Morphodynamics of rivers strongly affected by monsoon precipitation:

Review of depositional style and forcing factors

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Abstract

Rivers that receive significant amounts of their surface water supply from monsoon precipitation characteristically experience seasonal floods, and display seasonally highly variable discharge, controlled by the monsoon trough passage and its related cyclones. The intense summer rainfall causes high-magnitude floods, whereas rivers only transmit a low base flow during the dry winters. For many rivers in the sub-humid to arid subtropics, bordering the monsoon domain, the monsoon rain is also the main source of surface water recharge. However, such rivers may receive monsoon rain and transmit discharge only during abnormal or strengthened monsoon seasons. This annual discharge variability or range, as compared to the mean annual discharge, distinguishes the monsoonal and subtropical rivers from the rivers in equatorial tropics and temperate perennial precipitation zones, where the annual range is relatively small compared to the annual mean discharge.

This review explores the effects of this seasonal and yearly variable rainfall, and the resultant highly peaked discharge pattern on river morphodynamics, and presents a comparison of modern and ancient monsoonal and subtropical river deposits. The field datasets and literature analyses discussed herein provide recognition criteria for monsoon-controlled river deposits, by documenting the diversity of the sedimentary facies, macroforms (bar forms), and architectural

elements common in ancient and modern monsoon-controlled rivers. The review demonstrates that seasonal and inter-annual precipitation range is a key control on river morphodynamics, and resultant sedimentary facies characteristics, rather than the specific climate zone or average annual precipitation.

Keywords: Monsoon, subtropical, river morphodynamics, discharge range, high-magnitude river floods; monsoonal and subtropical river facies models

1. Introduction

The global monsoon defines essential features of the Earth's climate, and monsoon rainfall directly affects more than 70% of the world's population. The monsoon rainfall is the key parameter in global hydrological cycle and is most influential on human sustenance and society (Wang and Ding, 2008). The Asian monsoon alone directly affects the livelihood of 60% of the world population. The monsoon regions reside on either side of the equatorial perennial rainfall region and occupy about 19.4% of the total area of Earth's surface (Fig. 1a, b). However, the monsoon rainfall accounts for 30.8% of the total precipitation on the globe, and thus the monsoon regions are the most concentrated rainfall-regions in the world (Wang and Ding, 2008). Of all the atmospheric circulation systems, monsoons exhibit the most significant seasonal variations (Zhisheng et al., 2015). Monsoon means "season" (Zhisheng et al., 2015), and is defined by the high seasonal precipitation variability and an annual reversal of the low-level winds, as a function of the seasonal migration of the Inter-Tropical Convergence Zone (ITCZ; e.g., Wang and Ding 2008) (Fig. 1a, b). More than 80-90% of annual precipitation in this region falls during the summer monsoon. This annual precipitation variability or range, as compared to the mean annual precipitation, distinguishes the monsoon domain (encircled by black line in Fig.

1a) and its bordering sub-humid to arid subtropics (green shades outside the black line in Fig. 1a), from the perennial rainfall regime in equatorial tropics and temperate zones, where the annual precipitation range is relatively small compared to its annual mean (Wang and Ding, 2008) (Fig. 1a).

Monsoon rain is the main surface water supply to many of the world's rivers whose catchments fall within the tropical or subtropical monsoon domain (green shades in Fig. 1a). Such rivers characteristically experience seasonally highly variable discharge, controlled by the intense monsoon precipitation during the passage of the monsoon and its related cyclones (see also Latrubesse et al., 2005; Syvitski et al., 2014) (Fig. 1c-i). The intense summer rainfall causes high-magnitude floods, whereas these rivers only transmit a relatively low base flow during the dry winters (Fig. 1f). Some monsoonal rivers, such as the rivers in the Gangetic plains in India, can have discharges 40–50 times greater during the summer monsoon compared to the dry winter (Sinha and Friend, 1994; Sinha and Jain, 1998; Latrubesse et al., 2005). For example, the Ganges River delivers 80% of water discharge and 95% of its sediment load during the summer monsoon season (Goodbred, 2003), as only the monsoon flood discharge is geomorphically effective and able to transport sediment. This large discharge range, as compared to its annual mean is an inherent property of monsoonal rivers, as it characterizes monsoonal and sub-tropical rivers with a variety of drainage basin sizes, including those $> 10,000 \text{ km}^2$, and in a variety of geomorphologic-geologic settings (Latrubesse et al., 2005). Thus, this paper invokes that this large seasonal discharge variability or range, as compared to its annual mean, sets these rivers (Fig. 1f) apart from the perennial rainfall regime experienced by rivers in equatorial tropics and temperate zones (Fig. 1g).

Monsoon anomalies may cause droughts, catastrophic floods, and other extreme weather or climate events (Seneviratne et al., 2012; IPCC, 2014; Zhisheng et al., 2015). Monsoonal rivers are especially prone to catastrophic terrestrial flooding events (Sinha, 2009; Brakenridge, 2012; Syvitski and Brakenridge, 2013). Thus, the monsoon inter-annual variability and predictability, the consequent flood hazards, as well as the impact of anthropogenic climate change on monsoonal systems worldwide are topics of extensive research (e.g., www.wcrp-climate.org; IPCC, 2014). Yet, key questions remain unanswered about this most significant component of the global hydrological cycle, such as extracting long-term trends from high-frequency inter-annual variability, or climate forcing from human activity-induced changes (e.g., Seneviratne et al., 2012; IPCC, 2014). For instance, increased frequency and magnitude of terrestrial flooding have been suggested to result from the intensification of extreme monsoon precipitation due to anthropogenic global warming (e.g., Seneviratne et al., 2012; IPCC, 2014). Yet, similar effects can also occur due to human activity within the catchment, such as pasture and intensive agriculture that increases sediment supply and channel bed aggradation rates, which can cause abrupt channel course changes (avulsions) and large catastrophic floods (Assine, 2005). Similarly, flood mitigation by channel restriction (dam, embankment and barrage/levee construction) that reduces water and sediment conveyance capacity may also induce high magnitude floods, like the Kosi flood in 2008 or Indus flood in 2010 (Sinha, 2009; Brakenridge, 2012; Syvitski and Brakenridge, 2013). The latter examples demonstrate that the key morphodynamics of monsoon-controlled rivers are poorly understood, and there is a need for a process-based approach, as understanding the coupling between river form and processes is essential in finding long-term solutions to flood control and disaster risk management (Sinha, 2009; Brakenridge, 2012; Syvitski and Brakenridge, 2013).

This paper presents a different approach from previous syntheses of monsoonal rivers, where they have been treated separately from subtropical river systems and considered as rivers within tropical geographic zone that experience annual rainfall of over 700 mm/year (Latrubesse et al., 2005), or as rivers within the tropical climate zone (Syvitski et al., 2014). The tropical geographical zone is defined by the Tropics of Cancer (23°27'N) and Capricorn (23°27'S). The tropical climate zone is defined by minimum temperatures and average annual and monthly precipitation amounts (Syvitski et al., 2014). Instead, this paper explores the effects of the rainfall regime, and especially considers the large annual range as compared to its annual mean, and the resultant precipitation peakedness, characteristic for monsoon precipitation (Wang and Ding, 2008) (Fig. 1a); and how that translates into large annual discharge and sediment transport range (e.g., Henck et al., 2010) (Fig. 1c, f-i). Therefore, this analysis contains both tropical monsoonal and subtropical rivers that receive most of their surface water from monsoon precipitation, but excludes tropical or subtropical rivers that have a low annual discharge range, as compared to their annual mean, because they partially drain the perennial equatorial tropics or temperate zones (e.g., Parana, Japura, Orinoco in Fig. 1g).

Ancient sedimentary records of monsoon-controlled rivers provide a comparison for long-term climate forcing, as human effects are negligible, and long-term changes rather than high-frequency variability are preserved in the sedimentary record. Yet, examples of ancient river systems, identified as monsoonal, are few, and no specific facies models or recognition criteria have been proposed, other than the potential of monsoonal rivers for developing fluvial megafans (e.g., Singh et al., 1993; Shukla et al., 2001; Leier et al., 2005). Facies models for subtropical rivers expose a wide variety of sediment types and thus seem contradictory, in that some suggest that dominant Froude supercritical flow structures are the key characteristic (e.g., Tunbridge,

1981; Sneh, 1983; Stear, 1985; Deluca and Eriksson, 1989; Olsen, 1989; Bromley, 1991; North and Taylor, 1996; Billi, 2007; Fielding et al., 2009; Allen et al., 2013), whereas other challenge that (Abdullatif, 1989), and yet others propose that in-channel vegetation and the vegetation-induced heterogeneity as a key recognition criterion (Fielding et al., 2009; Allen et al., 2013). Furthermore, specific climate controls, such as the effect of highly seasonal monsoon precipitation are rarely analyzed in terms of sedimentary systems, as climate forcing is commonly implied as a simple increase or decrease in mean annual precipitation (aridity-humidity), and the effects of the precipitation peakedness are commonly excluded or difficult to identify. Yet, a significant change may occur in a river system dynamics with no change in annual mean precipitation, if the annual precipitation falls during a short wet season rather than over the course of the year.

This review attempts to reduce these knowledge gaps and aims to: (1) explore the effects of the seasonal and yearly variable rainfall (Fig. 1a), and the resultant highly peaked discharge pattern (Fig. 1c, f-i) on river morphodynamics; (2) document the diversity of the sedimentary facies, macroforms (bar forms), and architectural elements in rivers strongly affected by monsoon precipitation; and (3) demonstrate that the large seasonal and inter-annual precipitation range, as compared to its mean, is a key control on river morphodynamics and the resultant sedimentary facies characteristics, rather than the specific climate zone or the average annual precipitation (Fig. 1d).

2. Rivers in the monsoonal and subtropical domains

Monsoon precipitation pattern in tropics and subtropics (green area in Fig. 1a) is distinguished from the perennial precipitation regime in temperate and rainforest zones by the high

precipitation range, calculated as summer precipitation minus winter precipitation in Figure 1a (MJJAS minus NDJFM precipitation in the northern hemisphere and NDJFM minus MJJAS in the southern hemisphere; Wang and Ding, 2008). The range, as compared to annual mean precipitation, gives the monsoon precipitation index (MPI) that is > 0.5 for the monsoon domain (Wang and Ding, 2008) (Fig. 1a). The link between this monsoon precipitation pattern and the highly variable river discharge is demonstrated by the robust correlation between the precipitation seasonality index (calculated as average precipitation of the wettest minus driest month divided by the annual mean) and the discharge seasonality index (calculated as the average discharge of the wettest minus driest month divided by the annual mean) (Fig. 1c). In contrast there is poor correlation between the average annual precipitation and average annual discharge (Fig. 1d), or even between the average precipitation of the wettest month and its discharge (Fig. 1e). Figure 1f illustrates this large discharge range, as compared to its annual mean in monsoon domain and subtropical rivers (discharge seasonality index 2.00-5.08), as compared to perennial precipitation regime rivers in temperate and rainforest zones in Figure 1g (discharge seasonality index 0.50-1.68). The river systems in Figure 1c-i include a variety of drainage basin sizes (including those $> 10,000 \text{ km}^2$) and a variety of geomorphologic-geologic settings. In Figure 1c-e the data for each river are averaged over ca 4-80 years, and in Figure 1f-i over ca 10-60 years. The monsoonal and subtropical rivers are also differentiated by the relatively larger positive deviation from mean values (1.25-4.08 as compared to 0.30-0.92 in Fig. 1h), indicating considerably larger flood regime or magnitude as compared to the annual mean (see also Latrubesse et al., 2005). Thus, although the perennial precipitation zone rivers also experience floods, including floods of catastrophic proportions (e.g. on the Mississippi), the difference between their flood discharge and annual mean discharge is relatively small,

161 compared to the monsoonal zone and subtropical rivers. This difference is also demonstrated by
162 the average flood discharge (calculated as the normalized average discharge of the four
163 consecutive months with highest discharge) as compared to the annual mean (Fig. 1i). The latter
164 has a very close value to the normalized standard deviation that is > 0.8 for the monsoonal and
165 subtropical rivers, and < 0.5 for the rivers of the perennial precipitation zone (Fig. 1i).

