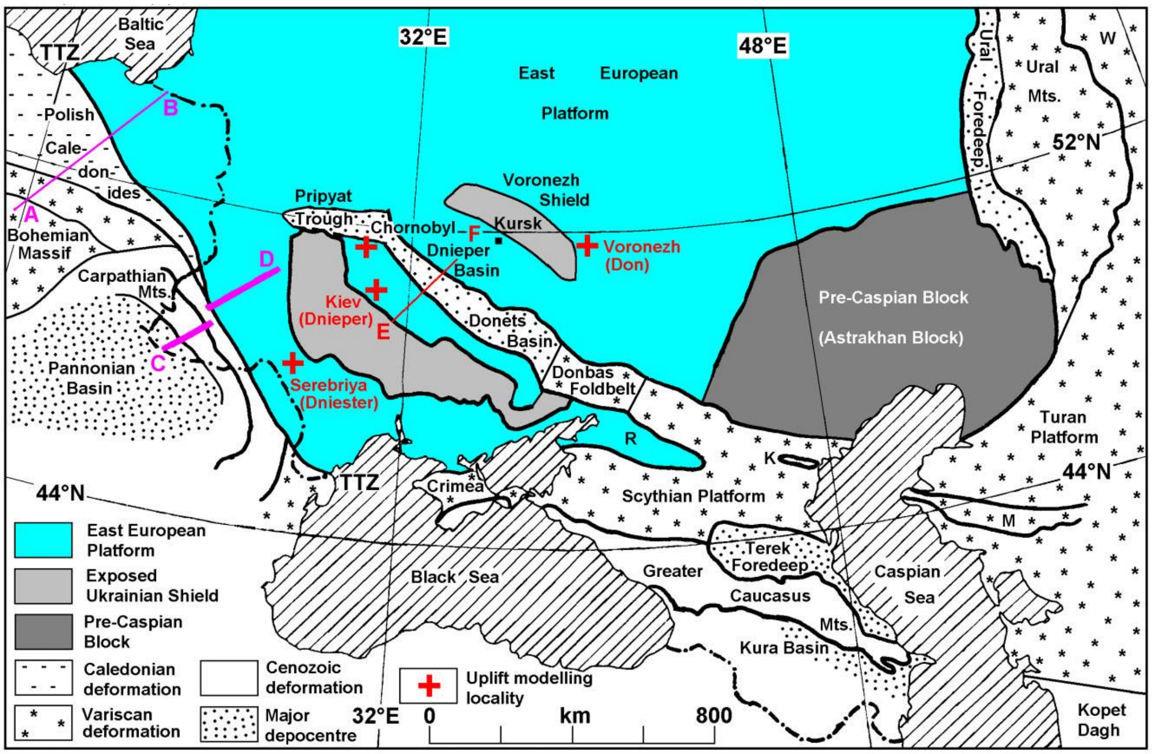
