Basin inversion: a worldwide Late Cenozoic phenomenon

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Abstract

The occurrence of endorheic basins on the Tibetan Plateau, both in its Pleistocene history and (fewer in number) at the present day, has been attributed to the 'basin-and-range' character of the orogen; the understanding of their conversion to exorheic drainage is key to interpreting the evolution of the Yellow River and other river systems of the plateau. However, such basins also occur in areas of lower altitude and relief and can be observed to have been considerably more common in pre-Quaternary times. In many areas, for example the Mediterranean region, such basins, previously having accumulated stacked sedimentary sequences, typically 'inverted' in the late Pliocene or at around the Pliocene–Pleistocene boundary, possibly as part of a response to the cooling of global climate and its effect on surface processes. Some basins have inverted later, at around the time of the Mid-Pleistocene Revolution, coinciding with the increased severity of climate resulting from the 100 ka Milankovich cycles that followed that change. The progressive incision into the fills of these inverted sedimentary basins has also been linked to this climatic influence, perhaps taking effect as a result of erosional isostatic uplift, which would have replaced the accumulation-induced subsidence (also isostatic) of the basins. NW Europe, including Britain, had sediment-accumulating basins in the Early Cenozoic; the timing of their inversion is poorly known as a result of the paucity of preserved evidence from the Late Cenozoic in such areas. Endorheic basins that survive at the present day are mainly in regions of relative aridity and are often controlled by active tectonic structures, such as the pull-apart basins of the Dead Sea Fault Zone and similar examples. This review discusses the evidence from different climatic regions, tectonic settings and areas of different crustal type, with a view to elucidating meaningful patterns that might throw light on this phenomenon.

Keywords: sedimentary basin; subsidence; endorheic; Tibetan Plateau; Yellow River

1 Introduction

Much of the precipitation on the NE Tibetan Plateau, the location of the field excursion within the meeting from which this special issue arises, fails to reach the global ocean, instead being intercepted by endorheic drainage-basin systems; the Qinghai Lake Basin, for example, drains internally to a lake surface at ~3198 m above sea level. This can readily be explained by the 'basin-and-range' structure imposed upon the plateau by the nature of the tectonism that has created it, a series of elongate basins of various sizes separated by mountain ranges formed from resistant rocks that have become involved in the deformation (e.g., Tapponnier et al., 2001; Fig. 1). Indeed, intermontane basins, often endorheic, are a typical feature of tectonically uplifted plateaux (Strecker et al., 1989; Meybeck et al., 2001; Sobel et al., 2003; Pingel et al., 2013; Ballato et al., 2019). Not all the basins of the NE Tibetan Plateau are endorheic at the present day, however. Many have been 'captured' by drainage to the ocean, in particular by the Yellow River (Huang He), often (but not always) leading to 'inversion' of the basins, with a change from sediment accumulation to fluvial incision into the basin-fill sequences. The Yellow River drains a succession of basins from its upper reaches to its lower course across the North China Plain: the Zaling–Eling, Zoige, Xinghai–Tongde, GongHe, GuiDe, Xunhua, Longzhong, Zhongwei–Zhongning, Yinchuan, Hetao, and Fenwei basins (Fig. 1). The fact that different basins show different stages of dissection is perhaps key to understanding the evolution of the Yellow River drainage system, a topic that has been debated for many years (Li, 1991; Westaway, 2002; Craddock et al., 2010; Hu et al., 2011; Pan et al., 2011; Hu et al., 2016, 2017; Zhang et al., 2017).

Endorheic river systems (i.e., drainage that fails to reach the ocean) are in the considerable minority, globally, covering <20% of the Earth's surface (Maybeck et al., 2001; Dorsaz et al., 2013). Some of these are rivers that dry up as they flow across desert areas, such as, in western China, the Yukong Kashi and Karakax and in the Hexi Corridor, the Heihe, Beida and Shule rivers (e.g., Hetzel et al., 2004; He et al., 2015; Fig. 1); the courses of such rivers, last occupied by drainage during wetter episodes in the past, might still exist under the desert sands, with the potential for reoccupation should humidity increase in future. The Sahara is well known to have once hosted extensive drainage systems, particularly in more humid pre-Quaternary times, these rivers having become inactive as aridity increased (Pickford, 2000; Griffin, 2002; Goudie, 2005). Similarly, former drainage lines can be detected beneath the sands of the Kalahari (Moore and Larkin, 2001; Burrough et al., 2009). Many (if not most) endorheic rivers drain into a lake or wetland at the centre of a basin, where the water level fluctuates according to the balance between evaporation and renewed supply (Tyler et al., 2006). The Lake Eyre system, in south-central Australia, is one of the largest such endorheic basins at the present day (e.g., Kotwicki and Isdale, 1991; Rust and Nanson, 1986; Habeck-Fardy and Nanson, 2014; Figs 2 and 3). It contrasts in several ways with Lake Qinghai. First, it is dry, or mostly dry, for much of the time, with its surface a vast salt-pan, and, second, it is close to sea level, with its lowest central areas (and the sediment sequence beneath the lake floor) falling >15 m below sea level. Both these lakes have excellent sedimentary sequences. Lake Eyre has >400 m of sediments that record past conditions, generally a change from a more humid Neogene environment to increasing aridity in the Quaternary (Alley, 1998; Habeck-Fardy and Nanson, 2014), although this intra-cratonic basin originated as a result of Late Cretaceous – Early Cenozoic tectonic subsidence (Krieg et al., 1990). Lake Qinghai would appear to have formed considerably more recently, perhaps in connection with the onset of the East Asian monsoon, by ~4.6 Ma, although its sedimentary fill exceeds 600 m in thickness and provides an important record of fluctuating climatic and hydrological conditions, particularly the interplay between the influences of the monsoon and the westerlies (An et al., 2012; Fu et al., 2013; Innes, 2017). Similar systems occur in other dry parts of the world, the aridity potentially being a factor in their continued existence and perhaps their initiation, particularly in combination with rapid uplift (Sobel et al., 2003; Heidarzadeh et al., 2017). Indeed, it has been suggested that the aridity caused by the precipitation-blocking effect of bounding ridges, by leading to endorheism and therefore the retention of sediment in intermontaine basins, is a positive-feedback factor in the formation of high plateaux (Garcia-Castellanos, 2007).

Examples of endorheic basins and drainage systems that are arid but, like Lake Eyre, not of high altitude include the Aral Sea, infamously shrunken in recent decades as a result of anthropogenic interference (e.g., Oberhänsli et al., 2007), Lake Chad (Griffin, 2002) and the Okovango (Gumbricht et al., 2001), which ends in an inland swamp, although in a more humid past the large palaeo-lake Makgadikgadi existed there (Shaw, 1988; Burrough et al., 2009; Podgorski et al., 2013). Such basins, since they are accumulating sediment (albeit relatively slowly, in many cases), will generally be subsiding isostatically under the load of that sediment, a process that provides additional accommodation space for further sedimentation; this activity has been confirmed for the Okovango delta, for example, by Gumbricht et al. (2001). Should sediment supply exceed the generation of new accommodation space, then the basin will fill and drainage will spill out of it, this being one potential mechanism for the capture and inversion of formerly endorheic basins (Douglass and Schmeeckle, 2007; Cunha et al., 2020 [X ref within this special issue, not yet published as hard copy]). The overflow of a basin could also occur if water level were to rise during a wetter climatic phase (e.g. Heidarzadeh et al., 2017), as again is envisaged for Iberia (see Cunha et al., 2020). This idea is contrary, and an important alternative, to the common interpretation of fluvial capture, triggered by headward erosion, as the mechanism for endorheic to exoreheic transformation (e.g., Craddock et al., 2010).