166 The two rivers closest to the discharge seasonality index 2 are the Brahmaputra (index 2.00;
167 Fig. 1f) and Orinoco (index 1.68 in Fig. 1g). Both these rivers receive part of their surface water
168 supply from monsoon rainfall, but the Brahmaputra has a large part of the drainage basin outside
169 of the monsoonal zone, and the Orinoco partially drains the rainforest zone. Also the Ganges,
170 Indus and Kosi (Fig. 1f) receive part of their surface water from snow melt at high elevations,
171 but still transmit the vast majority of their water discharge and sediment flux during the
172 monsoon-induced floods (seasonality index 3.80, 4.43 and 2.63, respectively). Thus, rivers
173 dominated by monsoon precipitation as their surface water supply are distinct from the rivers of
174 temperate or rainforest perennial precipitation zones; however there is a continuum of river
175 behavior, as many rivers have mixed surface water supply.

176 Compared to the monsoon domain (green within the black line in Fig. 1a) most
177 subtropical areas are typically drier, in many areas with evapotranspiration exceeding
178 precipitation. However, some sub-humid subtropical areas may have a precipitation range similar
179 to the monsoon domain, where it is greater than 300 mm (Wang and Ding, 2008). A critical
180 difference of the subtropical zone as compared to the monsoon domain, is a considerably larger
181 inter-annual variability (Wang and Ding, 2008). In some subtropics the intense monsoon-driven
182 rain may occur seasonally (mostly in sub-humid subtropics), whereas others have a recurrence
183 frequency of multiple years or even decades (mostly in semi-arid and arid subtropics). This

extreme inter-annual precipitation variability is especially distinct in areas that receive monsoon rain only during extreme monsoon trough migration or abnormal or strengthened monsoon seasons and associated cyclonic flow. For example, the normalized inter-annual variability for the monsoonal Ganges is 0.54 (calculated as the range of annual discharge values divided by annual mean discharge over 7 years), whereas for the subtropical Burdekin river it is 3.56 (Fig. 1j). Most monsoon domain rivers have a perennial base discharge, whereas many rivers in the subtropical zone transmit discharge only in response to monsoon rainfall events, and are otherwise dry, unless part of their drainage is in a different climate zone (e.g., the Nile). The monsoon rains in the subtropics may last for a whole monsoon season, similar to the monsoon domain itself, or the whole annual precipitation may fall in a few episodic downpours. As return frequency of such intense rainfall events may be from years to decades, some such rivers are ephemeral and may remain dry over these timescales. Such rivers characteristically experience high magnitude flash floods, especially in areas with mountainous hinterland and sparse vegetation.

In summary, both the monsoonal domain and the subtropical rivers (including dryland rivers) are strongly affected by the seasonal monsoon precipitation, and characteristically experience high-magnitude seasonal floods that last from just a few hours to months, with a yearly (seasonal) to up to decadal (ephemeral) recurrence frequency. The high-magnitude floods are caused by intense monsoonal rainfall events and the flood discharge is significantly higher than annual mean discharge (Fig. 1f-i). Monsoonal rivers may transmit a low perennial base discharge, or only transmit discharge throughout the monsoon season, or the river beds may remain dry for most of the year, or even for decades, except during extreme precipitation events.

3. Datasets

Datasets presented in Table 1 include 17 modern or recent, and 35 ancient examples. Since deposit characteristics are the focus of this review, only papers containing facies descriptions of modern or ancient river systems, identified as highly seasonal monsoonal or subtropical, are included in this analysis (Table 1). In addition detailed field data from the early Eocene Wasatch and Green River Formations in the Uinta Basin, the Cretaceous Williams Fork Formation in the Piceance Basin, the Triassic Chinle Formation in the Dinosaur National Monument, and the Pennsylvanian-Permian Lyons Formation in the Front Range of Colorado are incorporated (Plink-Bjorklund and Birgenheier, 2013; Plink-Bjorklund et al., 2014a, b; Plink-Bjorklund, 2015) (Table 1).

The ancient examples are overrepresented in the literature, because the objective of this review was to establish facies characteristics and recognition criteria for monsoonal and subtropical river deposits, and few examples of modern systems include detailed sedimentary descriptions. Table 1 shows a qualitative assessment of the characteristics of these rivers, as few papers present quantitative data. Quantitative data was used where available, such as presented in Figure 6.

The strength of the modern river systems lies in that we know and can measure the climate conditions and hydrological responses. However, the modern systems tend to have more limited data on the nature of sediments, as outcrop or core are very limited. In contrast, the ancient examples provide large and detailed datasets on sedimentary structures, bar forms and architecture. In order to reduce interpretive uncertainties, the majority of the incorporated ancient datasets (32 out of 35) present independent data on paleoclimate conditions, such as plant or animal fossil records, palesol and trace fossil records, or the presence of climatically distinct

facies (e.g., eolian, playa etc.). There are remarkable similarities between sedimentary features of the studied modern and ancient systems (Table 1). Thus, this paper chooses the comparative approach, and integrates the ancient and modern datasets. Every characteristic discussed below has been recognized in both modern and ancient datasets.

4. Characteristic facies

Below is a synthesis of facies characteristics that are common in modern and ancient monsoonal and subtropical rivers that have highly seasonal discharge. In contrast, these facies are uncommon in temperate perennial precipitation zone river deposits that are overrepresented in existing fluvial facies models (Table 2; see also e.g., Collinson, 1996; Miall, 1996; Bridge, 2006). Some of these characteristics, like the abundance of Froude transcritical and supercritical flow sedimentary structures (4.1) occur essentially in all the presented examples (see Table 1). Other features, like abundant high deposition rate sedimentary structures (4.2), or in-channel mud layers (4.3) are very common. Yet other features, like thick mud-clast conglomerates (4.4), abundant soft-sediment deformation (4.5), in-channel vegetation and vegetation-induced sedimentary structures (4.6) or in-channel bioturbation (4.7) only occur in less than half of the datasets (Table 1).

4.1. Froude transcritical and supercritical flow sedimentary structures

Sedimentary facies most commonly linked to seasonally variable-discharge river deposits in modern and ancient datasets are those associated with Froude transcritical and supercritical flow or upper flow regime (UFR) sedimentary structures, such as parallel or planar laminations, convex-up low-angle bedforms, scour and fill structures and humpback or sigmoidal cross strata

(e.g., McKee et al., 1967; Williams, 1971; Frostick and Reid, 1977; Foley, 1978; Tunbridge, 1981; Stear et al., 1985; Abdullatif, 1989; Bromley, 1991; North and Taylor, 1996; Alexander et al., 2001; Fielding, 2006; Billi, 2007; Fielding et al., 2009; Allen et al., 2013) (Figs. 2-5). A high abundance of these facies is distinctly different from river deposits from perennial precipitation zones that are dominated by Froude subcritical flow sedimentary structures, such as cross stratification (see e.g., Collinson, 1996; Miall, 1996; Bridge, 2006, for river facies models) (Fig. 6; Table 2).

This abundance of UFR structures is attributed to high flow velocities, and the characteristically rapid rise, sharp peak and rapid decline of the flood hydrograph, common in rivers that experience seasonal floods. High flood discharges allow for high flow velocities of around $5\text{--}6\text{ m s}^{-1}$ (Stear, 1985; Abdullatif, 1989). Experimental work shows that planar laminations (Fig. 2) can form due to very low-angle bedform migration (e.g., McBride et al., 1975; Bridge and Best, 1988; Cheel, 1990; Best and Bridge, 1992) or by the passage of sweeps and bursts (e.g., Bridge, 1978; Cheel, 1990), in most cases under Froude transcritical or supercritical flow conditions (Paola et al., 1989; Cheel, 1990; Bridge and Best, 1997). The convex-up low-angle laminations (Figs. 3-5), also referred to as antidunes (e.g., Middleton, 1965; Simons et al., 1965; Allen, 1966, 1984b; Middleton and Southard, 1984; Cheel, 1990; Alexander et al., 2001), are similar to in-phase waves produced under experimental conditions (e.g., McBride et al., 1975; Cheel, 1990). They form at flow velocities and Froude numbers higher than that required for plane bed deposition, and migrate downstream, related to stationary waves, or migrate upstream, depending on flow velocities (Cheel, 1990; Alexander et al., 2001). Scour and fill structures (Figs. 3-5) indicate initial erosion followed by sediment aggradation and infilling of the scour features, and have been related to antidune migration and chute and pool

formation (Kennedy, 1963; Leopold et al., 1966; Foley, 1978; Alexander et al., 2001). The humback or sigmoidal cross strata (Fig. 4) have been linked to the transition from dune to plane bed deposition, related to Froude numbers just below what is needed for UFR plane beds (Bridge, 1981; Bridge and Best, 1988).

Some seasonal river deposits contain a great variety of UFR structures (e.g., Abdullatif, 1989; North and Taylor, 1993; Fielding, 2006; Fielding et al., 2009; Allen et al., 2011, 2013, 2014; Plink-Björklund and Birgenheier, 2013) (Fig. 6). In many cases the different UFR structures occur together and transition vertically or laterally into other UFR structures (Figs. 2e, 3c-e, i, 4a-d). However, some river deposits display an abundance of planar laminations alone (e.g., McKee et al., 1967; Scott et al., 1969; Williams, 1971; Frostick and Reid, 1977; Tunbridge, 1981; Parkash et al., 1983; Stear, 1985; Deluca and Eriksson, 1989; Bromley, 1991; Billi, 2007), low-angle convex-up bedforms (e.g., Leopold et al., 1966; Foley, 1978; Parkash et al., 1983), or scour and fill structures (e.g., Hand et al., 1969; Singh and Bhardwaj, 1991) (Fig. 6). The specific proportion of UFR structures as compared to Froude subcritical flow sedimentary structures varies significantly between different seasonal river systems, as well as within the same system (Fig. 6). In some extreme cases up to 80-100% of sedimentary structures are UFR structures (e.g., Tunbridge, 1981; Billi, 2007; Plink-Björklund and Birgenheier, 2013) (Fig. 6). In Table 1, except for one example (Donselaar et al., 2013), document an abundance, and 43% display a dominance of Froude transcritical and supercritical flow sedimentary structures. The actual content of UFR sedimentary structures can be even higher than shown in Figure 6 and Table 1, because the low-angle convex-up laminae, the scour and fill laminae and the sigmoidal cross strata are commonly acknowledged as cross strata rather than UFR structures (see also Alexander et al., 2001).

299 The content of UFR structures falls into 3 groups in Figure 6, where examples 1-5 have 70-
300 100% UFR structures, 6-8 around 50%, and example 9, that is an example of a river from
301 perennial precipitation regime, less than 15%. There are two modern examples within the first
302 group, the Kobo Basin, Ethiopia (Billi, 2007) and Karoo, South Africa (Stear, 1985). Kobo basin
303 has annual mean rainfall of ca 750 mm (Billi, 2007), whereas Karoo only around 200-300 mm
304 (Stear, 1985). Common to both is that this annual rainfall is received by a few downpours during
305 the monsoon season (Fig. 1a). Both rivers only transmit discharge during these events, and are
306 otherwise dry. The flood event in the Karoo, described in Stear (1985) was caused by an extreme
307 rainfall event, where 300 mm of rain fell in a few hours. Such extreme rain events are likely to
308 cause flash flooding, especially in areas with sparse vegetation. Flash floods have especially
309 rapid rise, sharp peak and rapid decline of the flood hydrograph, and are thus likely to deposit
310 UFR bedforms. In contrast, in the modern example of the second group, the River Gash in Sudan
311 experiences discharge for most of the monsoon season with peak floods that last for up to a week
312 (Abdullatif, 1989). Thus the discharge is of longer duration with multiple high-magnitude peaks,
313 but also lower-magnitude flood discharges. Such longer duration and variable flood discharges
314 are likely to produce UFR bedforms during peak conditions, but also Froude subcritical
315 bedforms during lower discharges. The longer discharge duration also allows for more sustained
316 conditions, needed for dune bedform formation and migration. This is why the proportion of
317 cross stratification tends to be higher. Yet, this group is still distinctly different from the river
318 example of the perennial precipitation zone, where cross strata form over 80% of the
319 sedimentary structures identified (Fig. 6). It is cross stratification that is the most common
320 sedimentary structure of the sandy river channel fills in perennial precipitation zones, where

discharge conditions are of lower velocity and more sustained (e.g., Collinson, 1996; Miall, 1996; Bridge, 2006; see also discussion in Fielding et al., 2009) (Table 2).