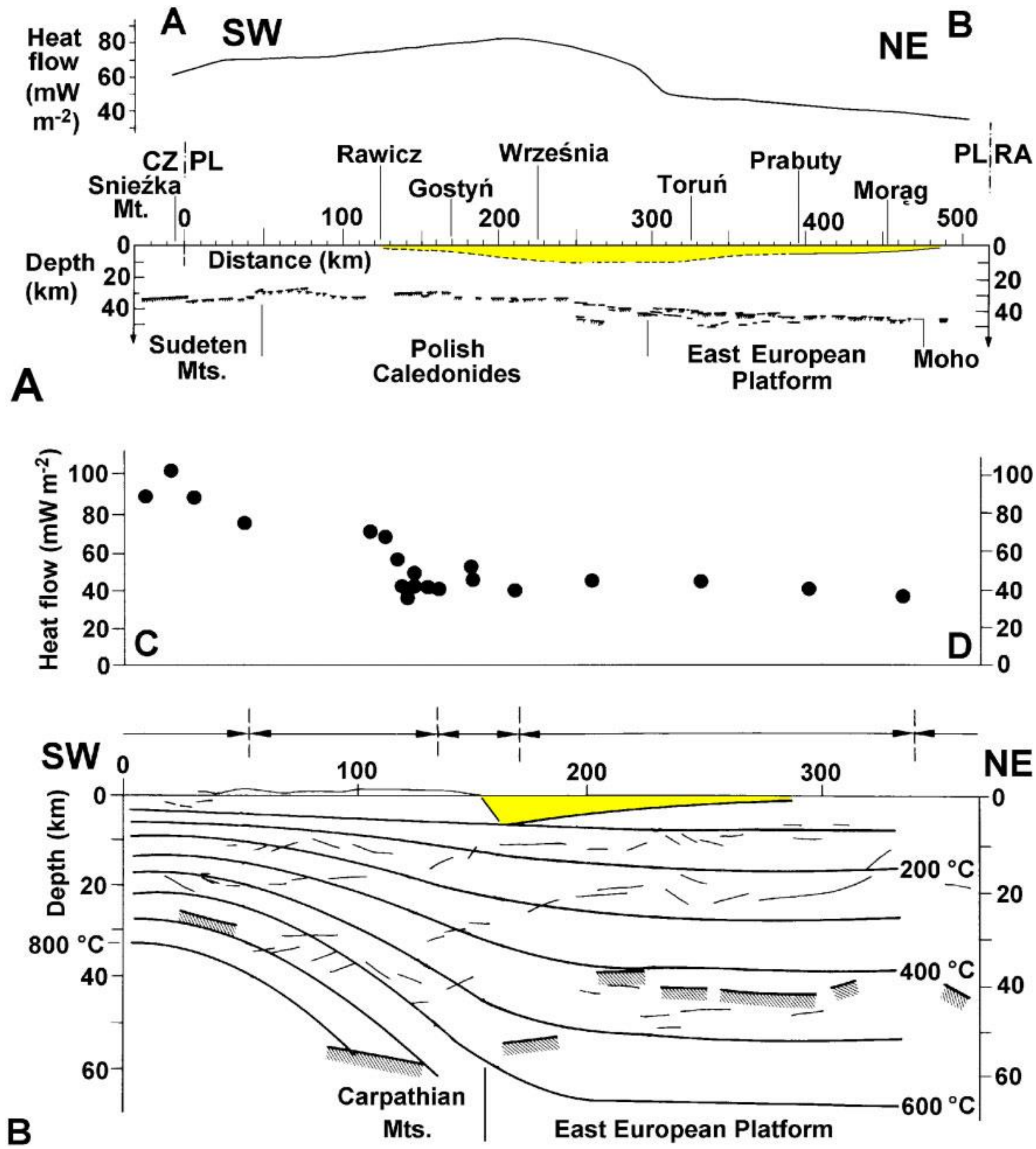