Modern-day fluvially-fed subsiding basins are not invariably endorheic. There are examples in which rivers flow into areas of sediment accumulation and subsidence but continue to outlets into the global ocean. This might be facilitated by these subsiding basins coinciding with the lowermost reaches of rivers, although this is by no means always the case. It is axiomatic that the stacks of sediment in such basins contrast markedly with the terraced sedimentary records of rivers in uplifting areas upstream and/or downstream (e.g., Maddy, 1997; Bridgland, 2000; Antoine et al., 2007; Cunha et al., 2012; cf. Bridgland and Westaway, 2008a, b, 2014). Equally clearly, it is the former that stand the better chance of becoming part of the long-term geological record, with the existing pre-Quaternary fluvial record invariably consisting of deposits that accumulated in basinal depocentres (cf. Bridgland et al., 2014). Compilation of fluvial-archive data from around the world, part of the activity of the early Fluvial Archives Group (FLAG) and carried out under the auspices of what is now the International Geoscience Programme (IGCP), revealed patterns in the style of preservation (Bridgland et al., 2007; Westaway et al., 2009b) that suggest that subsiding fluvially-fed basins were more common in the pre-Quaternary than at present and that there has been progressive capture and inversion of basins during the past few million years. This is more difficult to establish than the the figure of <20% for modern endoreheic drainage, but the prevalence of Neogene endorheic systems in Mediterranean regions such as Turkey and Iberia, where the evidence from the Late Cenozoic is generally well preserved, contrasts with the diminished proportion of internally draining basins at present in those regions (see, e.g., Demir et al. (2018) and later discussion). This paper will review examples of such evidence and assess the underlying patterns, seeking explanations where possible. A further issue, in the light of the inspiration behind this present special issue, will be to consider whether the basins of the NE Tibetan Plateau and other orogenic areas fit within the observed patterns, or whether their high-altitude and tectonic contexts separate them from other categories (see Table 1 for a summary of selected basins, organized by type).

2. Modern-day subsiding fluvial depocentre (sag) basins

There are several well-known examples of currently subsiding basins in which Quaternary sediments have accumulated as stacked sequences, including multiple reaches of both of Europe’s largest two rivers, the Rhine and Danube. The Lower Rhine has been cited many times as an example in which equivalents of the upstream terrace sequence are to be found within the >70 m of accumulated sediment, in a conventional stratigraphical (‘layer-cake’) sequence, that thickens progressively downstream of the German–Dutch border region (Brunnacker et al., 1982; Ruegg, 1994; Bridgland and D'Olier, 1995; Bridgland, 2000; Bridgland and Westaway, 2008a, b, 2014, Busschers et al., 2007; Fig. 4). Although this basin has always remained connected to the ocean (except perhaps during ice-blockage of the North Sea: Gibbard, 1995; Gupta et al., 2007), it is closed-ended, with the bedrock surface rising significantly downstream, beneath the sea floor of the Dover Strait (Bridgland and D'Olier, 1995), a phenomenon that can be accounted for, presumably, by the centering of the subsiding depocentre in the area beneath the Netherlands and immediately offshore from the Rhine Delta. Prior to the Elsterian/Anglian glaciation, which coincided with Marine oxygen Isotope Stage (MIS) 12, the Dover Strait did not yet exist and the Rhine fed a much more extensive basin, beneath an enlarged delta, in the southern North Sea; however, this former depocentre has since been bypassed by (cold-climate / low-sea-level) drainage via the English Channel (e.g., Gibbard, 1995; Westaway and Bridgland, 2010; Westaway, 2017).

Upstream, within Germany, are smaller fault-bounded crustal blocks that show relative subsidence in comparison with the more widespread area of the Middle Rhine Graben, an example being the Neuwied Basin (Bosinski, 1995; Meyer and Stets, 2002). The subsidence here is relative, in that there are Rhine terraces within the Neuwied Basin, albeit at lower heights than in nearby areas beyond its bounding faults, indicating that the basin has been uplifting during the Quaternary, but at a much reduced rate in comparison with the surrounding region (e.g., Cordier et al., 2009). In its earlier Cenozoic history, however, this was a subsiding basin that filled with marine and then terrestrial deposits (Boenigk and Frechen, 2006; Table 1).

Moving to the Danube, that river has also been building stacked sedimentary sequences in depocentres, in particular the Pannonian Basin, beneath the plains of Hungary (Ruszkiczay-Rüdiger et al., 2005; Gábris and Nádor, 2007), and at the edge of the Black Sea (Matoshko et al., 2009). As in the Rhine, in reaches where there is transition between uplifting and subsiding areas, the terrace staircase of the Danube becomes constricted in altitudinal range and then grades into a stacked sequence, as is first encountered in a downstream direction in the Danube Basin, giving rise to the 'Little Hungarian Plain' (Fig. 5). Downstream from here the subterranean fluvial archives of the Little Hungarian Plain gives way to a terrace staircase through the Transdanubian (Hungarian) mountain range, before this in turn gives way to the larger Great Hungarian Plain, beneath which are >700 m of accumulated Danube sediments (Ruszkiczay-Rüdiger et al., 2005; Gábris and Nádor, 2007; Fig. 5). The Pannonian Basin is an example of a formerly endorheic system that has not been inverted, in the sense that subsidence has continued: internal flow into the erstwhile 'Lake Pannon' gave way to an alluvial plain as the lake became filled, with the Danube maintaining external drainage once that had been initiated (Gábris and Nádor, 2007). The change to exorheic drainage occurred around the Miocene–Pliocene transition, or roughly at the time of the Messinian salinity crisis in the Mediterranean region; attempts have been made, using modelling, to determine the climatic and/or tectonic factors and mechanism responsible (e.g., Csato et al., 2007; Leever et al., 2011; Suc et al., 2011; Enciu et al., 2014), with some authors (e.g., Leever et al., 2011) attributing it to lake overflow and others (e.g., Suc et al., 2011) to river capture.

A further example of a river that passes through successive uplifting and subsiding reaches, with a comparable effect on the pattern of fluvial-archive preservation, is the Orontes, which flows northwards from Lebanon, through Syria, to Turkey, largely along the Dead Sea Fault Zone (DSFZ: Fig. 6). This river flows through two subsiding pull-apart basins within the DSFZ in which there has been long-term sediment accumulation, probably from pre-Quaternary times: the Ghab Basin in northern Syria and the Amik Basin in Hatay Province, southern Turkey. Characterized, like the plains of Hungary, by flat surfaces relatively little above river level, these are interposed between terraced and gorge reaches that have been uplifting (Bridgland and Westaway, 2008a, 2014; Bridgland et al., 2012; Fig. 6; Table 1).

Subsiding fluvially-fed basins are commonly present in the foreland areas adjacent to major tectonically generated mountain ranges (Beaumont, 1981; Flemings and Jordan, 1990), such as the Himalayas (e.g., Tandon, 1976; Molnar, 1984; Sinha et al., 2007) and the Alps (Beaumont, 1981). In the former case, the Ganges system drains much of the basinal area that formed in response to the large crustal load of the evolving Himalayas and sediment-load derived from the nearby upland (Tandon, 1976, 1991; Tandon and Rangaraj, 1979;

Lyon-Caen and Molnar, 1985). The basin infill, in the form of the Neogene Siwalik Group and its successors, represents a >4 km thick fluvial record dating back ~13 Ma (Keller et al. 1977; Sinha et al., 2007). An important Alpine-foreland example of a river flowing almost entirely across a stack of its own sediments, with no apparent incision into its valley-floor, is the Po in northern Italy, which flows across the subsiding Po Basin (Po Plain) to the northern Adriatic Sea (Ori, 1993; Table 1). Continued tectonic activity can divide foreland basins, leading them to become incorporated in orogenic upland areas (e.g., Ballato et al., 2008; DeCelles et al., 2011; Pingel et al., 2013) as, indeed, is considered to be the origin of some of the high basins of the Andes (Strecker et al., 1989; Kleinert and Strecker, 2001).