4.2. *High deposition rate sedimentary structures*

Some of the UFR sedimentary structures described above also indicate high deposition rates. The convex-up low-angle bedforms are aggradational bedforms with sub-vertical angles of climb (Fig. 3c, d), similar to climbing ripples (Fig. 7a-c). They form when sediment fallout rates exceed bedform migration rates (e.g., Allen, 1984). The humpback cross strata occur as cross strata with aggradational planar laminations in topsets and bottomsets (Fig. 3d-f), and thus indicate that these are also aggradational bedforms, and associated with high deposition rates (Bridge, 1981; Bridge and Best, 1988).

In addition to the aggradational UFR sedimentary structures, climbing dune cross stratification (Fig. 7d), and especially climbing ripple lamination (Fig. 7a-c) are common in seasonal river deposits. The latter are linked to the rapid decline of flood waters from upper to lower flow regime conditions (e.g., McKee et al., 1967; Picard and High, 1973; Croke et al., 1998). There are some documented examples of channel fills that are completely or dominantly filled with climbing ripple cross laminated sandstones, separated by erosion surfaces (Plink-Björklund and Birgenheier, 2013) (Fig. 8a), mud layers (Donselaar et al., 2013) (Fig. 8b), or both (Olsen, 1989) (Fig. 8c).

Another sedimentary structure, common in some seasonal river deposits, is gradational planar lamination (e.g., North and Taylor, 1996; Billi, 2007; Banham and Mountney, 2014; Plink-Björklund and Birgenheier, 2013) (Figs. 6, 7e-f). Such planar laminae are characterized by multi-mm or cm thick laminations that display gradational or graded boundaries. Essentially the

laminations consist of reverse and normal graded intervals (Fig. 7e-f). The gradational planar laminations are here interpreted to indicate high deposition rates. Experimental work shows that several mm to cm thick planar laminae are linked to migration of thick low-angle bedforms under high deposition rates (Paola et al., 1989; Cheel, 1990; Bridge and Best, 1997). Similarly, experimental work with turbidity currents indicates that planar laminations that formed at high sediment fallout rates are more diffuse or gradational, and several mm to a few cm thick (Sumner et al., 2008; Arnott, 2012). Not all planar laminations are gradational in seasonal river deposits, and in many cases they are thin, “flaky”, and with distinct lamina boundaries (Fig. 2c). Such planar laminations are a few grain diameters to a few mm thick and were deposited at lower sediment fallout rates (e.g., Paola et al., 1989; Cheel, 1990; Bridge and Best, 1997; Fielding, 2006).

High deposition rates are likely to be essential for accumulating large thicknesses of UFR deposits that otherwise have a very low preservation potential (Allen and Leeder, 1980; Bridge and Best, 1988, 1997; Paola et al., 1989; Cheel, 1990; Alexander et al., 2001), and thus high sediment flux and high deposition rates are characteristic for seasonal river deposits. High deposition rates can be readily recognized by gradational and climbing (aggradational) bedforms, but also by the thickness of single flood deposit and the degree of channel amalgamation (see further discussion in section 5). Including all these criteria, there is evidence for high deposition rates in 85% of the seasonal river deposits included in Table 1.

Furthermore, the climbing bedforms and gradational laminae signify deposition from suspension rather than bedload (Allen and Leeder, 1980). Suspension transport of sand to granule size sediment significantly increases coarse sediment transport rates and volumes, and emphasizes that seasonal rivers are likely to have a high coarse sediment yield compared to

367 rivers from perennial precipitation zones (see also Cecil et al., 2003). In some modern
368 subtropical rivers the suspended load has been suggested to be about 20 times higher than in
369 perennial rivers (Frostick and Reid, 1977). Sediment yield may be especially high in areas with
370 high-gradient drainage basins and sparse vegetation that promote flash flooding, landslide
371 initiation and floodplain erosion (Schumm, 1968; Stear, 1985; Fielding et al., 2009; Plink-
372 Björklund and Birgenheier, 2013). Together with high flow velocity, high deposition rates and
373 suspension fallout are also responsible for the relatively low abundance of dune cross
374 stratification, as bedform formation is suppressed in high sediment fallout conditions (e.g., Allen
375 and Leeder, 1980). Moreover, the high sediment load and resultant locally high deposition rates
376 are the main trigger of barrage (levee) breaks and catastrophic terrestrial flooding events in
377 seasonal rivers, such as the Kosi Flood in 2008 or Indus flood in 2010 (Sinha, 2009;
378 Brakenridge, 2012; Syvitski and Brakenridge, 2013; see further discussion in section 5).

380 4.3. *In-channel mud layers*

381 Another common feature in modern and ancient seasonal river deposits is the presence of in-
382 channel mud drapes and layers (e.g., Williams, 1971; Soleilhavoup, 1975; Tunbridge, 1981;
383 Sneh, 1983; Stear, 1985; Abdullatif, 1989; Olsen, 1989; Singh and Bhardwaj, 1991; Singh et al.,
384 1993; North and Taylor, 1996; Shukla et al., 2001; Billi, 2007; Nichols and Fisher, 2007; Fisher
385 et al., 2008; Fielding et al., 2009; Chakraborty and Gosh, 2010; Allen et al., 2011; Mader and
386 Redfern, 2011; Plink-Björklund and Birgenheier, 2013; Plink-Björklund et al., 2014a) (Fig. 9).
387 Note that these mud layers are different from a massive mud plug that signifies final channel
388 infilling, as the mud layers described here are interbedded with coarse sediments and thus part of
389 seasonally active channel-filling processes. In Table 1, 89% of the channel deposits are reported

to contain in-channel muds. There is only one case among the modern examples cited that does not demonstrate their presence, and of the ancient examples, in-channel mud has been reported in 85% of cases (Table 1).

The mud layers vary from veneers just a few mm thick (e.g., Williams, 1971; Stear, 1985; Singh and Bhardwaj, 1991; North and Taylor, 1996; Billi, 2007) (Fig. 9a-b) to several decimeters thick intervals (e.g., Abdullatif, 1989; Singh and Bhardwaj, 1991; Shukla et al., 2001) (Figs. 9c-d, 10). Mud layers occur as conformable undulating drapes (e.g., Stear, 1985; Turner, 1986; Abdullatif, 1989; Nichols and Fisher, 2007; Allen et al., 2011) (Fig. 10a-b,d), continuous layers (e.g., Sneh, 1983; Abdullatif, 1989; North and Taylor, 1996); discontinuous layers (e.g., Fisher et al., 2008), on lateral accretion surfaces (e.g., Stear, 1985; Hampton and Horton, 2007; Nichols and Fisher, 2007) (Fig. 9a-d), on lower parts of cross strata foresets and on bottomsets (e.g., Allen et al., 2011) (Figs. 4f, 10c), or as discrete lenses (e.g., North and Taylor, 1996; Fielding et al., 2009; Allen et al., 2011) (Fig. 9). In many cases, the mud layers show desiccation cracks (e.g., Tunbridge, 1981; Stear, 1985; Abdullatif, 1989; Olsen, 1989; Nichols and Fisher, 2007; Allen et al., 2011; Mader and Redfern, 2011) (Fig. 9c), pedogenic modification (e.g., Nakayama and Ulak, 1999; Shukla et al., 2001; Fielding et al., 2009; Allen et al., 2011; Mader and Redfern, 2011; Plink-Björklund and Birgenheier, 2013) (Fig. 9a) or root traces (e.g., Abdullatif, 1989; Shukla et al., 2001). In some cases the mud layers occur as curled-up fragments (e.g., Sneh, 1983) (Fig. 9e-g). Flaser-bedding-like structures have been described from the Ajay River, India (Bhattacharya, 1997). In most cases mud layers overlie sandy sediments conformably but sharply, but in some places sandstones grade into mudstones in upward-fining units (e.g., North and Taylor, 1996). Laterally, individual mud layers can be traced from just a few meters to at

least 100 m (Fig. 10a-b). Commonly the mud layers are discontinuous as they are eroded laterally (e.g., Abdullatif, 1989) (Fig. 10a, b).

Newly-formed mud drapes at the tops of bars, on accretion surfaces or at the channel bottom have been observed after major floods (e.g., Stear, 1985; Abdullatif, 1989; Singh et al., 1993; Billi, 2007) (Fig. 9a). For example, after a high magnitude flood of Kosi River in 1984, a 30 cm deep, 20 m wide and 100 m long mud layer was deposited in a channel (Singh et al., 1993).

Deposition of in-channel mud horizons has been linked to the characteristically rapid waning of high-magnitude floods. In some cases mud layers directly overlie UFR sedimentary structures such as planar laminations or low-angle convex-up laminations (Figs. 9a-c, g, 10d). In other cases climbing ripple laminated sets occur between the UFR structured units and the mud layers (Fig. 10b, d). Rapidly declining flows leave little trace of lower flow regime conditions, and thus UFR structures may be directly overlain by silt or mud drapes deposited at the final stage of the flood (Williams, 1971; Soleilhavoup, 1975; Tunbridge, 1981). Due to erosion during the peak of consecutive floods, the preservation of mud layers may be relatively low (Williams, 1971). Thus even mud layers that initially draped topography may be preserved only at topographically higher levels, such as bar tops which quickly emerge during waning floods (Fig. 10a-b). Mud layers are preserved if rapidly buried during later discharge events, rather than eroded (Allen et al., 2011). It is only the large-discharge high-magnitude floods that can drape topography, while smaller-discharge events result in mud being deposited in the deeper portions of the channel and on bar edges only (Allen et al., 2011) (Fig. 10c).

Mud drapes may be desiccated and pedogenically modified as bars become emergent after flood waters descend (e.g., Singh et al., 1993, 2007; Shukla et al., 2001). In rivers with perennial base discharge, like many rivers within the monsoon domain, only the upper levels in channels

are desiccated and pedogenically modified (e.g., Singh and Bhardwaj, 1991; Singh et al., 1993; Nakayama and Ulak, 1999; Shukla et al., 2001; Chakraborty and Gosh, 2010; Chakraborty et al., 2010). In channels that transmit discharge only during floods, like many rivers in the semi-arid and arid subtropical zones, the whole channel bottom may be desiccated, bioturbated and pedogenically modified (e.g., Tunbridge, 1981; Stear, 1985; Abdullatif, 1989; Fielding et al., 2009; Donselaar et al., 2013; Plink-Björklund and Birgenheier, 2013).

4.4. *Thick soft-sediment clast conglomerates*

Thick soft-sediment clast or intra-formational conglomerates have been suggested to be another common feature in seasonal river deposits (e.g., North and Taylor, 1996). These conglomerates differ from soft-sediment clast trains or lags that are a few grain-diameters thick and commonly line erosion surfaces also in perennial precipitation zone rivers. In Table 1, 48% of the river deposits contain conglomerates composed of soft-sediment clasts. The conglomerates range from just a few decimeters to multi-meter thick packages (Fig. 11a-c) that are common at the bases of the channels and above internal erosion surfaces (e.g., Deluca and Eriksson, 1989; North and Taylor, 1996; Zaleha, 1997; Hinds et al., 2004; Cain and Mountney, 2009; Plink-Björklund and Birgenheier, 2013; Banham and Mountney, 2014) (Fig. 11). Locally up to 4 m thick conglomerates have been documented (Hulka and Heubeck, 2010; Plink-Björklund and Birgenheier, 2013) (Fig. 11a-c). The conglomerates occur as restricted lenses or are up to several hundred meters in lateral extent (e.g., Deluca and Eriksson, 1989).

The conglomerates mostly consist of angular to rounded silt and mud clasts, commonly in granule to pebble size (Fig. 11). Up to meter-size clasts occur in places (e.g., North and Taylor, 1996). The clasts are locally derived and their composition recognized as, in places

pedogenically modified, floodplain or in-channel muds and silts. In some settings, the conglomerates contain caliche clasts, ferruginous concretions, sandstone clasts and wood fragments in addition to mud clasts (e.g., Turner, 1986; Deluca and Eriksson, 1989; Cain and Mountney, 2009; Plink-Bjorklund et al., 2014). The conglomerates vary from clast to matrix supported, and from disorganized to imbricated or stratified (e.g., North and Taylor, 1996; Plink-Björklund and Birgenheier, 2013; Banham and Mountney, 2014) (Fig. 11). In places there is a vertical transition from disorganized to stratified conglomerates (Fig. 11d). Most conglomerates have a sandy matrix, but in places silt and mud rich matrix occurs in chaotically organized conglomerates (e.g., Plink-Björklund and Birgenheier, 2013). The conglomerates may grade upward into sandstones (e.g., North and Taylor, 1996; Plink-Björklund and Birgenheier, 2013) (Fig. 12a), or are erosionally bounded below and above (Tunbridge, 1981).