Figure 1



A
SW
NE
B
Heat flow (mW m-2)
CZ PL
Sneżka Mt.
Rawicz
Gostyń
Września
Toruń
Prabuty
Morąg
PL RA
Depth (km)
Distance (km)
Sudeten Mts.
Polish Caledonides
East European Platform
Moho
A
Heat flow (mW m-2)
C
D
SW
NE
Depth (km)
800 °C
200 °C
400 °C
600 °C
Carpathian Mts.
East European Platform
B

Figure 2

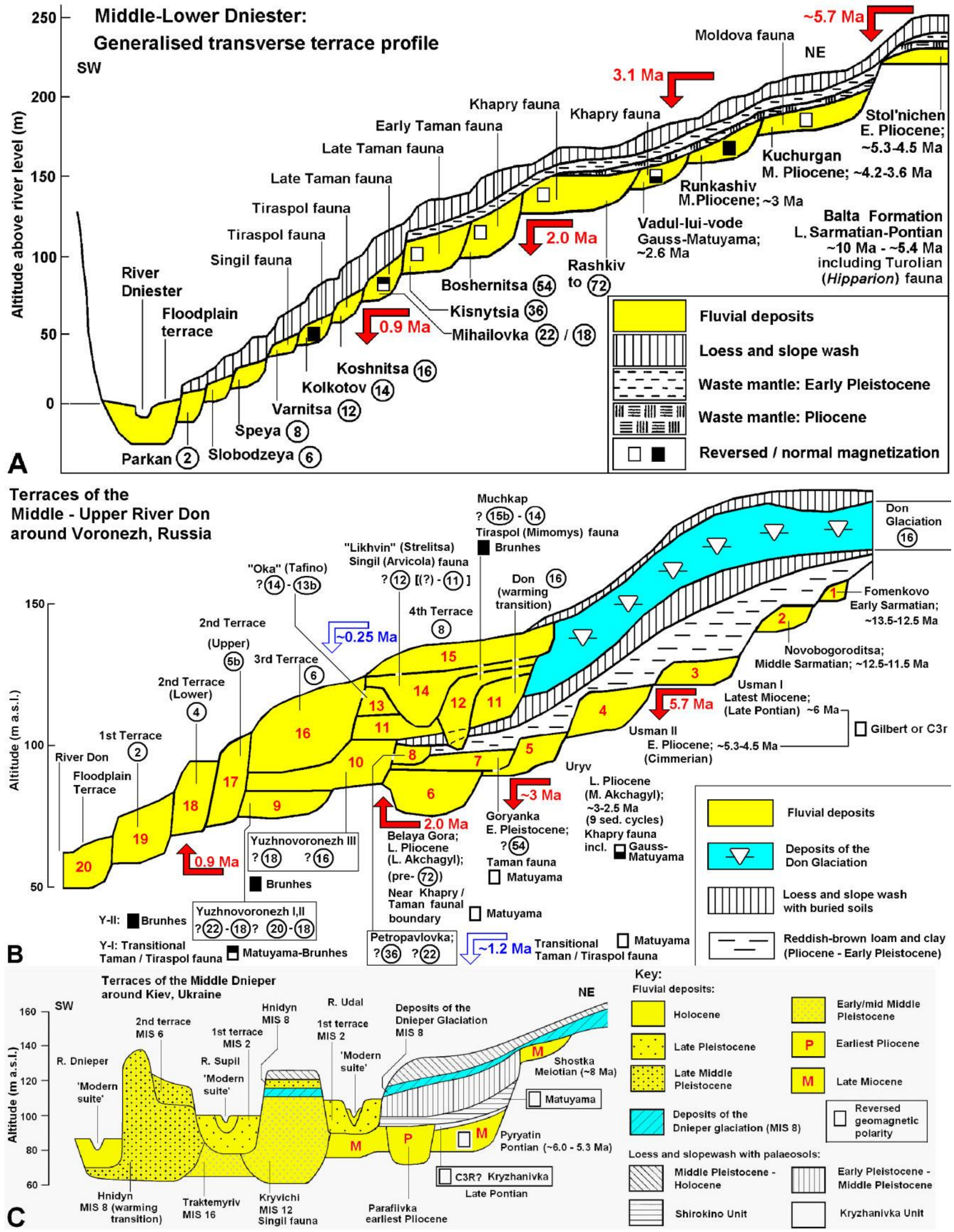

Middle-Lower Dniester:
Generalised transverse terrace profile
SW
NE
Altitude above river level (m)
River Dniester
Floodplain terrace
Parkan (2)
Slobodzeya (6)
Speya (8)
Varnitsa (12)
Kolkotov (14)
Koshnitsa (16)
Mihailovka (22) / (18)
Kisnytsia (36)
Boshernitsa (54)
Rashkiv to (72)
Vadul-lui-vode Gauss-Matuyama; ~2.6 Ma
Runkashiv M.Pliocene; ~3 Ma
Kuchurgan M. Pliocene; ~4.2-3.6 Ma
Stol'nichen E. Pliocene; ~5.3-4.5 Ma
Balta Formation L. Sarmatian-Pontian ~10 Ma - ~5.4 Ma including Turolian (Hipparion) fauna
Singil fauna
Tiraspol fauna
Late Taman fauna
Early Taman fauna
Khapry fauna
Moldova fauna
0.9 Ma
2.0 Ma
3.1 Ma
~5.7 Ma
Fluvial deposits
Loess and slope wash
Waste mantle: Early Pleistocene
Waste mantle: Pliocene
Reversed / normal magnetization
A
Terraces of the Middle - Upper River Don around Voronezh, Russia
Altitude (m a.s.l.)
River Don
Floodplain Terrace
1st Terrace (2)
2nd Terrace (Lower) (4)
2nd Terrace (Upper) (5b)
3rd Terrace (6)
4th Terrace (8)
"Oka" (Tafino) ?(14)-(13b)
"Likhvin" (Strelitsa) Singil (Arvicola) fauna ?(12) [(?)-(11)]
Muchkap ? (15b) - (14) Tiraspol (Mimomys) fauna Brunhes
Don (16) (warming transition)
Don Glaciation (16)
Fomenkovo Early Sarmatian; ~13.5-12.5 Ma