2.1 Fluvially-fed sedimentary basins on the NE Tibetan Plateau

The basins on the Tibetan Plateau that are probably subsiding under sediment load are essentially those that remain unconnected to global ocean, such as the Qinghai Lake Basin, which, as already noted, is endorheic at the present day, although it might once have had an outflow to the Yellow River (Pan, 1994; Kang et al., 2017). Indeed, it was perhaps once a sub-basin of a much larger lacustrine depocentre that also incorporated the now exorheic Xinghai, GongHe, GuiDe and Xining basins (Fig. 1), a view based on potential stratigraphical continuity between the fills of these basins, as well as coincident 'culminant surface' levels of the basin fills and reconstructed maximum lake levels (e.g., Pan, 1994; Perrineau et al., 2011; cf. Liu et al., 2013). The eastern end of the GongHe Basin is drained by the Yellow River but the major part of that basin is separated from this drainage system and remains endorheic, draining to small lacustrine centres in the west, particularly Lakes Chaka, Dalian and Duolonggou, the last having the same water level as Lake Qinghai (Perrineau et al., 2011; Zhang et al., 2012, 2014). This is perhaps because the basin was only joined to the Yellow River in relatively (geologically) recent times (cf. Li, 1991; Pan, 1994) and drainage evolution has not yet led to capture of other parts of the basin; indeed, cosmogenic dating (^{10}Be and ^{26}Al) by Perrineau et al. (2011) has suggested that incision into the GongHe Basin infill began as recently as ~150 ka. In contrast, the infill of the GuiDe Basin, immediately downstream, has become incised by ~1000 m since ~1.8 Ma, the overall age span of the resulting succession of river terraces being securely dated by magnetostratigraphy (Fang et al., 2005). These two basins are separated by Mesozoic granodiorite bedrock of the Waligong Range, the Yellow River having cut the Longyang Gorge, 500 m deep, through this barrier; the granodiorite barrier has, however, been exhumed from beneath connecting basin sediments, the GongHe and GuiDe sedimentary fills having been connected in their upper parts, and there has been uplift of the barrier from movement on a reverse fault (Fang et al., 2005; Perrineau et al., 2011). From the above-mentioned cosmogenic dating, this gorge is implied to have formed with surprising rapidity, the required incision into the resistant bedrock being up to 6 mm per year (cf. Bridgland and Hu, 2017). The GuiDe, Xunhua and Xining basins were joined during the early part of their (predominantly lacustrine) infill history and subsequently separated by tectonic movements before the final filling of the disassociated endorheic GuiDe and Xunhua sub-basins by fluvial conglomerates, 50–150 m thick, during the Early Pleistocene (Liu et al., 2013). Liu et al. (2013) described these sub-basins as overfilled by this time and noted that the subsequent evolution was deposition of Middle and Upper Pleistocene sediments within the newly incised Yellow River valley (i.e., terrace deposits), placing the timing of inversion after 1.7 Ma.

In a similar vein, the uppermost Yellow River flows out of the Zaling–Eling Basin and through the Zoige Basin (Zhu et al., 2009), with only the latter showing inversion (Table 1). Thus far, the Yellow River has eroded only a few metres into the Zoige basin-fill, perhaps reflecting its connection to the river system in recent geological times (Pan, 1991). There are several sub-basins, parts of which would appear to be internally draining, with rivers, not yet captured by the Yellow River, feeding into lacustrine and wetland centres.

3. Examples of fluvial depocentre basins from the recent geological past, now inverted

The inverted Neogene basins of the NE Tibetan plateau have many parallels elsewhere in the world, albeit not created from such extreme tectonic activity. Small structural basins, similar to the Ghab and the Amik basins of the Orontes system except that they have inverted, occur in other parts of the Mediterranean region, notably Iberia (Harvey and Wells, 1987; Visera and Fernandez, 1992; Mather, 1993; Mather et al., 1995; Wenzens and Wenzens, 1997; Stokes and Mather, 2000, 2003; García-Meléndez et al., 2003) and Turkey (Westaway, 1993; Westaway et al., 2004, 2006a; Maddy et al., 2005, 2007, 2008, 2012, 2017). In Iberia, rather larger Neogene basins, such as the Madrid Basin, once existed in the interior of the peninsula, becoming inverted when connected to the Atlantic by the formation of rivers such as the Tagus, Duero and Miño (e.g., Pérez-González, 1994; Cunha et al., 2008, 2016, 2020; Pais et al., 2012; Gouveia et al., 2020); understanding how the capture of the interior endorheic basin(s) of central Iberia came about is fundamental to explaining Late Cenozoic landscape evolution over a much wider area (Cunha et al., 2020 [this issue]).

Neogene subsiding basins also existed close to the Atlantic coast of Iberia, such as the Lower Tagus (Lisbon) Basin (e.g. Pais et al., 2012; Cunha, 2019), formed by Cenozoic lithospheric folding and faulting (de Vicente et al., 2011, 2018). It was filled successively by marine, lacustrine and finally fluvial sediment, the last representing an ancestral Tagus system that was receiving drainage from the previously endorheic Madrid Cenozoic Basin as early as ~3.65 Ma (Cunha, 1992; Pérez-González, 1994). The incision stage of this basin is believed (from modelling) to have commenced by 1.8 Ma (Cunha et al., 2012, 2016, 2017).

In western Turkey, around the Kula volcanic complex, the River Gediz has incised a terrace system (Maddy et al., 2012; Fig. 7) into >300 m of poorly consolidated sediments filling the Selendi Basin (Table 1), comprising the predominantly fluvial Hacıbekir and İnay groups, the latter culminating in lacustrine limestones (Seyitoğlu, 1997; Purvis and Robertson, 2004, 2005; Ersoy et al., 2010). This is one of a number of similar extensional basins that accounted for ~50% of the surface area of central western Turkey in the Late Cenozoic, of which the Selendi sediments are probably the best exposed and most extensively researched; these include the Gördes, Demirci and Uşak–Güre basins and, further south, the somewhat larger Alaşehir and Büyük–Menderes grabens (Purvis and Robertson, 2005). Whether basin infill continued into the Pliocene is a topic of debate (see Westaway et al., 2004 for discussion) but inversion had occurred by the Early Pleistocene, as closely-spaced gravel terraces are preserved beneath basalt dated by Ar–Ar to ~1320 ka and have recently been attributed to MIS 48–38 (Maddy et al., 2017). Bridgland and Westaway (2014) suggested that this and similar occurrences of Late Cenozoic basin inversion were an indirect result of global cooling, related to the acceleration of surface processes this caused and the resultant enhancement of erosional isostatic uplift.

Further east in Turkey, two substantial endorheic Late Neogene palaeolakes existed before the establishment of the upper catchment of the Euphrates. The first of these was the Adiyaman palaeolake, which existed between ~6.0 and ~3.7–3.6 Ma, impounded by localized tectonic (hanging-wall) uplift on the northern flank of the Bozova Anticline, which led to blockage of the Euphrates valley near the site of the modern-day Atatürk Dam (Demir et al., 2012). The drainage of this lake basin probably resulted from a decrease in the slip rate on the fault, although lake overflow cannot be ruled out as a mechanism. Further upstream in the Euphrates system, the Malatya palaeolake formerly existed in the area of that city, elongated N–S on its western side, where it extended along the Malatya–Ovacık Fault Zone, satellite system to the East Anatolia Fault Zone (Seyrek et al., 2008; Westaway et al., 2008). Like that at Adiyaman, this large Miocene lake basin appears to have been disrupted by tectonic movements, although in both cases there is arguably enhancement of uplift and lake disruption as a response to climatic change, through the mechanism of erosional isostasy. In addition to numerous smaller endorheic basins that existed in the Pliocene – Early Pleistocene, there was another substantial fault-generated palaeolake in the Diyarbakır area (coincident with the Diyarbakır Basin), within the present catchment of the River Tigris; its inversion was certainly pre-Quaternary and probably pre-Pliocene (see Westaway et al., 2009a).