Formation of thick soft-sediment clast conglomerates in seasonal rivers is linked to UFR flow conditions and rapid lowering of water level during the waning stage of floods (e.g., Gohain and Parkash, 1990; Singh et al., 1993; North and Taylor, 1996; Gibling and Tandon, 1997). In Kosi and Brahmaputra rivers, mud clasts form during early phases of the falling stage when UFR conditions occur, the banks are undercut, and develop shear cracks, whereby parts of the bank slump (Coleman, 1969; Gohain and Parkash, 1990; Singh et al., 1993) (Fig. 12b). Where the banks are overlain by, or contain, muddy sediment, the slumped sediments are cohesive and break into large blocks up to 2 m in length (Gohain and Parkash, 1990; Singh et al., 1993; Gibling and Tandon, 1997) (Fig. 12d). Areas about 300 x 200 m across with scattered slumped blocks have been observed at the end of the monsoon season in the Kosi River, suggesting large-scale erosion of the banks. The Brahmaputra had an average of three slumps per kilometer over a 50 km stretch of bankline after the 1963 and 1967-1968 monsoon seasons (Coleman, 1969). In

some areas, more than ten bank failures per kilometer occurred. The size of the slumps was quite variable, ranging from large features in excess of 300 m long to small 15-25 m long slumps (Coleman, 1969). The localized bank failures initially form isolated scattered large blocks (Fig. 12d), that then develop into angular and then to rounded clasts, when broken into smaller pieces and rolled along the channel floor for some distance (Singh et al., 1993) (Fig. 12e-f).

Flowage of bank material into channels is another failure mechanism. As observed in the Brahmaputra River (Coleman, 1969), when the water level falls rapidly, pore water in the channel banks develops overpressure and results in failure. This produces a bowl-shaped shear failure in the overlying cohesive deposits (Coleman, 1969) (Fig.12e). This type of failure is rather limited in extent compared to bank undercutting (Coleman, 1969), but may be responsible for formation of conglomerates or diamictons with a muddy or organic rich matrix. Similarly, porewater overpressure generates dendritic tributary patterns and sediment flowage, as small bowl-shaped failures propagate headward onto the backing floodplain, as observed after the Colorado flood in 2013 (Fig. 12g). Matrix-supported disorganized conglomerates are found where minimal reworking occurs after bank failure, whereas stratified conglomerates indicate current reworking and transport. Conglomerates may grade into sandstones if the soft-sediment clasts and sand are transported during the same flood, but are erosionally separated where deposited during separate flood events (Tunbridge, 1981).

Where sediments exposed in river banks are partially consolidated or contain caliche nodules, the conglomerates may also contain sand and caliche clasts (e.g., Gibling and Tandon, 1997). In Rupen River, India, erosion of a few thousand-years-old river bank sediments produces consolidated clast conglomerates (Gibling and Tandon, 1997) (Fig. 12d-e).

4.5. *Soft-sediment deformation structures*

Soft-sediment deformation has been documented in 39% of the modern and ancient seasonal river deposits included in Table 1. The deformation is abundant in some rivers (e.g., McKee et al., 1967; Coleman, 1969; Deluca and Eriksson, 1989; Bromley, 1991; Shukla et al., 2001; Hinds et al., 2004; Hampton and Horton, 2007; Fielding et al., 2009; Chakraborty and Gosh, 2010), rare in others (e.g., Sneh, 1983; Stear, 1985; Plink-Björklund and Birgenheier, 2013), and extremely rare in some (e.g., Williams, 1971). However, in contrast to the other criteria presented in this paper, soft-sediment deformation is also common in some perennial precipitation zone rivers, for example the Mississippi (e.g., Coleman, 1969). Thus, although included here, the significance of this criterion for recognizing variable-discharge rivers needs further evaluation. Furthermore, soft-sediment deformation may be connected to seismicity and earthquakes rather than river discharge regimes (e.g., Owen et al., 2011; Gosh et al., 2012). However, the scale of soft sediment deformation described below occurs on a single bed scale within a single bar form, rather than on multi-bed scale or as an extensive layer, as would be more likely the case if related to seismically-triggered soft-sediment deformation (see discussion in Owen et al., 2011).

In the datasets described here (Table 1), soft sediment deformation occurs as convolute bedding with vertical (e.g., Tunbridge, 1981) (Fig. 13a-e) or overturned limbs (e.g., McKee et al., 1967), or as overturned cross strata (Coleman, 1969; Bromley, 1991; North and Taylor, 1996) (Fig. 13f-g). Convolute bedding occurs as regularly or irregularly spaced folds that mainly have vertical limbs and may be ruptured in some cases (e.g., Tunbridge, 1981; Singh and Bhardwaj, 1991) (Fig. 13b). In places, the folds are weakly to moderately asymmetrical with overturned limbs (e.g., McKee et al., 1967). Convolute laminations occur in the upper part of

channel fill accretion units, on bar tops, as well as in floodplain deposits (Singh and Bhardwaj, 1991; Shukla et al., 2001; Chakraborty and Gosh, 2010). The convolute laminated units are a few decimeters to 0.5-1.0 m thick and commonly associated with planar laminations or climbing ripple laminations (e.g., McKee et al., 1967, Bromley, 1991; Singh and Bhardwaj, 1991; Plink-Björklund and Birgenheier, 2013). In places, climbing ripple-laminations have been observed to laterally transition into convolute bedding (McKee et al., 1967). Deformed zones can be more than 100 m wide (McKee et al., 1967). Overturned cross strata are recumbent folds, overturned in a downcurrent direction, less than 5 cm to a meter thick (Coleman, 1969; Bromley, 1991; North and Taylor, 1996).

Modern examples demonstrate that convolute bedding is a result of pore fluid expulsion generated during late flood stages when current velocities have slowed down and sediment is in the condition of quicksand (McKee et al., 1967), or during rapid loading by overlying sediments (Singh and Bhardwaj, 1991). The convolute laminations with downcurrent inclined axial planes and the overturned cross strata were formed during high flow through partial liquefaction and deformation of water-saturated sediment under the influence of lateral shearing (e.g., Stear, 1985; Singh and Bhardwaj, 1991). The latter is especially likely to occur in association with a rapid increase in bottom shear resulting from highly turbulent flow during abrupt transition from lower to upper flow regimes (Stear, 1985). The rapid rise and fall of river level, heavily sediment-laden currents, the fine-grained nature of the deposits, and intensive erosion and slope undercutting and steepening are likely to produce contortions of varying magnitudes in seasonal rivers (Coleman, 1969). A high percentage of the contorted stratification is associated with slumping and bank caving (Coleman, 1969). Some seasonal river deposits completely lack soft-sediment deformation or the deformation structures are extremely rare (Williams, 1971). A high

water table has been suggested as the extra condition necessary for the production of bedding disturbances. In arid conditions, such as in central Australia, water tables are generally low and the well drained alluvial sands contain no disturbances (Williams, 1971). Thus soft-sediment deformation is suggested to be less likely in well drained sandy sediments and in arid conditions than in the monsoonal domain or in sub-humid subtropical climates. However, even in well drained conditions groundwater seepage through flooring of bars or dunes, at the recession of floodwaters, may produce local quicksand bedding disturbances at the lower parts of the bedforms (Williams, 1971). In summary, although likely to occur in any river deposits with water level variations, soft-sediment deformation is suggested to be more common in seasonal rivers due to supercritical flow that promotes bank undercutting and bottom shear, and highly and rapidly variable water levels (in places full channel exposure) that promote porewater overpressure and deformation in channel banks and bars.

4.6. *In-channel vegetation and vegetation-induced sedimentary structures*

In-channel vegetation and vegetation-induced sedimentary structures have been suggested as another characteristic feature of seasonal river deposits, especially in sub-humid subtropics (Fielding et al., 2009; Allen et al., 2011, 2013, 2014). In Table 1, 20% of the seasonal river deposits, mostly in monsoonal and sub-humid subtropical climates, are reported to contain in-channel vegetation (e.g., Singh and Bhardwaj, 1991; Singh et al., 1993, 2007; Croke et al., 1998; Shukla et al., 2001; Fisher et al., 2008; Fielding et al., 2009; Chakraborty et al., 2010; Allen et al., 2011, 2013, 2014).

In-channel vegetation occurs as *in situ* and *ex situ* plant fossil adpressions and carbonaceous traces (Fielding et al., 2009; Allen et al., 2011, 2013, 2014) (Fig. 14). Such

intervals may be chaotically bedded or contain a variety of vegetation-induced sedimentary structures, such as centroclinal cross-stratification, bowl-shaped depressions, upturned beds around *in situ* tree stumps, scour-and-mound beds, and mudstone filled hollows with downturned beds (Rygel et al., 2004; Fielding et al., 2009; Allen et al., 2011, 2013, 2014) (Fig. 14). Some cases show downstream-tilted trunks and upstream anchoring root traces (Fielding et al., 2009; Allen et al., 2011, 2013, 2014). Other cases preserve closely spaced groups of upright lycopsid or calamitid trunks and/or stems, together with their radially propagated, mainly horizontal root systems (Fielding et al., 2009; Allen et al., 2011, 2013, 2014) (Fig. 14). Abundant allochthonous plant debris may also be common in close proximity to upright tree fossils (Fielding et al., 2009).

Colonization of the channel floor indicates periods of very low flow or lack of base flow, in order for vegetation to become established (Fielding et al., 2009). In the Burdekin River, Australia stable arborescent vegetation (*Melaleuca argentea*) occurs mainly in flow-parallel, linear groves (Fielding and Alexander, 1996; Fielding et al., 2009) (Fig. 14i). The trees adopt a prostrate, downstream-facing stance and have upstream-extending roots that give stability against flood (Fielding et al., 2009). The trees are often buried in the channel deposits, and some accumulate considerable logjams that can also become buried (Fielding et al., 1997, 2009; Alexander et al., 1999; Nakayama et al., 2002) (Fig. 14j). These groves engineer their environment and bedforms by trapping sediment (Fielding et al., 1997, 2009). The vegetation-induced sedimentary structures indicate that trees acted as obstacles to flow, generating areas of flow separation with associated scour and deposition (Fielding et al., 1997; Nakayama et al., 2002; Rygel et al., 2004). These structures are formed due to hydrodynamic interactions between living vegetation and the sediments (Rygel et al., 2004; Fielding et al., 2009; Allen et al., 2011, 2013, 2014).

4.7. In-channel trace fossils

Long, vertical, terrestrial trace fossils have been linked to seasonal climates (e.g., Hasiotis et al., 2002; Chakraborty et al., 2010), because the characteristic water table fluctuations between the dry and wet seasons are indicated by the length of the burrows. In most cases such traces are associated with flood plain deposits (e.g., Hasiotis, 2004). In some settings, however, they also occur in channels on accretion surfaces, and thus indicate in addition to the water table fluctuations that channels were dry for sustained periods between the floods (e.g., Uba et al., 2005; Saez et al., 2007; Plink-Björklund and Birgenheier, 2013) (Fig. 15). The in-channel vertical traces in the Eocene Wasatch and Green River Formations, USA, include multi-meter deep insect nests and crayfish burrows (*Camborygma*) (Fig. 15a-c). In places also *Ancorichnus* (polychaetes insect larvae burrows), *Scoyenia* (beetle larvae, cicada burrows) and adhesive meniscate burrows (insect burrows) occur on such surfaces (Fig. 15e). Root traces are another common in-channel trace fossil on accretion set boundaries. Root traces may occur on upper parts of bars only, if the channel maintains some base flow perennially (e.g., Singh and Bhardwaj, 1991; Singh et al., 1993, 2007; Shukla et al., 2001; Chakraborty et al. 2010), or occur throughout the channel fills, where channels are seasonally dry (e.g., Fielding et al., 2009; Allen et al., 2011, 2013, 2014; Plink-Björklund and Birgenheier, 2013; Bashforth et al., 2014). Variable degree of pedogenic modification accompanies the occurrence of terrestrial traces, in places indicating sustained subaerial exposure (e.g., Hasiotis, 2004) (Fig. 15f-g).