Novobogoroditsa; Middle Sarmatian; ~12.5-11.5 Ma
Usman I Latest Miocene; (Late Pontian) ~6 Ma
Usman II E. Pliocene; ~5.3-4.5 Ma (Cimmerian)
Gilbert or C3r
Uryv L. Pliocene (M. Akchagyl); ~3-2.5 Ma (9 sed. cycles) Khapry fauna incl. Gauss-Matuyama
Goryanka E. Pleistocene; ?(54) Taman fauna Matuyama
Belaya Gora; L. Pliocene (L. Akchagyl); (pre-(72)) Near Khapry / Taman faunal boundary Matuyama
Petropavlovka; ?(36) ?(22)
Transitional Taman / Tiraspol fauna Matuyama
Yuzhnovoronezh III ?(18) ?(16) Brunhes
Yuzhnovoronezh I,II ?(22)-(18)? (20)-(18)
Y-II: Brunhes
Y-I: Transitional Taman / Tiraspol fauna Matuyama-Brunhes
~0.25 Ma
0.9 Ma
2.0 Ma
~3 Ma
5.7 Ma
~1.2 Ma
Fluvial deposits
Deposits of the Don Glaciation
Loess and slope wash with buried soils
Reddish-brown loam and clay (Pliocene - Early Pleistocene)
B
Terraces of the Middle Dnieper around Kiev, Ukraine
SW
NE
Altitude (m a.s.l.)
R. Dnieper
'Modern suite'
2nd terrace MIS 6
1st terrace MIS 2
R. Supil
Hnidyn MIS 8
R. Udal
1st terrace MIS 2
Deposits of the Dnieper Glaciation MIS 8
Shostka Meiotian (~8 Ma)
Matuyama
Pyryatin Pontian (~6.0 - 5.3 Ma)
C3R? Kryzhanivka Late Pontian
Hnidyn MIS 8 (warming transition)
Traktemyriv MIS 16
Kryvichi MIS 12 Singil fauna
Parafiivka earliest Pliocene
Key:
Fluvial deposits:
Holocene
Late Pleistocene
Late Middle Pleistocene
Deposits of the Dnieper glaciation (MIS 8)
Early/mid Middle Pleistocene
Earliest Pliocene
Late Miocene
Reversed geomagnetic polarity
Loess and slopewash with palaeosols:
Middle Pleistocene - Holocene
Early Pleistocene - Middle Pleistocene
Shirokino Unit
Kryzhanivka Unit
C