Long-lived endorheic basins persist in Turkey, however. Two obvious examples are those occupied the largest lakes in the country, Lake Van and Lake Tuz. Lake Van, in eastern Anatolia, is a high-level (1648 m above sea level) highly alkaline (pH up to 9.8) and saline lake in a tectonically generated basin; the world's largest soda lake by volume, it has accumulated sediments that provide a valuable climatic record from the last ~600,000 years (Litt and Anselmetti, 2014; Stockhecke et al., 2014a, b). It is thought to have formerly been exorheic, with an outlet to the south-west that was blocked as a result of volcanic activity during the formation of the large stratovolcano 'Nemrut', located ~15 km to the west, probably during the Early Pleistocene (Degens et al., 1984). Although subidence is now confined to the internally draining lake basin, this sits above larger infilled and linked Cenozoic Hinis–Mus–Van basins (Bati and Sancay, 2007). Turkey's second-largest lake, also saline and endorheic, is Lake Tuz, situated in central Anotolia, 105 km NE of Konya. It is located in another structurally generated subsiding basin, larger than that at Lake Van (Çemen et al., 1999; Aydemir and Ates, 2006; Fig. 2).

Moving north-eastwards from the Mediterranean region, there is evidence for inversion of the western part of the extended Black Sea Basin at around the Miocene–Pliocene transition, to judge from the record from the particularly well developed Dniester terrace sequence, as documented by Matoshko et al. (2004; Fig. 8). The highest terraces within this sequence, formed by the Kuchurgan, Runkashiv, Vadul-lui-Vode and Rashkiv sedimentary suites, are incised progressively below the surface of a Miocene stacked fluvial sequence, the Balta Series, within which as many as 11 depositional cycles, potentially climatically driven, have been distinguished (Bilinkis, 1992). The Balta Series is capped, with no apparent incision, by the Stol'nichen Series, attributed to the Lower Pliocene, the dating of this and the underlying Miocene deposits being achieved using mammalian faunas, with *Hipparion* (Matoshko et al., 2004). This places the inversion of the basin in which the Balta Series accumulated as a mid-Pliocene event, assuming that the Stol'nichen Series is the culminant division of the basin-fill sequence

Further to the Lake Eyre example that was contrasted with Lake Qinghai in China, in Eastern Australia the Murray–Darling, the largest river system of that arid country, originated in the Early Cenozoic within a basin that had much older origins and which has accumulated several hundreds of metres of marine, lacustrine and fluvial sediments. It shows little indication of inversion, despite localized uplift and resultant diversion of river channels during the last climate cycle and an incised reach through Cenozoic limestone in its lowermost course (Bowler, 1978; Stephenson and Brown, 1989; Glen, 1992), where epeirogenic uplift has been invoked, based on coastal aeolianite sediments dated using the thermoluminescence (TL) and amino-acid methods (Murray-Wallace et al., 2010). In contrast, the Shoalhaven River, which drains to the Tasman Sea ~150 km south of Sydney and has a Quaternary terrace record that extends back to around the Early–Middle Pleistocene transition (Nott, 1992; Nott et al., 2002; Bridgland and Westaway, 2008a, 2014), flows above Cenozoic valley fills (Nott, 1992; Fig. 9). In this case, therefore, there is clear evidence for Pleistocene landscape inversion prior to the formation of the Pleistocene valley (Nott et al., 2002; Fig. 9C), perhaps attributable to the same post-Miocene and/or accelerating Middle–Late Pleistocene uplift seen in other parts of the world (e.g., Bridgland and Westaway, 2008a, 2014). During the Neogene the Shoalhaven was disrupted by basaltic eruptions that produced lava dams and led to episodes of lacustrine deposition in the palaeovalleys, which were effectively narrow, linear basins. The dating of the Shoalhaven terraces (Nott et al., 2002; Fig. 9C), however, points to most of the incision of the modern valley into the Cenozoic fill having occurred since the Mid-Pleistocene Revolution (MPR), perhaps in relation to the greater severity of glacials that has characterized the longer climatic cycles that began at that time (cf. Bridgland and Westaway, 2014).

Relatively little attention has been paid thus far to the Americas in this review, notwithstanding their numerous and in some cases well-known endorheic systems, such as the Great Salt Lake/Lake Bonneville and the Salton Sea, USA (Fig. 2). Parts of the western US make for valuable comparison with the Tibetan Plateau, however, because of structural and drainage similarities. For example, in SE California an ancestral (Pleistocene) river system that interconnected lakes in several Basin and Range Province valleys, including Death Valley (Table 1), was traced by Blackwelder (1954). This river had a catchment of ~100,000 km^2 and probably debouched into the Colorado. Thus, a large region that is now desert, with only isolated local and endorheic drainage systems, might well have drained to the Pacific Ocean. Recent studies (e.g., Jayko et al., 2008; Knott et al., 2008) have demonstrated that this drainage has recurred several times since the Mid-Pliocene. Further NE, Lake Bonneville once occupied much of Utah and parts of eastern Nevada and southern Idaho: an area of >50,000 km^2. Its depth was up to ~300 m and during highstands it overflowed northward into the Snake River and thence via the Columbia to the Pacific Ocean (e.g., Eardley et al., 1973; Pederson et al., 2016). Pederson et al. (2016) established that the development of the giant Lake Bonneville was unique to the latest Pleistocene, owing its existence to the diversion of the Bear River, which formerly flowed directly into the Snake and was thus exorheic, as a result of a volcanic eruption at ~55 ka. Another very large endorheic basin is occupied by Lake Lahontan in NW Nevada. Parts of this have a long history of intermittent lake development, dating back to the Pliocene (e.g., Reheis et al., 1993), but no exorheic connection has been identified. Former lacustrine basins such as these played an important role in the initial human settlement of North America in the Late Pleistocene (e.g., Adams et al., 2008). Setting aside river diversions such as affected Lake Bonneville, the principal influence in the development of these lake systems is thought to have been the reduced evaporation at times of colder climate (e.g., Ibarra et al., 2014).

Another case study from the United States is of particular interest: the record from the Rio Grande near Albuquerque, New Mexico (Connell et al., 2007; Westaway et al., 2009b). The river hereabouts flows within the Rio Grande Rift, in which sediment accumulation gave way around the start of the Middle Pleistocene to incision (effectively basin inversion), implying uplift of the graben interior and leading to the formation of a terrace system and a modern-day channel that is ~100–150 m below the top of the stacked sequence (e.g., Mack et al., 2002; Westaway et al., 2009b; Fig. 10). Westaway et al (2009b) cited this as an example of relative uplift within a rift valley, noting that the landscape beyond the bounding walls of the graben has been uplifting faster than its interior, in this case causing the inversion of this further example of a narrow, linear (structural) basin. They linked the inversion to the acceleration of surface processes and resultant erosional isostasy brought about by the longer and more extreme climatic cycles following the MPR.

South America is also host to numerous endorheic basins, particularly on the Andean Plateau, where they bear comparison with those of the Tibetan Plateau, being arid and fault-bounded as part of a high-altitude 'basin and range' setting (Ballato et al., 2019). These include the Salar de Atacama and Pampa del Tamarugal basins of Chile (Table 1), which are located, respectively, in the pre-Andean and central Andean depressions (Dorsaz et al., 2013). The Santa María, Humahuaca, and Quebrada del Toro basins of the Argentinian Andean plateau are considered to be parts of a former foreland basin that was subjected to Pliocene fragmentation and uplift (Kleinert and Strecker, 2001; Pingel et al., 2013; Ballato et al., 2019). The basins of this region are thought to have been subjected to repeated cycles of infill and excavation (inversion), with the unloading caused by the latter linked with renewed fault activity and promotion of the cyclicity (Hilley and Strecker, 2005; Ballato et al., 2019).