Trace fossils indicate specific soil moisture and water table conditions. Insect, ant and termite nests are constructed in subaerial conditions and thus indicate channel bed exposure. Deeper nests signify lower soil moisture and water-table levels (Hasiotis, 2004). *Camborygma*

burrows also indicate subaerial exposure, and were constructed by freshwater and terrestrial crayfish (Hasiotis and Mitchell, 1993). More than meter long *Camborygma* traces indicate low water tables, and that sediments were often subaerially exposed (Hasiotis, 2004). The adhesive meniscate burrows are constructed by soil bugs in sediment undergoing pedogenesis with moisture levels of 10–37%, and their occurrence has been linked to the A and uppermost B soil horizons (Hasiotis, 2004). *Ancorichnus* are insect burrows, most likely constructed by a beetle living at or just below the sediment–air interface, in sediments that were very moist to saturated (50–100% porewater), and signify the proximity of the capillary fringe and the water table (Hasiotis, 2004). *Scoyenia* is constructed by deposit-feeding insect larvae and indicate very high soil moisture, approaching 100% saturation of freshwater (Hasiotis, 2004). The latter traces (*Ancorichnus*, *Scoyenia* and adhesive meniscate burrows) can occur any type of river systems. However, in combination with palaeosols that only indicate sustained dry conditions, they bear vestige of the wet monsoon seasons. The long vertical burrows are indicative of seasonally wet climates with large water table fluctuations.

5. Characteristic architecture

Below follows a synthesis of architectural characteristics that are common in modern and ancient monsoonal domain and subtropical rivers (Table 2). The abundance of flood units (5.1), poorly developed macroforms or bar forms (5.2), and the dominance of avulsions (5.3) over systematic channel migration, were each reported in more than 50% of the datasets included in Table 1.

5.1. Flood units

In Table 1, 54% of the datasets report distinct depositional units, bounded by erosion surfaces, or by an erosion surface below and a mud drape above. In systems with conglomeratic sediments, these depositional units are 0.5 m to multi-meter thick, and fine upwards from conglomerates to sandstones (e.g., Turner, 1986; Rhee and Cough, 1993; Nichols and Hirst, 1998; Hinds et al., 2004; Uba et al., 2005; Chakraborty and Gosh, 2010; Chakraborty et al., 2010; Mader and Redfern, 2011) (Fig. 16a-e). In the Ganga River, some depositional units consist of erosionally based conglomerates with root traces at the top (Singh and Bhardwaj, 1991).

In sandy channel fills and within the sandy parts of coarser systems, there is a considerable variability in such depositional units, as they may contain one or all of the following stratigraphic elements from base up: (a) scour and fill to mud drape (Fig. 10d), (b) scour and fill to planar laminae (Figs. 16f, g, 17e), (c) scour and fill to planar laminae to climbing ripples (Figs. 10d, 17d), (d) scour and fill to climbing ripples (Fig. 17b), (e) convex-up low angle laminae, (f) convex-up low angle laminae to climbing ripples (Figs. 17b, 18b), (g) convex-up low-angle to planar laminae to climbing ripples (Fig. 18b), (f) planar laminae to climbing ripples (Figs. 10d, 17c, f-h, 18b), (e) planar laminae (Figs. 16f, 17a), (g) cross strata to mud drapes (Fig. 18a), and (h) climbing ripples to mud drapes (e.g., Tunbridge, 1981; Sneh, 1983; Turner, 1986; Abdullatif, 1989; Deluca and Eriksson, 1989; Nichols and Hirst, 1998; Shukla et al., 2001; Hinds et al., 2004; Hampton and Horton, 2007; Chakraborty and Gosh, 2010; Chakraborty et al., 2010; Mader and Redfern, 2011; Donselaar et al., 2013; Plink-Björklund and Birgenheier, 2013; Plink-Björklund et al., 2014a). Any of these depositional unit types may or may not be capped by a mud drape. The depositional units do not necessarily fine upwards, other than at the tops of the units, where the transition from UFR structures to ripples or to mud drape occurs (Deluca and

Eriksson, 1989; Billi, 2007). This transition from sand to mud is gradational in some river deposits, as a muddy sandstone occurs at the top of the unit rather than a distinct mud drape or layer (e.g., Chakraborty and Gosh, 2010). In some cases convoluted laminations (e.g., Sneh, 1983; Chakraborty and Gosh, 2010) or structureless sandstones (Nichols and Hirst, 1998) occur at the top of the depositional units. The tops of the units are quite commonly erosively truncated (e.g., Banham and Mountney, 2014) (Fig. 10a-b, d).

The sand to mud couplets tend to be especially well developed in channels with lateral accretion sets (Figs. 9d-e, 19a, d). The accretion deposits form inclined heterolithic strata, as each accretion unit consists of a sand and mud couplet, a few dm to multiple meters thick (e.g., Chakraborty and Gosh, 2010; Chakraborty et al., 2010; Plink-Björklund and Birgenheier, 2013). Thickness of the depositional units that do not form lateral accretion sets, but rather occur as downstream or vertically accreted units, varies from a few dm to 5-10 m (e.g., Sneh, 1983; Stear, 1985; Turner, 1986; Nichols and Hirst, 1998; Hinds et al., 2004; Hampton and Horton, 2007; Banham and Mountney, 2009; Chakraborty and Gosh, 2010; Chakraborty et al., 2010; Plink-Björklund and Birgenheier, 2013) (Figs. 10, 16-19). Lateral extent varies from tens of meters to more than 100 m (Hinds et al., 2004; Chakraborty and Gosh, 2010; Chakraborty et al., 2010; Plink-Björklund and Birgenheier, 2013) (Fig. 18c). The tops of the depositional units are in places desiccated and pedogenically modified (Fig. 15).

Common to all these depositional units is an abrupt transition from UFR structures to climbing ripples or to a mud drape, indicating rapid flow deceleration, characteristic of flood events (e.g., Chakraborty and Gosh, 2010). Lack of ripples between the transitions from UFR structures to overlying mud drapes indicates an extremely rapid decline in flow strength, with no low-stage reworking of the flood deposit (e.g., Jones, 1977; Tunbridge, 1981). Because of this

vertical transition, the commonly erosional base, and the desiccation and pedogenic modification at the top of the units, they are commonly interpreted as deposits of single flood events (e.g., Tunbridge, 1981; Sneh, 1983; Turner, 1986; Abdullatif, 1989; Deluca and Eriksson, 1989; Nichols and Hirst, 1998; Shukla et al., 2001; Hinds et al., 2004; Hampton and Horton, 2007; Chakraborty and Gosh, 2010; Chakraborty et al., 2010; Mader and Redfern, 2011; Donselaar et al., 2013; Plink-Björklund and Birgenheier, 2013; Plink-Björklund et al., 2014a). However, observations from the Laingsburg flood, South Africa, in 1981 show that multiple depositional units may form during a single flood due to multiple flood waves (Stear 1985) (Fig. 18c). If bounded by desiccation, bioturbation or pedogenic features, the flood units can conclusively be identified as corresponding to individual flood events. The flood units may be interbedded with lower-discharge deposits in rivers that transmit perennial base discharge, or experience low- as well as high-magnitude floods, whereas deposits are likely to entirely consist of flood units in rivers that transmit effective discharge only during high-magnitude floods.

5.2. *Poorly developed macroforms*

Lack of barforms or well developed accretion sets has been suggested as one of the characteristic features of rivers with seasonally variable discharge (e.g., Rhee and Cough, 1993; Uba et al., 2005; Billi, 2007; Hampton and Horton, 2007; Fielding et al., 2009; Hulka and Heubek, 2010; Allen et al., 2011). In Table 1, half of the papers do report the presence of downstream or lateral accretion sets or bars, but 58% note that macroforms are not well developed and vertical or gentle downstream accretion prevails (e.g., Tunbridge, 1981; Stear, 1985; Olsen, 1989; North and Taylor, 1996; Zaleha, 1997; Thomas et al., 2002; Kumar et al., 2003; Allen et al., 2013, 2014; Plink-Björklund and Birgenheier, 2013) (Figs. 18a, 20a).

Low-angle downstream accretion sets are relatively common (e.g., Williams, 1971; Abdullatif, 1989; Deluca and Eriksson, 1989; Bromley, 1991; Kumar, 1993; Sadler and Kelly, 1993; Willis, 1993; North and Taylor, 1996; Zaleha, 1997; Nichols and Hirst, 1998; Nakayama and Ulak, 1999; Uba et al., 2005; Nichols and Fisher, 2007; Leleu and Hartley, 2010; Leleu et al., 2010; Allen et al., 2013, 2014; Plink-Björklund and Birgenheier, 2013; Trendell et al., 2013; Plink-Björklund et al., 2014a), but may be difficult to observe in small outcrops (Fig. 19). The downstream accretion sets record very rapid deposition due to flow expansion. Such low-angle and poorly developed accretion sets contrast with the systematic bar migration and formation of well-developed accretion sets, which are common in existing facies models, based on perennial precipitation zone river systems. Vertical aggradation (e.g., Tunbridge, 1981; Stear, 1985; Olsen, 1989; Zaleha, 1997; Thomas et al., 2002; Kumar et al., 2003; Allen et al., 2013, 2014) (Fig. 18a) dominates some seasonal river systems, and is formed due to vertical channel bed aggradation. Lateral accretion sets occur in numerous settings, especially in downstream reaches of rivers and in crevasse channels, but are mostly reported as subordinate rather than dominant (e.g., Williams, 1971; Tunbridge, 1981; Stear, 1985; Abdullatif, 1989; Deluca and Eriksson, 1989; Olsen, 1989; Bromley, 1991; North and Taylor, 1996; Zaleha, 1997; Nichols and Hirst, 1998; Newell et al., 1999; Ray and Chakraborty, 2002; Thomas et al., 2002; Kumar et al., 2003; Uba et al., 2005; Nichols and Fisher, 2007; Saez et al., 2007; Leleu and Hartley, 2010; Leleu et al., 2010; Allen et al., 2013; Donselaar et al., 2013; Plink-Björklund and Birgenheier, 2013; Trendell et al., 2013; Plink-Björklund et al., 2014a) (Figs. 9a-d, 19a). In many cases the lateral accretion sets transition into vertical accretion across short distances (Figs. 9c, 10c). In some cases such as in the Ganga River channel bank exposures, in the braided zone, low-angle downstream, upstream and vertical accretion sets occur (Shukla et al., 2001) (Fig. 15a).

Modern seasonal rivers form sand waves and transverse and longitudinal bars in braided and anastomosing channels, and point bars in meandering channels that become emerged during low flow (e.g., Coleman, 1969; Williams, 1971; Abdullatif, 1989; Singh and Bhardwaj, 1991; Singh et al., 1993, 2007; Shukla et al., 2001; Jain and Sinha, 2004; Fielding et al., 2009; Chakraborty and Ghosh, 2010; Chakraborty et al., 2010) (Fig. 20d-e), and thus seem to contradict with the preserved sedimentary strata that tend to show poorly developed macroforms. Even recent deposits of the Tista River, Himalayan Foreland, exposed in channel banks, contain laterally continuous vertically aggrading units rather than bar slipface cross strata (e.g., Chakraborty and Ghosh, 2010) (Fig. 18a).

Rapid discharge variations of floods enhance rapid deposition and aggradation (Abdullatif, 1989), whereas sustained flow conditions are needed for systematic bar migration. Furthermore, high concentration of sediment carried in suspension during floods inhibits bar formation (Allen and Leeder, 1980), because of the effect of suppressing flow separation and turbulence involved in the formation of bar growth (North and Taylor, 1996). In some modern systems bars form from lower flow regime conditions, during perennial base flow or during the initial flood stage, whereas during UFR conditions of the peak flood, plane bed transport and deposition prevail (e.g., Kosi River; Gohain and Parkash, 1990). Thus, the initial lower-flow regime bedforms are not preserved, as sediment gets remobilized during the peak flood. In the Burdekin River, the short-duration, large discharge events ($<15,000 \text{ m}^3 \text{ s}^{-1}$) have the capacity to completely mobilize alluvium down to bedrock (Fielding et al., 1999; Amos et al., 2004; Allen et al., 2014). If lower flow regime conditions prevail during the post-peak flow, the channel bed is exposed, incised and dissected, and in places bar migration and aggradation can occur (Gohain and Parkash, 1990) (Fig. 20d, e). In summary, although bar forms can be observed to develop in modern seasonal

756 rivers, especially during more sustained lower flow regime conditions, their deposits are
757 dominated by low-angle downstream and vertical aggradation sets due to prevailing UFR
758 conditions and high deposition rates during peak conditions of high-magnitude floods
759 (Chakraborty and Ghosh, 2010; Chakraborty et al., 2010). This is in contrast with the Mississippi
760 for example, where flooding conditions are also common, and stage changes expose bars during
761 low flow (e.g., Coleman, 1969). However, here the bar forms are active during the floods, and
762 inactive during low flow, rather than being completely or partially remobilized.