Figure 3

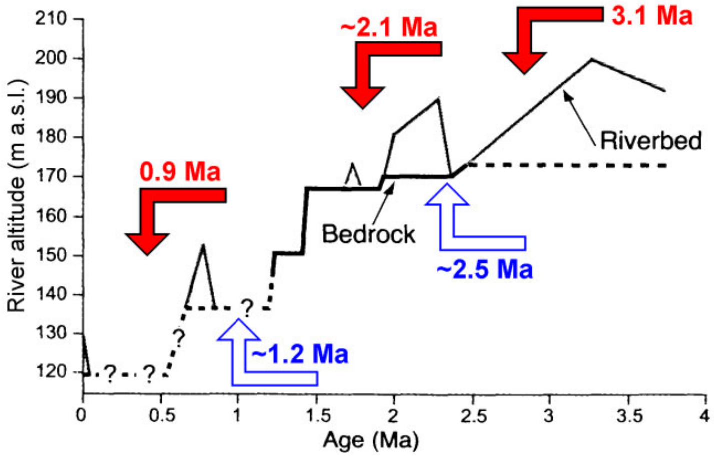


Figure 4

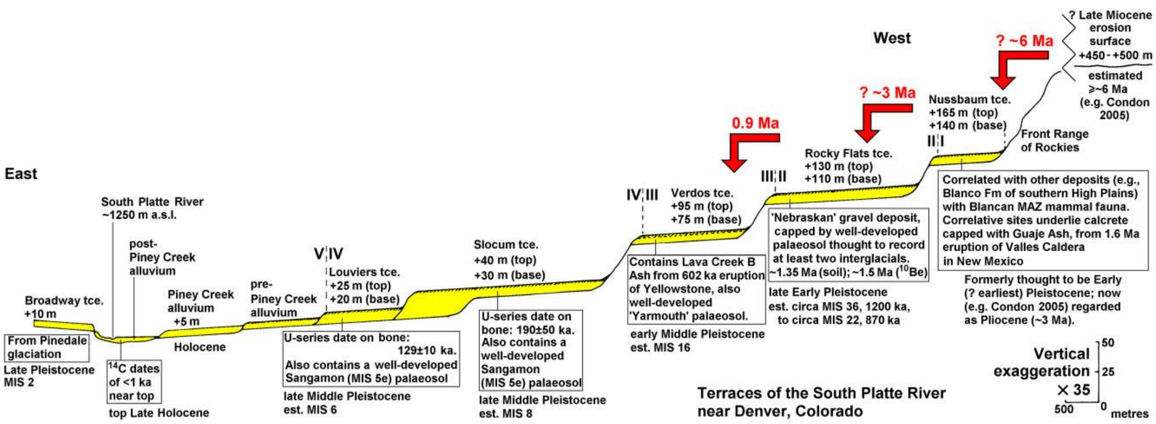
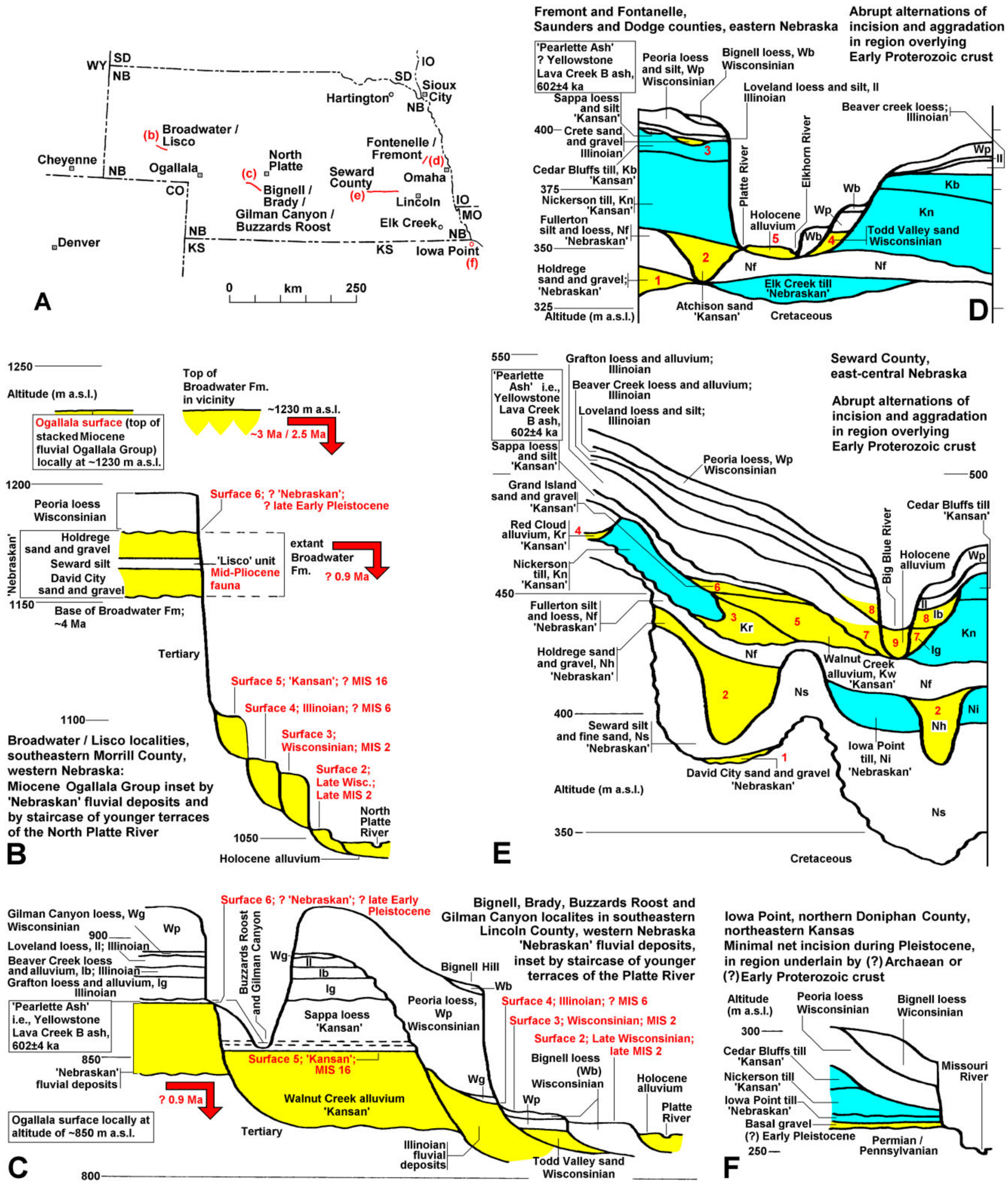


Figure 5



Fremont and Fontanelle, Saunders and Dodge counties, eastern Nebraska
Abrupt alternations of incision and aggradation in region overlying Early Proterozoic crust
Seward County, east-central Nebraska
Abrupt alternations of incision and aggradation in region overlying Early Proterozoic crust
Broadwater / Lisco localities, southeastern Morrill County, western Nebraska: Miocene Ogallala Group inset by 'Nebraskan' fluvial deposits and by staircase of younger terraces of the North Platte River
Bignell, Brady, Buzzards Roost and Gilman Canyon localites in southeastern Lincoln County, western Nebraska 'Nebraskan' fluvial deposits, inset by staircase of younger terraces of the Platte River
Iowa Point, northern Doniphan County, northeastern Kansas
Minimal net incision during Pleistocene, in region underlain by (?) Archaean or (?)Early Proterozoic crust
'Pearlette Ash' i.e., Yellowstone Lava Creek B ash, 602±4 ka
Ogallala surface (top of stacked Miocene fluvial Ogallala Group) locally at ~1230 m a.s.l.
Surface 6; ? 'Nebraskan'; ? late Early Pleistocene
Surface 5; 'Kansan'; ? MIS 16
Surface 4; Illinoian; ? MIS 6
Surface 3; Wisconsinian; MIS 2
Surface 2; Late Wisc.; Late MIS 2
Ogallala surface locally at altitude of ~850 m a.s.l.
Altitude (m a.s.l.)
A
B
C
D
E
F