3.1 Basins that inverted in the mid-Cenozoic

It has been noted that many of the best examples of pre-Quaternary basin filling and inversion, the latter usually taking place in the late Neogene, have been from the Mediterranean regions. Further north in Europe many of the classic river terrace sequences are also incised into the infill of basins, although these sediments are heavily dissected, having inverted significantly earlier, in the mid-Cenozoic. Thus the terraces of the Middle and Lower Thames are emplaced above the Palaeogene infill strata of the synclinal London Basin, which are dominated by marine sediments but also include the terrestrial fluviatile Reading Formation. A quartz-rich facies of this formation, in Buckinghamshire, has been suggested as deposited by a precursor of the River Thames, carrying material from the north-west into the basin (White, 1906; Wooldridge and Gill, 1925; Bridgland, 1994). Another major British Quaternary river was the Solent, which formed the axial drainage of the Hampshire Basin, the fill of the latter being heavily incised by that river and its tributaries. Now drowned beneath the strait called 'The Solent' (part of the English Channel), its left-bank terrace staircase covers much of the southern–central coastal region of England (Everard, 1954; Allen and Gibbard, 1993; Briant et al., 2006; Westaway et al., 2006b), overlying the Palaeogene fill of the Hampshire Basin. The fluvial drainage systems represented by the terrestrial part of the Cenozoic Hampshire Basin were recently reviewed by Newell (2014), with their potential extension into Devon, as represented by the Haldon Formation, examined by Hamblin (2014).

The youngest deposit in either of these two Palaeogene basins is the Oligocene Bouldnor Formation of the Hampshire Basin. There is a substantial hiatus and a gap in evidence, therefore, between the basin-fill deposits and the Quaternary terrace sequences that are incised into them, with only a scattering of possible Miocene and Pliocene marine outliers on the highest ground, some perhaps lowered by solution of the Chalk, to hint at what is missing (e.g., Worssam, 1963; John and Fisher, 1984; Cameron et al., 1992; Daley and Balson, 1999). It is thus difficult to determine the timing of the inversion of these basins, or of the Paris Basin (e.g., Guillocheau et al., 2000) on the nearby continent, into which the terraces of the Seine system are incised (Antoine et al., 2007).

4. Further discussion

4.1 Small-scale basins or atypically thick terrace formations

In the upper parts of terrace staircases in certain fluvial systems are found atypically thick high-level deposits that are somewhat intermediate between normal terrace 'treads' and basin-fill sequences. Indeed, the formation of river terrace sequences in some fluvial systems can be seen to have been interrupted by intervals of subsidence during which basin-fill conditions prevailed. Such sequences are peculiar to certain crustal conditions, as was discussed in detail by Westaway and Bridgland, 2014); these are conditions in which there is a constricted mobile lower crustal level, either as a result of the crust being of near-cratonic, Early–Middle Proterozoic age (fully cratonic, Archaean crust is typically brittle throughout its depth, with no lower mobile layer, and so is ultra-stable, exhibiting minimal evidence for vertical movement; e.g., Westaway et al., 2003) or having experienced mafic underplating. The sequence in the valley of the River Don at Voronezh, Russia, ~800 km ENE of the Dniester (with its classic terrace sequence, as described above), is one example, showing evidence for two periods of sediment accumulation within a sequence that otherwise records progressive incision, presumably in response to uplift. These accumulation phases were from ~2 Ma to MIS 16, when the valley was glaciated and there was a major incision event, and again between that incision event and MIS 8; alternatively, this might represent a single phase of accumulation interrupted by glacially-induced incision (Matoshko et al., 2004; Westaway and Bridgland, 2014; Fig. 11). The vertical thickness of accumulated fluvial sediment here is in the region of 80 m, without the addition of the ~60 m of incision that took place following glaciation.

More straightforward, because the area is unglaciated, is the record from the Euphrates on the northern Arabian Platform. Here, in Syria and southern Turkey, the progressive uplift and incision recorded by the Euphrates terrace staircase was interrupted by periods of accumulation, and presumably subsidence, estimated to have occurred between ~5 and ~3 Ma, and again between about MIS 36 and MIS 22 (Demir et al., 2008; Bridgland and Westaway, 2014; Westaway and Bridgland, 2014; Bridgland et al., 2017). The stacked sequences have thicknesses of up to 40 m and ~50 m, respectively (Fig. 12) and should perhaps be regarded as small-scale basin-fills. It should be noted that the timing of the inversion of these basins approximates to the late Pliocene and to the MPR.

Accumulations of modest thickness are recognized in the uppermost part of certain terrace records that may be less readily attributed to crustal properties. The Vltava, around Prague, Czech Republic, has formed a staircase of ~12 terraces at the head of which is a relatively thick (~45 m) sequence of Pliocene sands (the Kobylisy Sands) associated, in terms of height range, with gravels of the Zdiby Group / Zdiby Terrace (Tyráček et al., 2004). Whether this uppermost (12th) terrace is part of an incision record or instead represents an early accumulation phase is unclear, as is the stratigraphical relation between the two sediment bodies; the Zdiby Group is described as occupying an infilled and abandoned early fluvial course (Tyráček et al., 2004).

A more striking example comes from the terrace record of the Yarlung Zangbo in the eastern Himalayas. Here, in an earlier FLAG special issue, Zhu et al. (2014) described what they termed the highest river terrace of this system, a body of sand and gravel 200–350 m thick, ~550 m above the valley-floor and ~150 m above the next highest terrace. These sediments are intermediate in age between the Oligocene–Miocene Dazhuka Formation, a body of sandstones, conglomerates and volcaniclastic rocks up to 1200 m thick that extends along the Yarlung Zangbo valley for some 1500 km and might also represent an ancestral version of the river, and the highest of the ten conventional terraces, dated at ~1.8 Ma by electron-spin resonance (Zhu et al., 2014; Fig. 12). Bridgland and Westaway (2014) argued from this evidence that, even in an area of spectacular Quaternary uplift such the Himalayan Massif, river-terrace formation was preceded by late Neogene 'basin' inversion, the thick high terrace being seen as a 'perched' basin-fill.

Of lesser thickness, but comparable in many ways, are localized occurrences of unusually thick gravel bodies within terrace sequences that have been interpreted as filling small tectonic basins, associated with active faulting. Generally falling outside the scope of this review, such small-scale features are probably common in tectonically active areas. For example, Viveen et al. (2012) have recorded small tectonic basin-fills within the terrace staircase of the River Miño, NW Iberia, with localized gravel thicknesses of ~30 m. Similar phenomena have been reported in the Huang Shui system downstream of Xining on the NE Tibetan Plateau (see above; Fig. 1) by Vandenberghe et al. (2011), who have referred to the atypically thick (again ~30 m) gravel bodies as 'accumulation terraces'. There is perhaps a continuum between the smaller recognized basins, the thick terrace deposits and valley-fill accumulations such as in the Euphrates and the high-level gravel in the Yarlung-Zangbo, and the smaller-scale examples in the Miño and Huang Shui, although further data compilation will be required to demonstrate whether that is so.

4.2 Coincidence of basins with cratons

The cold, fully brittle crust of the Archaean cratons has been linked to the lack of evidence for uplift in such provinces, as demonstrated by fluvial records with sediments of various ages disposed laterally rather than in terrace-staircase form (Westaway et al., 2003, 2009b; Bridgland and Westaway, 2008a, b, 2014). Although isostatically generated uplift in these areas cannot be sustained in the absence of a mobile lower crust (see above), isostatic depression from sediment loading can of course take place. Logically, a stable cratonic crustal province surrounded by more dynamic crust that is uplifting in response to erosional isostasy is likely to become a sediment-accumulation area. Once this occurs, isostatic subsidence will provide further accommodation space and the resultant depocentre will be sustained by positive feedback. This can perhaps explain the occurrence of several currently endorheic and/or inverted basins in Asia on cratonic crust (or Proterozoic crust where the mobile layer is relatively thin), such as the Tarim, Qaidam and Sichuan basins (cf. Yu et al., 2015; Demir et al., 2018; Table 1). Métivier et al. (1998) suggested that the area of the present endoreheic Qaidam Basin might, in pre-Quaternary times, have drained into the Yellow River (see Fig. 1). Middleton (1989) suggested that intracratonic basins have developed as a result of sediment loading, although his model invoked downward-trending mantle plumes in the initiation of sediment-accumulating depressions.