764 5.3. *Avulsions*

765 In Table 1, at least 58% of the datasets demonstrate that avulsions, rather than channel migration,
766 dominate river morphodynamics. Avulsions are a key component of the modern monsoonal and
767 subtropical river behavior (e.g., Abdullatif, 1989; Makaske, 2001; Jain and Sinha, 2003, 2004;
768 Assine, 2005; Billi, 2007; Assine and Silva, 2009; Sinha, 2009; Chakraborty et al., 2010;
769 Chakraborty and Gosh, 2010; Donselaar et al., 2013). Avulsion involves a sudden channel shift
770 at a nodal point and occurs when an event of sufficient magnitude (e.g., a flood) occurs along a
771 river that is at or near avulsion threshold (Jones and Schumm, 1999), defined by the changing
772 channel instability through time (Sinha, 2009). This implies that even a small flood can trigger
773 avulsion if a river is close to avulsion threshold. Experiments and analyses of modern and
774 ancient systems show that channel bed superelevation, that is a function of channel bed
775 aggradation (deposition rates), is the main trigger of avulsions (Bryant et al., 1995; Mohrig et al.,
776 2000; Makaske, 2001; Stouthamer and Berendsen, 2001; Jerolmack and Mohrig, 2007; Sinha,
777 2009; Hajek and Edmonds, 2014). Incisional avulsions also occur (Mohrig et al., 2000; Hajek
778 and Edmonds, 2014), and in both cases large fluctuations in river discharge and the occurrence

of floods serve as avulsion-triggering events (Jones and Schumm, 1999). Avulsion may occur as rapid paleo-channel reoccupation, when the avulsed channel occupies a pre-existing channel in the vicinity. Avulsions may also occur by crevasse splaying that involves a more gradual process of breaching through the banks and development of a new channel through time. For example, a major Kosi River avulsion in 2008 occurred during the annual monsoon flood, where the Kosi River shifted more than 100 km eastward (Sinha, 2009; Sinha et al., 2013). The avulsed channel ‘reoccupied’ one of its former palaeochannels and 80–85% flow of the river was diverted into the new course. Since the new course had a much lower carrying capacity, the water flowed like a sheet, 15–20 km wide and 150 km long with a velocity of 1 m s^{-1} at the time of breach. The avulsion occurred as a result of channel bed aggradation and superelevation and a resultant barrage (constructed levee) break (Sinha, 2009). This single event affected more than 30 million people, caused >400 fatalities, displaced 10 million people, and caused severe loss of property (Sinha, 2009; Brakenridge, 2012). The Indus River avulsion in 2010 flooded $\sim 8000 \text{ km}^2$ of agricultural land as part of the river system avulsed 50–100 km west of its previous location (Brakenridge, 2012; Syvitski and Brakenridge, 2013). Similarly, barrage failure during seasonal monsoon flood, due to channel bed aggradation and superelevation, led to the avulsion (Syvitski and Brakenridge, 2013).

In the geological record, avulsions are evidenced by the dominance of downstream and vertical accretion, and thus lack of systematic channel migration, as well as the lack of well-developed macroforms (e.g., Tunbridge, 1981; Stear, 1985; Olsen, 1989; Rhee and Cough, 1993; North and Taylor, 1996; Zaleha, 1997; Thomas et al., 2002; Kumar et al., 2003; Uba et al., 2005; Billi, 2007; Hampton and Horton., 2007; Fielding et al., 2009; Hulka and Heubek., 2010; Allen et al., 2011, 2013, 2014; Plink-Björklund and Birgenheier, 2013; Plink-Björklund et al., 2014a).

Furthermore, frequent avulsions result in a high degree of channel amalgamation in areas with high channel return frequency, and in splay-rich floodplains in areas with lower channel return frequency (e.g., Shukla et al., 2001; Alexander and Fielding, 2006; Hajek and Wolinsky, 2012; Plink-Björklund and Birgenheier, 2013; Plink-Björklund et al., 2014a) (Fig. 21). Channel return frequency is a function avulsion frequency and the area across which a channel avulses (see e.g., Hajek and Wolinsky, 2012).

In summary, avulsions are common in monsoonal and subtropical rivers due to the frequent occurrence of floods. The high and highly variable flood discharges and the high sediment concentrations result in rapid channel bed aggradation and resultant superelevation that is a key threshold for occurrence of avulsions.

6. Link to fluvial fan systems

In Table 1, 60% of the river systems are identified as fluvial fans (Table 1), as fluvial megafans (Singh et al., 1993, 2007; Kumar, 1993; Willis, 1993; Zaleha, 1997; Nakayama and Ulak, 1999; Shukla et al., 2001; Thomas et al., 2002; Kumar et al., 2003; Uba et al., 2005; Chakraborty et al., 2010; Chakraborty and Ghosh, 2010; Plink-Björklund and Birgenheier, 2013; Bashforth et al., 2014; Plink-Björklund et al., 2014a), terminal fans (Abdullatif, 1989; Bromley, 1991; Kelly and Olsen, 1993; Rhee and Cough, 1993; Sadler and Kelly, 1993; North and Taylor, 1996; Newell et al., 1999), terminal distributary systems (Billi, 2007), fluvial distributary systems (Nichols and Hirst, 1998; Nichols and Fisher, 2007; Fisher et al., 2008), distributive fluvial systems (Leleu and Hartley, 2010; Leleu et al., 2010), and undifferentiated fluvial fans (Turner, 1986; Croke et al., 1998; Hampton and Horton, 2007; Nielsen et al., 2007; Saez et al., 2007; Donselaar et al., 2013). Fluvial fans are large fan shaped sedimentary bodies that are built by rivers (Fig. 22).

They are distinct from alluvial fans as they have a low gradient, are considerably larger, fine outward to silts and muds, and do not contain sediment gravity flow deposits (e.g., DeCelles and Cavazza, 1999; Horton and DeCelles, 2001; Leier et al., 2005). Fluvial fan deposits are recognized by their large lateral extent ($>10^4$ km²), and characteristic vertical and lateral facies transitions. In fluvial megafans, lateral facies transitions characteristically reflect updip to downdip changes in channel character and size (e.g., from large braided to smaller anastomosing and to even smaller meandering channels), as well as in the channel to floodplain proportion (e.g., Singh et al., 1993; Shukla et al., 2001; Assine and Silva, 2009; Chakraborty and Gosh, 2010) (Fig. 22). Proximal parts of the fans, where the fan area is smaller and channel return frequency higher, are dominated by amalgamated channels, whereas the distal parts are dominated by floodplain fines (Shukla et al., 2001). Crevasse splays are a common floodplain component, especially in the anastomosing zone (Plink-Björklund and Birgenheier, 2013; Plink-Björklund et al., 2014a). Vertical facies transitions indicate fan progradation, aggradation or backstepping by vertical transitions from floodplain-dominated successions with small channels to amalgamated large channels (e.g., DeCelles and Cavazza, 1999; Uba et al., 2005; Chakraborty et al., 2010).

Terminal fans, terminal distributary systems and fluvial distributary systems are conceptually suggested to form by bifurcations (channel splitting) rather than avulsions, and linked to high rates of evapotranspiration (e.g., Kelly and Olsen, 1993; Rhee and Cough, 1993; Sadler and Kelly, 1993; Nichols and Hirst, 1998; Nichols and Fisher, 2007; Fisher et al., 2008). However, this concept has received significant critique, as the proposed modern analogues have shown to build by avulsions rather than bifurcations, and both processes can generate similar morphological patterns (see discussion in North and Warwick, 2007).

Modern fluvial megafans occur in the monsoonal domain and in the subtropical zone as a function of the highly seasonal discharge and the associated high-magnitude floods that trigger frequent avulsions (Assine, 2005; Leier et al., 2005). Ancient megafans indicate that the most active fan construction episodes are linked to an increase in monsoon intensity, such as between ca 10 and 8 Ma in the Himalayas (Dettman et al., 2000, Leier et al., 2005). Occurrence of terminal fans, terminal distributary systems, and fluvial distributary systems has also been linked to arid to semi-arid subtropical conditions (e.g., Abdullatif, 1989; Nichols and Hirst, 1998; Hinds et al., 2004; Billi, 2007; Nichols and Fisher, 2007; Fisher et al., 2008). High evapotranspiration and water losses to floodplain and seepage into the fan have been suggested to control the formation of terminal fans and terminal distributary systems (e.g., Abdullatif, 1989; Billi, 2007).

In summary, there is a tendency for monsoonal domain and subtropical zone rivers to form fluvial fans, if they have large enough discharge and high enough sediment supply (see discussion in Leier et al., 2005). However, not all seasonal rivers form fluvial fans. Some rivers may enter narrow or small basins and cannot migrate laterally and form fan-shaped sediment lobes. The spacing between channel outlets can also be an important factor, because if the outlets are closely spaced (e.g., Ravi and Chenab Rivers in the western Himalayan foreland), channel migration is limited by the adjacent river's deposits (Geddes, 1960; Leier et al., 2005). Some of the large rivers are fed by fluvial megafans rather than forming fan-shaped sedimentary bodies themselves, such as the Ganges in the Himalayas (e.g., Sinha et al., 2010), or the Paraguay in South America (Assine, 2005; Assine and Sylva, 2009) (Fig. 22).

7. Forcing factors on morphodynamics of rivers strongly affected by monsoon precipitation

871 Based on the above synthesis of facies and architectural elements common in rivers that receive
872 most of their surface water from the intense seasonal monsoon precipitation, their
873 morphodynamics can be concluded to be a function of: (1) dominant deposition from floods with
874 Froude transcritical and supercritical flow conditions; (2) the characteristically rapid rise, sharp
875 peak and rapid decline of the flood hydrograph; (3) rapid lowering of water level during the
876 falling stage of floods; (4) lack of efficient perennial base flow; (5) seasonal partial or complete
877 channel bed exposure; (6) high sediment concentrations during floods; (7) high suspended
878 sediment load (including sand and gravel) during floods; (8) water oversaturation of floodplain
879 sediment during floods; (9) seasonally variable water table; (10) high evapotranspiration; and
880 (11) water loss through seepage and floodplain flooding. All these features are the result of
881 precipitation seasonality, characteristic for the monsoonal domain and the subtropical zone (Fig.
882 1). All these characteristics are linked to the inherent large discharge range of monsoonal and
883 subtropical rivers as compared to the mean (Fig. 1f-i). The positive discharge deviation (Fig. 1i)
884 is a measure of flood magnitude that is considerably higher in monsoonal and subtropical rivers
885 as compared to rivers in the perennial precipitation zone. This large positive deviation
886 demonstrates that flood discharge is considerably higher than the annual mean, and even more so
887 when compared to the low dry-season base flow. Thus, the flood conditions in such rivers are
888 more different as compared to the non-flood conditions, and in most cases sediment is
889 transported and deposited only during the seasonal flood events. Sediment transport and
890 deposition in perennial precipitation zone rivers is also linked to floods, but the flood conditions
891 are not as extreme compared to the annual mean or the low-flow conditions. The Froude
892 supercritical flow and the dominant suspension transport further strengthens that effect, as
893 sediment is flushed through the system and rapidly deposited, preserving individual flood events,

rather than gradually and successively transporting sediment as bedload, and thus inevitably subject to hysteresis effects. The large magnitude of floods together with rapid floodwater waning are the key for triggering channel bank instability and collapse, and produce soft-sediment deformation and soft-sediment clast conglomerates.

There is a considerable variability in the analyzed monsoonal and subtropical river sediments as can be deduced from the presence of the characteristic facies and architectural elements. Only 6% of the examples included in Table 1 contain all 11 characteristics (Fig. 23). However, 89% of examples contain at least 5 of the characteristics (Fig. 23). The majority (60%) of the river deposits contain 5-7 characteristics, whereas 29% contain 7 or more, and only 11% 4 or fewer (Fig. 23). Furthermore, the proportion of the characteristic facies and architectural elements also varies considerably between different river systems (e.g., Fig. 6). Below, the seasonal river deposits are subdivided into four categories to illustrate some of this variability and to discuss their varied forcing factors. These sub-categories should be treated as examples or end members, rather than distinct facies models.