Figure 6

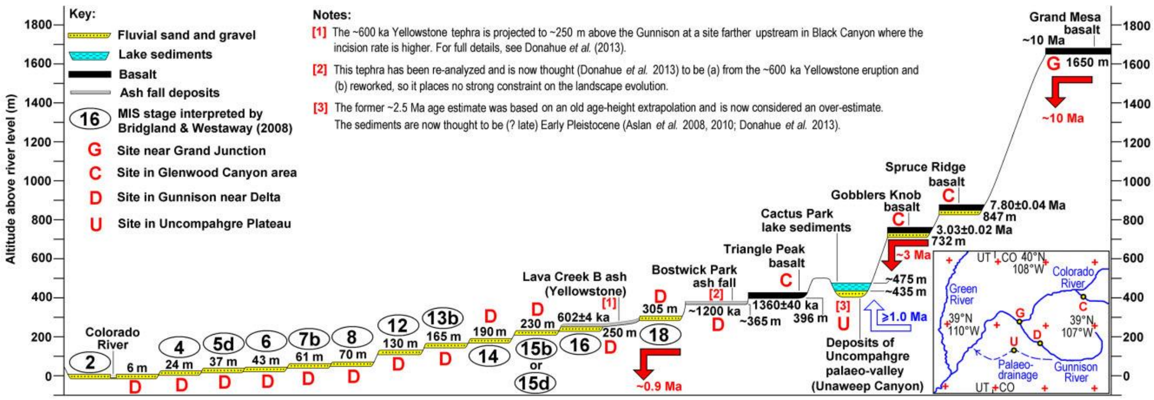


Figure 7

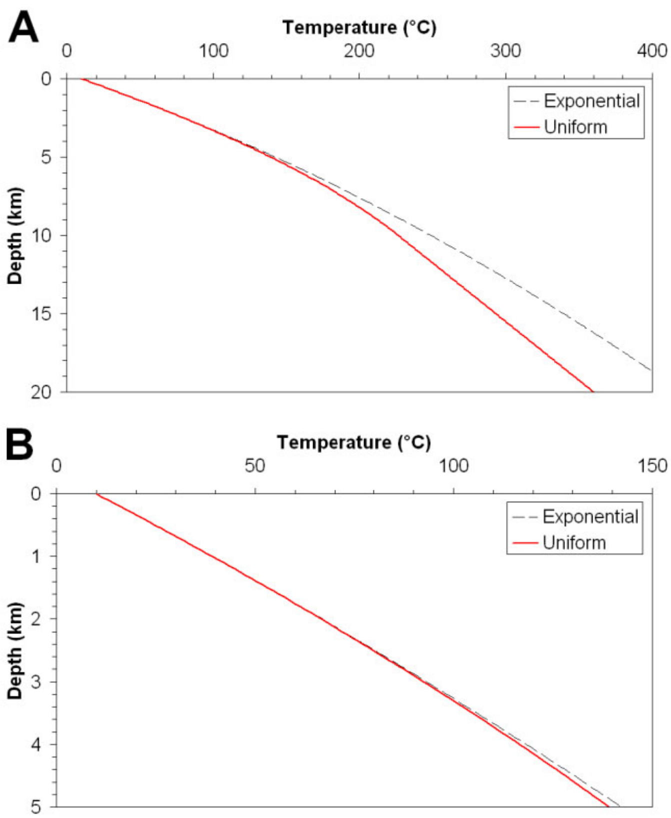


Figure 8