In marked contrast to the high-level cratonic basins of China, the lower-level cratonic regions of Africa generally lack surrounding dynamically uplifting crustal regions, so basin accumulation is less of a feature on that continent, notwithstanding the above-mentioned Okovango example.

4.3 Declining proportion of endorheic drainage during the Quaternary

It has been suggested already that the areas of the world occupied in the pre-Quaternary by subsiding fluvially-fed depocentres must have considerably exceeded the occurrence of such basins at the present day, although no quantification will be attempted here. The basins were (and are) of a variety of types, although many (perhaps most), past and present, have a structural origin. Some, usually the larger ones, are synclinal, while others are fault-bounded, these varying in size from substantial graben structures to small pull-apart basins such as in the Orontes examples and the even smaller Miño tectonic basins (see above).

A further reason for supposing the opening statement of this section to be true is that there are few currently active fluvial depocentres that can be shown to have formed since the Late Neogene, whereas numerous pre-existing basins have inverted during that interval. Tibetan Plateau basins that might have joined Pacific drainage systems and then become re-isolated, as has been suggested for the Qinghai Lake Basin, represent an exception, as do examples from the Basin and Range Province of the USA and Lake Van (see above); in both cases continuing tectonic activity (as well as volcanism in the USA and Turkey) can explain fluctuation between endorheic and exorheic drainage within particular basins. Cyclicity between subsidence and incision (inversion) is also identified for a small minority of the basins that appear in Table 1, notably three examples in the Andes.

4.4 Timing of and mechanism for inversion of the Yellow River basins

The timing of inversion is another variable, the understanding of which could be of value in improving comprehension of basin mechanisms in general. Some of the most recently inverted basins are those in the Upper reaches of the Yellow River, reinforcing the notion that these might be exceptional on account of their involvement in the structural evolution of the Tibetan Plateau. The degree of incision into basin-fill sediments is a readily apparent indication of time since inversion, albeit qualitative (see Fig. 14). The recent nature of inversion here should mean that evidence for causal mechanisms can be more readily obtained. As already noted, the upper basins of Zoige and GongHe remain partly endorheic, showing that basin-capture by drainage to the ocean need not produce rapid and complete inversion, at least in relatively arid regions like this. Controversy exists in the case of the Yellow River over whether basin capture has resulted from headward erosion or from basin overspill (Craddock et al., 2010; Perrineau et al., 2011; Hu et al., 2016, 2017; Kang et al., 2017). The increasing incision of basin infill in successive basins downstream within the system might be regarded as supporting the former view, which probably represents the majority opinion. Certainly the Zoige Basin shows very little incision of its sedimentary infill, which remains largely intact. The degree of incision into the GongHe and GuiDe basins (Fig. 14B) shows a marked contrast, with the former less incised that its downstream neighbour, seemingly in keeping with the fact that the GongHe basin was added to the Yellow River system quite recently, as implied by cosmogenic dating (Perrineau et al., 2011; see above), and the continued endorheic nature of a large part of that basin. Much hinges on those recent cosmogenic ages, however. The incision of the GongHe Basin infill amounts to ~700 m (Perrineau et al., 2011), compared with ~1000 m in the GuiDe basin (Fang et al., 2005; see above). Although this has been interpreted as a consequence of a later start to the incision (see above), it might instead be a reflection of the descent of the Yellow River by ~300 m across the reach occupied by these basins (which share an approximately level culminant surface), much of this being accommodated in the narrow Longyang Gorge between the two basins, where the river has a steep longitudinal gradient (cf. Perrineau et al., 2011). In detail, the history of vertical crustal motion in this region is complex, local effects of active faulting (warping and dislocating river terraces and the dissected lake sediments; e.g. Fang et al., 2005; Perrineau et al., 2011) being superimposed on the dramatic epeirogenic uplift of the Tibetan Plateau. Further work will be needed to elucidate these effects fully and to distinguish them from consequences of gorge entrenchment. Nonetheless, at Linxia, ~150 km downstream of GuiDe and only ~40 km inside the Tibetan Plateau, lower rates of regional uplift are evident: ~500 m of uplift since ~1.6 Ma being indicated by magnetostratigraphically dated river terraces (Li et al., 1997; Westaway, 2002). Rates of regional uplift evidently taper towards this margin of the plateau, as might be expected.

The African continent can perhaps provide an interesting comparison with the Tibetan Plateau and the basin-and-ridge character of the upper Yellow River. As noted by Goudie (2005), the interior of that continent is characterized by a basin and swell topography, which he considered to have been crucial in terms of African drainage development and he attributed to swells forming above mantle plumes. This is an interpretation linked to Africa's uniquely stationary situation in terms of plate-tectonic movements, which is thought to have been important in the development of numerous hot-spots above mantle plumes (e.g., King and Ritsema, 2000). Goudie envisaged dominantly endorheic drainage during drier intervals giving way to overspill in more humid episodes, the latter leading to the formation of such through-going drainage as the SSE course of the Niger towards its present coastal delta; indeed, he considered that in the more humid periods Lake Chad, currently endorheic, would have overflowed into the Niger system (after Talbot, 1980). The emphasis on overflow as a driver for changes from endorheic to exorheic drainage in Africa, as also envisaged for Iberia (see above; Cunha et al., 2020 [this issue]), invites comparison with the Yellow River, for which headward erosion has tended to be favoured as a mechanism (see above). The Tibetan Plateau shares the general aridity of Africa, but its greater elevation and tectonic (collision) setting are important differences. In addition, much of the interior of Africa is cratonic, including the bulk of the Niger catchment, the catchment of the River Congo and large parts of Vaal–Orange system (see Zhao et al., 2004); although much of China is also cratonic, the orogenic Tibetan Plateau crust is, in contrast, readily deformable on account of its thick lower crustal mobile layer and is thus considerably more dynamic.

5. Conclusions

Basin depocentres and their greater rarity during the Quaternary than earlier in the geological record represent one of the patterns determined from earlier work by the Fluvial Archives Group that demands further attention in the future. This review is intended to whet the appetite for such work rather than to draw definitive conclusions. It is clear that many formerly endorheic and/or subsiding basins have been inverted in the late Neogene and early Quaternary as the areas concerned have switched from subsidence to uplift, perhaps in response to progressive climatic cooling; some, such as the Rio Grande Rift, have inverted sufficiently recently to invoke the further cooling effects of the Mid-Pleistocene Revolution. There is much information from the Mediterranean region, where the crust is generally dynamic and where basin inversion has typically occurred within the past few million years, in contrast to regions further north in Europe, where inversion was somewhat earlier and the evidence from that time has been largely lost to erosion.

Areas of the world in which basin infill continues at the present day are typically, but not universally arid. They include currently and/or recently endorheic systems in Australia, Asia and Africa, as discussed. Where basins have been inverted in geologically recent times, having been connected to external drainage, it is often unclear whether this has come about because of capture by that drainage or by overflow of the basin at a time of greater humidity. The basins of the NE Tibetan Plateau, of particular relevance to this special issue, contrast in many ways with the low-lying basins elsewhere and their evolution may also differ, closely linked, as they are, to the active orogenic tectonism of the plateau. Further research aimed at determining which of the two mechanisms, overflow or capture, has been dominant in the evolution of these basins might well assist with the understanding of basins and their evolution elsewhere.

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Figure Captions:

Figure 1 – The Yellow River, showing its passage through the basin and ridge structure of the NE Tibetan Plateau, major faults, nearby basins and its course to the ocean.