7.1. Amalgamated lithosomes with thick flood units

These lithosomes consist of up to 100% of Froude transcritical and supercritical flow (UFR) sedimentary structures and high deposition rate (HDR) sedimentary structures (e.g., examples 1-2 in Fig. 6), lack mudstones, and contain simple, thick low-angle, erosionally bounded flood units (e.g., Tunbridge, 1981; Stear, 1985; Billi, 2007; Plink-Björklund and Birgenheier, 2013) (Figs. 2a, d, e; 17h; 18a; 19b, c; 24a). The lithosomes have a massive or uniform appearance, and form laterally and vertically amalgamated, tabular sandstones (e.g., Stear, 1985; Plink-Björklund and Birgenheier, 2013) (Figs. 19b, 21a). In the Uinta Basin the flood units consist of 82% of

UFR and 86% HDR structures (example 2 in Fig. 6), are up to 12 m thick, and uniform for hundreds of meters laterally. Long vertical burrows (Fig. 10a), and in places pedogenic modification (Fig. 10f, g), occur on the erosional bounding surfaces. Literature examples cite up to 90-100% UFR structures (example 1 in Fig. 6; e.g., Tunbridge, 1981; Billi, 2007). As these lithosomes lack finer-grained sandstones or mud drapes, the only lithologic heterogeneity is provided by thick mud-clast conglomerates that occur at the bases of the channels and of the erosionally-bound flood units (Sneh, 1983; Stear, 1985; Deluca and Eriksson, 1989; North and Taylor, 1996) (Figs. 16g, 17a). Macroscopic plant fragments or in-channel vegetation is absent. These lithosomes resemble flash flood or megaflood deposits (e.g., Schumm, 1961; McKee et al., 1967; Williams, 1971; Stear, 1985; Billi, 2007) (Fig. 24a), rather than confirming to existing fluvial facies models for the perennial precipitation zone (e.g., Collinson, 1996; Miall, 1996; Bridge, 2006) or even for subtropical rivers (Fielding et al., 2009, Allen et al., 2013, 2014).

The dominance of the flood units signifies that deposition of these lithosomes occurred only during floods. The dominance of UFR and HDR sedimentary structures indicates prevailing deposition at high flow velocities and high deposition rates. Lack of transitions from UFR structures to ripples or dunes implies extremely rapid decline in flow strength (Jones, 1977). The downstream accreting flood units reflect very rapid deposition due to flow expansion in a downstream direction across previous flood deposits. The thick erosionally-bounded flood units indicate initial erosion and local deposition of thick accumulations during single flood events. The lack of systematic bar migration and the high degree of amalgamation signify high avulsion rates. The thick conglomerates indicate high floodplain erosion and channel bank collapse and flowage rates. Collectively, these lithosomes consist entirely of high-magnitude flood deposits, with no evidence for low-magnitude or non-flood discharge, and this is why they resemble

flashflood deposits (cf., Tunbridge, 1981; Stear, 1985; Billi, 2007). The in-channel bioturbation and pedogenic modification indicate that the channels only transmitted discharge during floods and had no perennial base flow. The length of vertical burrows (e.g., insect nests and *Camborygma*) demonstrates fluctuating but sustained low water tables (Hasiotis et al., 2002).

Modern examples of this type of lithosome occur where the total annual precipitation falls in a few downpours that last from hours to a few days, whereas mean annual rainfall amounts vary. Examples include the Kobo Basin, Ethiopia where the mean annual rainfall is 750 mm (Billi, 2007), Sinai Peninsula and Karoo (Sneh, 1983; Stear, 1985), where the mean annual rainfall is less than 100 mm. Such a precipitation pattern is inevitably linked to intermittent droughts that may last for most of a year (Billi, 2007), to decades (Sneh, 1983; Stear, 1985). Droughts impede vegetation cover. The intense rainfall, together with sparse vegetation, promotes flash floods, high erosion rates, and high sediment flux. Such precipitation extremes transform rivers into particularly flashy systems that can experience 100% of sediment transport and deposition during high-magnitude floods, and are prone to rapid suspension deposition and avulsions. Such channels transmit discharge only in response to the extreme rainfall events, and are intermittently dry.

7.2. *Lithosomes with variable flood units*

These channel lithosomes also consist of stacked flood units, but the sedimentary structures and architectural elements vary considerably both vertically and laterally. Characteristically, the UFR sedimentary structures are overlain by lower flow regime (LFR) structures such as climbing ripples and/or mud drapes (e.g., McKee et al., 1967; Stear, 1985; Abdullatif, 1989; Deluca and Eriksson, 1989; Olsen, 1989; Bromley, 1991; North and Taylor, 1996; Croke et al., 1998;

Thomas et al., 2002; Kumar et al., 2003; Chakraborty and Gosh, 2010; Plink-Björklund and Birgenheier, 2013; Plink-Björklund et al., 2014a) (e.g., Figs. 9c, 10d, 17, 18c, 19a, d, 24b). These lithosomes have a lower proportion of UFR and HDR sedimentary structures, and a higher proportion of ripples or dunes (e.g., examples 3-5 in Fig. 6), as compared to the lithosomes with thick flood units. The flood units are not as highly amalgamated as in the latter lithosomes, due the presence of finer-grained sandstones in which the lower flow regime structures occur, and the presence of mud drapes or layers (Figs. 10, 19a, d). In some cases there is also a greater lateral variability of facies and internal architecture (e.g., Fig. 10a, d), whereas other examples display laterally extensive flood units (e.g., Fig. 19d). Flood unit thickness varies from a few decimeters (Fig. 9c) to multi-meter thick packages (Fig. 19d). In places long vertical burrows (Fig. 15c) occur on the bounding surfaces, in other places the mud layers are pedogenically modified (Fig. 9c). Macroscopic plant fragments are common in some systems and absent in others.

Also in these lithosomes, the flood units signify that deposition dominantly occurred during high magnitude floods. However, the transition from UFR to LFR structures and to mud layers indicates a less rapid decline in flow strength compared to the uniform lithosomes (Jones, 1977). In-channel bioturbation and pedogenic modification is common along the whole flood units, and shows that the channels only transmitted discharge during floods and had no base flow. Where bioturbation and pedogenic modification only occur in the upper parts of the channel fills, perennial base flow occurred.

Modern examples range from monsoonal domain rivers (e.g., Singh and Bhardwaj, 1991; Singh et al., 1993; 2007; Shukla et al., 2001; Jain and Sinha, 2004; Chakraborty et al., 2010; Chakraborty and Gosh, 2010), where mean annual precipitation may exceed 1100-1600 mm (Sinha et al., 2014), to arid subtropics with mean annual precipitation of only a few hundred mm

(e.g., Williams 1971; Sneh, 1983; Fisher et al., 2008). Similar to the uniform lithosomes, the variable lithosomes with flood units reveal the lack of efficient base flow, and prevailing deposition from high-magnitude floods.

7.3. *Lithosomes with flood units and lower flow regime deposits*

These lithosomes display a lateral and vertical alternation of flood units and dominantly cross-stratified intervals, where the latter resemble perennial river deposition (Williams, 1971; Sneh, 1983; Abdullatif, 1989; Nakayama and Ulak, 1999; Shukla et al., 2001; Donselaar et al., 2013) (e.g., examples 7-8 in Fig. 6; Figs. 10b; 18b; 23c). Although still abundant, the content of UFR and especially the HDR structures is considerably lower, with 52% and 36% respectively in the Uinta Basin, and with 49% of UFR structures in River Gash (Abdullatif, 1989; examples 7-8 in Fig. 6). Cross stratification is more common, with 23% on average in the Uinta Basin (Fig. 1), and 36% in River Gash deposits (Abdullatif, 1989). Moreover, vertical and lateral transitions between distinct and gradational or climbing UFR structures, cross strata and ripples, are common. In places such lithosomes have better developed macroforms and steeper accretion sets (e.g., Figs. 10b, 19a). Plant and wood fragments are more common, and bioturbation on flood unit boundaries is more variable and includes *Ancorichnus*, *Scoyenia* and adhesive meniscate burrows in places, in addition to the long vertical burrows. Although with higher cross-strata content, these lithosomes are distinct from perennial precipitation zone rivers that are characteristically dominated by cross stratified sandstones (80-100% in the Uinta Basin; example 9 in Fig. 6), have well-developed macroforms, lack in-channel bioturbation, and have conglomerates as thin lags along erosion surfaces (cf., Collinson, 1996; Miall, 1996; Bridge, 2006) (Table 2).

1009 Variability in sedimentary structures, lithology and accretion set thickness and orientation
1010 all signify frequent variations between high and low flow velocities and deposition rates, linked
1011 to flood stage variations (Abdullatif, 1989). Variations between distinct and gradational UFR
1012 structures indicate variations in deposition rates during the high magnitude floods, as a larger
1013 proportion of sediment is transported and deposited during lower-magnitude events. Simple
1014 downstream sediment dumping alternated with systematic subaqueous dune or bar migration
1015 during more sustained flow conditions. The in-channel mudstones and ripple bedforms indicate
1016 preservation of waning-phase deposits. The occurrence of cross strata in the transition from UFR
1017 strata to ripples indicates a less rapid decline of flood strength. The in-channel burrows and
1018 pedogenic modification indicate that channels were partially or fully dry at times. The
1019 *Ancorichnus*, *Scoyenia* and adhesive meniscate burrows indicate high soil moisture content and
1020 sustainably high water table (Hasiotis et al., 2002). The presence of plant debris suggests riparian
1021 tree growth and a more abundant vegetation cover, and signifies more sustained soil moisture.

1022 Modern examples range from monsoonal domain rivers (e.g., Shukla et al., 2001; Singh
1023 et al., 2007), to semi-arid subtropics with mean annual precipitation of only 100-150 mm
1024 (Abdullatif, 1989). Common to these rivers is that they transmit effective discharge for a whole
1025 monsoon season that lasts 3-4 months, and thus experience high- as well as low-magnitude
1026 floods. Peak discharges, that cause high-magnitude flood events and deposition of UFR/HDR
1027 sedimentary structures and thick accretion sets, may last for only a few days (Abdullatif, 1989),
1028 and alternate with longer periods of lower discharge. As a result, a smaller proportion of the total
1029 sediment budget is transported during the peak flood, and thus the proportion of gradational and
1030 climbing UFR structures is lower. During the less intense phases of flood or at non-flood
1031 discharges, cross strata and ripples form, and in some cases bar forms and channels can migrate

more systematically. High-intensity peak floods may not occur during every flood season, but rather have a longer return period of up to several decades (Fielding et al., 2009).

7.4. *Lithosomes with in-channel vegetation*

Even more complex lateral facies changes and internal architecture characterize channel lithosomes where in-channel in-situ trees and vegetation-induced sedimentary structures are common (Fielding et al., 2009, Allen et al., 2013, 2014; Bashforth et al., 2014) (Fig. 14, 23d). They may show abrupt and complex lateral variations across only tens of meters, in lithology, sedimentary structures, and the thickness and orientation of accretion sets. A modern example comes from the Burdekin River in the sub-humid subtropics (Fielding et al., 2009), where annual mean precipitation is 1000-1200 mm. The Burdekin River transmits discharge only during the monsoon season, with recurrence frequency of high magnitude floods of 3-20 years (Fielding et al., 2009). During the sustained droughts, trees grow in channels due to the locally higher soil moisture, and interact with the flows during floods.

8. Monsoonal morphodynamics and catastrophic floods

This morphodynamic analysis identifies the inherent instability of monsoonal and subtropical rivers due to their frequent high-magnitude floods. This instability makes them especially prone to avulsions that generate catastrophic floods with potentially severe consequences (see discussion in 5.3; see also Fig. 25). Flood mitigation is commonly approached by barrage (levee) and dam construction, which regrettably does not consider the inherent morphodynamics of these flood-prone extremely dynamic rivers (Sinha, 2008; Syvitski and Brakenridge, 2013). There is no evidence for appreciable flood moderation in the Kosi and other rivers of north Bihar, for

example, after the construction of embankments and barrage (designed with a flood cushion; Sinha, 2008). On the contrary, both the recent most catastrophic flood events, the Kosi flood in 2008, and the Indus flood in 2010 (Fig. 25), were considerably amplified due to dam and barrage construction (Sinha, 2008; Syvitski and Brakenridge, 2013). The dams, embankments and barrages cause local backwater effects that result in reduced water and sediment conveyance capacity, and thus in increased local deposition and channel bed superelevation (Syvitski and Brakenridge, 2013). Furthermore, the locally reduced water and sediment conveyance capacity and the resultant local channel bed aggradation, or the combination of the local aggradation and associated cutbank erosion, push the rivers close to their avulsion threshold, and allow even moderate meteorological events to trigger catastrophic floods (Sinha, 2008; Syvitski and Brakenridge, 2013). It follows that reinforcing the existing engineering structures is not a sustainable strategy for avoiding future flood catastrophes, and major translocations of river flow will continue to occur regardless of whether or not flood warning is improved (see discussion in Sinha, 2008; Syvitski and Brakenridge, 2013). A process-based understanding of these rivers and the coupling between river form and processes can help develop a long-term solution to flood control (Sinha, 2008). Solutions that consider morphodynamics of these sediment-rich and avulsion-prone rivers may include planning for major water and sediment flow diversions for effective flood control (Syvitski and Brakenridge, 2013).