Figure 2 – Location of the basins discussed in this paper. A - Crustal provinces of the Eurasian continent, modified from Demir et al. (2018), with basins highlighted and the locations of figures 1, 4B, 5A and 6 (inset) indicated. B – World map showing basins described in the paper that lie outside the area of A (the outline of which is shown.

Figure 3 – The Lake Eyre endorheic drainage basin, central Australia. The inset shows the extent of the basin within Australia.

Figure 4 – The Lower Rhine depocentre (reproduced, with modifications, from Bridgland and Westaway, 2014). A - Stratigraphy of stacked Lower Rhine deposits in relation to terraces further upstream; the numbers in circles indicate marine isotope stages. For location, see B. B - Map of the Lower Rhine channel system in the latest Pleistocene (after Busschers et al., 2007). C. - Cross section through the Late Pleistocene Rhine–Meuse sedimentary sequence beneath the central Netherlands (after Busschers et al., 2007).

Figure 5 – Subsiding and uplifting areas along the course of the Danube (from Gábris and Nádor, 2007). A - Map of the Pannonian Basin and its surroundings, showing the course of the Danube and the major Pleistocene alluvial on the Great Hungarian Plain. B - Longitudinal section crossing the Little Plain, Transdanubian Range and Great Plain, showing the contrast between uplifting areas, with terraces, and stacked sequences in the subsiding areas.

Figure 6 – Subsiding and uplifting areas along the course of the River Orontes in Syria and Turkey (after Bridgland et al., 2012). The inset shows the course of the river, broadly northwards along the Dead Sea Fault Zone. Different rates of uplift have been calculated for different uplifting reaches, as shown. The Amik and Ghab basins (the latter with its sattelite, the Acharneh Basin) continue to subside at the present day. The 'Pontian' marl was deposited in a Miiocene lacustrine basin.

Figure 7 – Sections through the River Gediz valley, near Kula, western Turkey (reproduced from Bridgland and Westaway, 2014). A. Transverse profile through the Early Pleistocene terrace staircase overlying Miocene İnay Group (predominantly fluviatile) sediments filling the Selendi Basin and preserved beneath the basalt capping of the Burgaz Plateau (after Maddy et al., 2008). B. Transverse profile across the Gediz valley near Eynehan and Karabeyli, showing basin-fill sediments of the Hacıbekir and İnay groups incised by the river and overlain by river-terrace remnants (after Seyitoğlu, 1997, and Westaway et al., 2004).

Figure 8 – Generalized cross section through the Middle–Lower Dniester valley, which is incised through Miocene fluvial basin-fill deposits; terrace sediments of various ages are shown, with MIS correlations shown at key levels (after Matoshko et al., 2004; reproduced from Bridgland and Westaway, 2014). For the location of this river, which flows into the northern Black Sea, see Fig. 2.

Figure 9 – The River Shoalhaven, SE Australia, an example of a small-scale Neogene basin or valley fill, subsequently incised during the Quaternary. Reproduced from Bridgland and Westaway (2014), after Nott (1992) and Nott et al. (2002). A - Cross section, compiled from borehole data, through

valley fill of the Mongarlowe Palaeochannel, near its confluence with the palaeo-Shoalhaven (for location, see D). B - Schematic cross section through valley fill (Nadgigomar Subgroup) of the palaeo-Shoalhaven in the region of Spa Creek (see D), showing dissection by the modern river. C - Schematic cross section through the Shoalhaven at Larbet, showing post-inversion terraces and TL dates (after Nott et al. 2002; inferred MIS correlations from Bridgland and Westaway, 2008a). D - Map of the Middle Shoalhaven, showing the footprints of Late Cenozoic (Oligocene) palaeovalleys, as well as the outcrops of older valley-fill sediments. The Oligocene basalt flows that dammed the system are also shown.

Figure 10 – Schematic transverse profile through the Rio Grande Rift and river valley around Albuquerque, New Mexico (35.1°N, 106.8°W). The zone represented is ~15 km wide. Stacked deposition of fluvial sands and gravels interspersed with deposits of lateral alluvial fans gave way in the latest Early Pleistocene to incision and terrace formation by the river. After Connell et al. (2007); reproduced, with modifications, from Westaway et al. (2009b). For the location of this river, see Fig. 2.

Figure 11 – Transverse profile through the deposits of the Upper Don near Voronezh (after Matoshko et al., 2004; reproduced from Bridgland and Westaway, 2014). For the location of this river, which flows into the northern Black Sea (Sea of Azov), see Fig. 2.

Figure 12 – Small-scale basin-fill episodes within the Late Cenozoic record of the River Euphrates, southern Turkey, as seen in an idealized transverse section through the sequence in the Birecik area; Holocene overbank deposits that cover the terraces assigned to MIS 6 and 2 are omitted. From Demir et al. (2018). For the location of this river, see Fig. 2.

Figure 13 – The River Yarlung Zangbo: high-level basin fills or thick terrace sequences (after Zhu et al., 2014). Cross sections through the Yarlung Zangbo terrace sequence at (A) Luobujue and (B) Xiari, both in Gyaca County, China. Ages from electron spin resonance (ESR) dating by Zhu et al. are indicated. For the location of this river, north of the Himalayas, see Fig. 2.

Figure 14 – GoogleEarth images of different basin reaches in the Yellow River valley across the NE Tibetan Plateau, showing differing degrees of incision into basin-fill sediments. Top left is a thumbnail from Fig. 1 for location. A - Zoige Basin. B - GongHe and GuiDe basins. C - Xunhua Basin.

Table 1. Selected basins worldwide, organized by type (starting with Yellow River basins)

Name, location	Size (km^2)	Sediment thickness (km)	Start of infill (Ma)	Date of inversion (Ma)	Endorheic or Exorheic (at present)	Principal literature
Basins along (or adjacent to) the course of the yellow River (see Fig. 1)						
Zaling–Eling	~1070	>0.23	>2.2	Extant	Exorheic	Zhu et al. (2009)
Zoige, Tibet	>6300	>0.3	>0.8	0.04-0.03	Exorheic	Pan (1994); Li (1991); Chen et al. (1999); Hu et al. (1999)
Xinghai–Tongde	~3200	>0.6	>3.0	<0.5	Exorheic	Craddock et al. (2010); Zheng et al. (1985)
GongHe	12500	2.3	~11	0.12-0.25	Exorheic	Li (1991); Pan, (1994), Perrineau et al. (2011); Zhang et al. (2012)
Qinghai Lake	~4300	~0.6	4.6	Extant	Endorheic	Li (1991); Pan, (1994), Perrineau et al. (2011); Fu et al. (2013); Zeng et al. (2014); Hu et al. (2017)
GuiDe	~1135	>1.5	~20	~0.18	Exorheic	Fang et al. (2005); Perrineau et al. (2011); Miao et al. (2012); Hu et al. (2017)
Xunhua	~3300	~2.6	>28	1.1	Exorheic	Xu (2015); Pan et al. (1996); Ji et al. (2010)
Linxia	3160	>1.1	29	1.7	Exorheic	Li et al. (1996, 1997); Fang et al. (2003); Li et al. (2014)
Xining	1725	~0.9	~54	4.8	Exorheic	Fang et al. (2019)
Lanzhou	~350	>1.5	~47	6.9	Exorheic	Wang et al. (2016); Li et al. (2017)
Zhongwei–Zhongning	~800	>1.05	>38	Extant (?)	Exorheic	Su et al. (2019); Zhang et al. (2004); Shen et al. (2001)
Yinchuan	7790	>7.5	Pre-	Extant (?)	Exorheic	Research Group AFSOM (1988); Pan

			Eocene			(1991); Wang et al. (2015); Su et al. (2019)
Hetao	28000	~14.8	Pre-Oligocene	Extant	Exorheic	Research Group AFSOM (1988)
Fenwei	>20000	~6.0	50	Extant	Exorheic	Research Group AFSOM (1988); Liu et al. (2013)
Other orogenic/tectonic (including foreland and extensional basins; grabens)						
Baltic	~300,000	~0.1	pre-Cambrian	Extant	Exorheic	Poprawa et al. (1999); Shogenova et al. (2009)
Neuwied, Germany	~600		~40	Palaeogene	Extant	Exorheic Bosinski (1995); Meyer and Stets (2002); Boenigk and Frechen, (2006); Cordier et al. (2009)
Alpine (N)	~70,000	up to 12	34	~5	Exorheic	Beaumont (1981); Pfiffner (1986)
Po (Alpine S)	46,000	>4	~5	Extant	Exorheic	Ori (1993)
Ebro, Spain	~28,000	~5	34	By ~8	Exorheic	Garcia-Castellanos et al. (2003); Nichols (2007); Stange et al. (2014)
Madrid, Spain	>20,000	~3.8	Final Mesozoic	~3.65	Exorheic	Pérez-González (1994); Cunha (2019); Cunha et al. (2008, 2016, 2020); Pais et al. (2012); de Vicente and Muñoz-Martín (2013)
Duero, Spain	~50,000	~3.5	~48	0.5–0.6	Exorheic	Herrero et al. (2010); Cunha et al. (2020); Pais et al. (2012)
Selendi, Turkey	~2300	≤0.4	~17	By 2.6	Exorheic	Seyitoğlu (1997); Westaway et al. (2004); Purvis and Robertson (2005); Ersoy et al. (2010)
Adiyaman, Turkey	~4000	≤0.3	~6	~3.6	Exorheic	Demir et al. (2007, 2012); Seyrek et al. (2008); Westaway et al.

						(2008)
Malatya, Turkey	~800	≤2.0	Late Miocene	~3.6 (?)	Exorheic	Türkmen et al. (2007); Seyrek et al. (2008); Westaway et al. (2008)
Diyarbakır, Turkey	~3600	~0.6	Late Miocene	~3.6 (?)	Exorheic	Westaway et al. (2009a); Karadoğan et al. (2014)
Lake Van Turkey	>3500	>0.2	before 0.6	Extant	Endorheic	Degens et al. (1984); Litt and Anselmetti (2014); Stockhecke et al. (2014a, b)
Lake Tuz, Turkey	>20,000	up to 13	~68	Extant	Endorheic	Çemen et al. (1999); Aydemir and Ates (2006)
Alamut–Taleghan (Iran)	~1000	7.3	17.5	by 3	Exorheic	Guest et al. (2007); Ballato et al. (2008, 2015)
Mianeh, Iran	~7000	>0.1	>7	<4	Exorheic	Heidarzadeh et al. (2017)
Himalayan Foreland	~800,000	>4	~13	Extant	Exorheic	Tandon (1976); Molnar (1984); Sinha et al. (2007)
Hexi Corridor, Tibet (Fig.1)	160,000	>3	pre-Cenozoic	Extant	Exorheic	Métivier et al. (1998); Hetzel et al. (20014)
Rio Grande, New Mexico	~15,000	>7.5	~35	~0.8	Exorheic	Mack et al. (2002); Connell et al. (2007); Repasch et al. (2017)
Salar de Atacama, Chile	17,318	>1	~30	Extant	Endorheic	Naranjo et al. (1994); Hartley and Chong (2002); Dorsaz et al. (2013)
Pampa Tamarugal, Chile	15,397	>1	26	Extant	Endorheic	Dorsaz et al. (2013), Garcia et al. (2011); Labbé et al. (2019)
Humahuaca, Argentina	~150	~0.35	pre-Cenozoic	Cyclic	Exorheic	Pingel et al. (2013); Ballato et al. (2019)
Quebrada del Toro, Argentina	~120	>0.7	pre-Cenozoic	Cyclic	Exorheic	Marrett et al. (1994); Hilley and Strecker (2005); Tofelde et al. (2017); Ballato et al.

						(2019)
El Cajón–Campo Arenal, Argentina	~1200	1.2	11	from ~6	Exorheic	Mortimer et al. (2007)
Santa María, Argentina	~70	~2.0	~13	Cyclic	Exorheic	Strecker et al. (1989); Bossi et al. (2001); Kleinert and Strecker (2001); Ballato et al. (2019)
N. Patagonia–Colorado	~180,000	~7	pre-Cenozoic	Extant	Exorheic	Zambrano and Urien (1970); Folguera et al. (2015)
Depocentre (sag basin)						
Lower Rhine, NW Europe	~20,000	~1.0	>2.5	Extant	Exorheic	Brunnacker et al. (1982; Ruegg (1994); Busschers et al. (2007)
Paris	~140,000	~4	250	By ~30	Exorheic	Brunet and Le Pichon (1982); Guillocheau et al. (2000)
Pannonian, Hungary	>100,000	>0.7	~7	Extant	Exorheic	Ruszkiczay-Rüdiger et al. (2005); Gábris and Nádor (2007)
Black Sea periphery	>500,000	>0.15	pre-Cenozoic	~3.5	Exorheic	Matoshko et al. (2009)
Murray–Darling	~300,000	up to 0.6	Cenozoic	Extant	Exorheic	Bowler (1978); Stephenson (1986); Stephenson and Brown (1989); Glen (1992)
Cenozoic Syncline						
London	~8000	~0.3	58	By ~25	Exorheic	Ellison et al. (1994); Daley and Balson (1999)
Hampshire	~4000	~0.5	58	By ~25	Exorheic	Edwards and Freshney (1986); Newell (2014)
Lisbon (L. Tagus)	~9000	~2	~45	By 1.8	Exorheic	de Vicente et al. (2011, 2018); Pais et al. (2012); Carvalho et al. (2017); Cunha (2019)

Pull-apart						
Ghab, Syria	~900	>0.8	~60	Extant	Exorheic	Devyatkin et al. (1997); Brew et al. (2001); Bridgland et al. (2012)
Amik, Hatay, Turkey	~900	~1	>3.7	Extant	Exorheic	Bridgland et al. (2012); Seyrek et al. (2014)
Aridity-driven (generating endorheic sag basin)						
Chad	2,500,000	~3.8	pre-Cenozoic	Cyclic	Endorheic	Servant and Servant-Vildary (1980); Griffin (2002)
Okavango, Botswana	~725.000	0.3	>60?	Extant	Endorheic	Gumbricht et al. (2001)
Lake Eyre	1,140,000	>0.4	~60	Extant	Endorheic	Rust and Nanson (1986); Habeck-Fardy and Nanson (2014)
Death Valley, USA	>500		~3	~1	Extant	Endorheic
*Great Basin**	*~550,000*	0.3	<2?	Extant	Endorheic	Morrison (1965); Eardley et al. (1973); McCoy (1987); Oviatt et al. (1999)
`* (including Lakes Bonneville & Lahontan)						
Intracratonic/oceanic (generating endorheic sag basin)						
Caspian Sea	371,000	>20	pre-Cenozoic	Extant	Endorheic	Brunet et al. (2003); Green et al. (2009)
Tarim, China	560,000	3	24	Extant	Endorheic	Sobel et al. (2003); Yu et al. (2015); Demir et al. (2018)
Qaidam, China	103,000	>11	pre-Cenozoic	Extant	Endorheic	Métivier et al. (1998); Yu et al. (2015); Demir et al. (2018)
Sichuan, China	180,000	2	M.Jurassic	Extant	Exorheic	Yu et al. (2015); Liu et al. (2016); Demir et al. (2018)

Highlights

- Late Cenozoic sedimentary basins are compared with reference to type and evolution
- Patterns have been determined in relation to crustal type and climate
- Internally (endorheic) draining basins show a post-Neogene decline in number
- Change to external drainage may be linked to increased climatic severity
- Yellow River basins may be atypical because of association with India–Asia collision