9. Climate linkage

Climate forcing of depositional systems is commonly inferred in the literature, but mostly in an implied way through oversimplified mean aridity-humidity or temperature changes. This is because specific controls are seldom invoked, and are difficult to identify in the paleo record.

Recognition of monsoonal and subtropical river morphodynamics and the resultant deposits provides one of the few direct climate proxies for estimating precipitation intensity and range (seasonality).

9.1. Monsoonal river dynamics as climate proxy

Deposition rates, avulsion frequency, fluvial fan aggradation and progradation, and incision in monsoonal and subtropical rivers are commonly considered as intimately linked to monsoon intensity and pattern changes (Pratt et al., 2002; Goodbred 2003; Thomas, 2003; Bookhagen et al., 2006; Sinha et al., 2007; Uba et al., 2007; Ray and Srivastava, 2010; Roy et al., 2012; Assine et al., 2014; Fontana et al., 2014). Yet, the specific link to monsoon pattern changes remains unclear, as monsoon intensification is often looked at as a simple increase in annual mean precipitation. This analysis signifies that the morphodynamics of monsoonal and subtropical rivers are linked to the monsoon precipitation intensity and range (peakedness) rather than total annual mean precipitation. Seasonal and inter-annual monsoon precipitation variability results in flood magnitude, duration and peakedness changes. This recognition provides a complementary climate proxy that directly reflects precipitation seasonality and the occurrence of extreme precipitation events. Furthermore, analysis of the proportions of high-magnitude flood indicators preserved in river channel lithosomes, allows for assessment of relative changes in monsoon rain intensity and duration over time. For example, recognition of amalgamated lithosomes with thick flood units (Fig. 24a) would suggest conditions where short-duration high-magnitude floods prevailed, and thus precipitation conditions where monsoon rain falls as relatively short-duration extreme precipitation events rather than distributed over a whole monsoon season (i.e. similar to some arid or semi-arid subtropics). In contrast, recognition of lithosomes with variable flood

units and lower flow regime deposits (Fig. 24b) would suggest conditions where precipitation season is longer and lasts for the whole monsoon season with variable strength.

A high seasonal precipitation variability or range inevitably causes seasonal or more sustained droughts. This recognition explains the seemingly contradictory future climate predictions (e.g., Seneviratne et al., 2012) or palaeoclimate interpretations (see discussion in McInerney and Wing, 2011; Plink-Björklund et al., 2014b) that invoke hydrological cycle intensification, as well as increased droughts. The in-channel desiccation indicators allow an estimation of duration of channel bed exposure and thus provide a measure for the intermittency of droughts. Moreover, due to dominant deposition during floods, seasonal river systems are likely to be highly sensitive to climate forcing, as they experience fewer buffering or hysteresis effects due to a fast downstream propagation of sediment (see also Goodbred, 2003).

As demonstrated in this review, the morphodynamics and the resulting sedimentary record of the monsoon-affected seasonal rivers is distinct from rivers of perennial precipitation climate zones (Table 2). However, polar-domain rivers may also exhibit distinct seasonality due to winter freezing and intense summer thawing (e.g., Flaig et al., 2015). This similarity and the potential differences need further assessment, but it is likely that some polar river deposits may contain abundant evidence for floods, and thus may display partially similar facies characteristics. However, the permafrost, freeze-thaw and other cryogenic processes, slower weathering and pedogenic processes, and differences in vegetation and insect habitats are likely to produce distinct differences.

9.2. *Monsoonal palaeoclimates*

In modern climate conditions, the monsoonal and subtropical zones occur outside the tropical rainforest zone and within ca 35° of latitude. In contrast, some of the early Eocene river systems that are consistent with the evidence presented herein, occur between ca 35-49° palaeolatitude (e.g., Schmitz and Pujalte, 2007; Foreman et al., 2012; Plink-Björklund et al., 2014b). A possible mechanism for such an increase in precipitation seasonality in the midlatitudes is increased poleward tropical heat and moisture flux (e.g., Emanuel, 2008) that is largely defined by the Hadley cell. Recent work indicates that the Hadley cell or tropical belt is currently expanding, with consequences such as poleward shift of the jet streams, poleward shift of midlatitude storm tracks, poleward shift of the zonal mean atmospheric circulation, and poleward expansion of the subtropical zone (Frierson et al., 2007; Lu et al., 2007; Seidel et al., 2008). Thus, if applied to Greenhouse-world river systems, the range of facies presented herein could indicate strong monsoonal precipitation and climate conditions controlled by seasonal monsoon precipitation, rather than palaeolatitude.

10. Conclusions

This review demonstrates that morphodynamics of rivers that receive most of their surface water from the seasonal, intense monsoon precipitation are controlled by dominant sediment transport and deposition during high-magnitude floods. Other properties include sharpness of the flood discharge hydrograph, together with high peak discharge, rapid lowering of water level during the falling stage of floods, lack of efficient perennial base flow, seasonal partial or complete channel bed exposure, high sediment concentrations and high suspended sediment load during floods, water saturation of floodplain sediments during floods, variable water table, high evapotranspiration, and water loss through seepage and floodplain flooding. The characteristics

or recognition criteria for such rivers include (1) the abundance of Froude transcritical and supercritical flow sedimentary structures; (2) abundance of high deposition-rate sedimentary structures; (3) in-channel mud layers; (4) thick mud-clast conglomerates; (5) abundant soft-sediment deformation; (6) in-channel vegetation and vegetation-induced sedimentary structures; (7) in-channel bioturbation; (8) abundance of flood units; (9) poorly developed macroforms or barforms; (10) dominance of avulsions over channel migration; and (11) abundance of fluvial fans.

Such rivers giving rise to these sedimentary characteristics occur in the monsoonal domain as well as in the subtropics, and are linked to the large precipitation range and inter-annual variability, rather than latitude or mean annual precipitation. They are inherently prone to avulsions and catastrophic flooding, and successful flood control needs to consider their inherent instability. When used for palaeoclimate assessment, they indicate a monsoon precipitation driver rather than palaeolatitude.

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1562 **Figure Captions:**

1563 **Fig. 1.** Monsoonal domain (green colors encircled by black lines) is characterized by (a) a large
1564 annual precipitation range (summer minus winter precipitation) as compared to annual mean
1565 (monsoon precipitation index), and by (b) monsoon westerly wind index (from Wang and Ding,
1566 2008). The bordering subtropical areas (green shades) also receive monsoon precipitation, but
1567 experience larger inter-annual variability. (c) The effect of this seasonal precipitation variability
1568 on river discharge is demonstrated by the robust correlation between the precipitation seasonality
1569 index and the discharge seasonality index. Both are here calculated as average of wettest month

1570 minus driest month divided by annual mean, and data for each river is averaged over ca 4-80
 1571 years (data from Leier et al., 2005). In contrast, there is a low correlation between the annual
 1572 means (d) or the annual monthly maxima (e). Examples of river discharge hydrographs
 1573 (averaged over 6-63 years) for (f) rivers that receive large amount of surface water supply from
 1574 monsoon rain and thus have a large annual discharge range as compared to annual mean
 1575 (discharge seasonality index >2); compared to (g) rivers that receive surface water from temporal
 1576 perennial precipitation and rainforest zone, and thus have a small annual discharge range
 1577 compared to annual average (discharge seasonality index <2). (h) Characteristically, the positive
 1578 deviation is proportionally larger in rivers with discharge seasonality index >2 . (i) Comparison of
 1579 discharge seasonality index, positive deviation, standard deviation and average flood discharge
 1580 (average discharge of four consecutive month of highest discharge divided by annual mean.
 1581 Discharge data in h-i from <http://www.sage.wisc.edu/riverdata>, except for Kosi and Naryani
 1582 (<http://icimod.org>), Rhein (van Bokhoven, 2006), Kassala (Abdullatif, 1989). (j) Comparison of
 1583 inter-annual variability between a monsoonal river (Ganges) and a subtropical river (Burdekin),
 1584 over 8 years (data from <http://www.sage.wisc.edu/riverdata>).
 1585 **Fig. 2.** Froude transcritical to supercritical or upper flow regime (UFR) planar laminations are
 1586 the most common sedimentary structures in seasonally variable discharge rivers. (a) Planar-
 1587 laminated, erosionally bounded (black arrows) channel fill sandstones, early Eocene Wasatch
 1588 Fm., Uinta Basin, USA. Person for scale. Palaeocurrent direction is obliquely to the right. (b)
 1589 Planar-laminated sandstones, Permian Organ Rock Fm., Utah, USA (from Cain and Mountney,
 1590 2009). (c) Bedding plane view of planar-laminated sandstone with well-developed parting
 1591 lineation, Pennsylvanian-Permian Maritimes Basin, Cape Breton Island, Nova Scotia, Canada
 1592 (from Allen et al., 2014). (d) Planar laminated sand of 1981 Laingsburg flood, South Africa,

1593 compared to (e) planar-laminated sandstone of Permian Beaufort Fm., Karoo Basin, South Africa
1594 (from Stear, 1985).

1595 **Fig. 3.** Convex-up low-angle bedforms and scour and fill structures that preserve antidune and
1596 chute and pool formation are also common in seasonally variable discharge rivers. (a) Convex-up
1597 low-angle bedform that preserves sub-vertical aggradation of antidunes, Mississippian-Permian
1598 Cumberland Basin, Canada (from Allen et al., 2013). (b) Bedding plane view of convex-up low-
1599 angle laminated sandstone with well-developed parting lineation, Triassic Wolfville Fm., Nova
1600 Scotia, Canada (from Leleu et al., 2010). (c) A photo and (d) line drawing of planar laminated
1601 and convex-up low-angle laminated sandstone, Eocene Wasatch Fm., Uinta Basin, USA. The
1602 bedform has a sub-vertical angle of climb (arrows). (e) Planar laminations commonly pass
1603 laterally or vertically into convex-up low-angle laminations or scour and fill structures, Eocene
1604 Wasatch Fm., Uinta Basin, USA. Palaeocurrent direction is obliquely into the cliff. (f) Scour and
1605 fill structures (bases marked by black arrows) and convex-up structures (gray arrows; compare to
1606 Fig. 5), Eocene Wasatch Fm., Uinta Basin, USA. Palaeocurrent direction is parallel with the cliff
1607 towards right. (g) Scour and fill structures with characteristic upward flattening of lamina,
1608 Eocene Wasatch Fm., Uinta Basin, USA. Palaeocurrent direction is obliquely to the right. (h)
1609 Meter-scale scour and fill structures, Permian Lyons Fm., Morrison, USA. Palaeocurrent
1610 direction is into the cliff. (i) Decimeter-scale scour and fill and convex-up structures, Ganga
1611 River, Himalayan foreland (from Singh and Bhardwaj, 1991).

1612 **Fig. 4.** Froude transcritical to supercritical or upper flow regime structures commonly pass
1613 laterally and vertically from one form to another, creating complex laminations. (a) Photo and (b)
1614 line drawing of complex scour and fill and convex-up structures, Eocene Wasatch Fm., Uinta
1615 Basin, USA. Palaeocurrent direction is obliquely to the left. (c) Large scour and fill structures

1616 (arrows). Note how there is some cross stratification at the edge of some of these structures (gray
1617 arrow) due to the migration of the bedform, Eocene Wasatch Fm., Uinta Basin, USA.
1618 Palaeocurrent direction is obliquely to the right. (d) Humpback or sigmoidal cross strata that are
1619 characterized by aggradational topsets and bottomsets pass vertically into planar laminations,
1620 Eocene Wasatch Fm., Uinta Basin, USA. Palaeocurrent direction is parallel with the cliff
1621 towards right. (e) Sigmoidal cross strata with aggradational topsets become gradually more
1622 aggradational from left to right, Devonian Chloritic Sandstone, Ireland (from Kelly and Olsen,
1623 1993). Palaeocurrent direction is parallel with the cliff towards right. (f) Sigmoidal cross strata
1624 showing preservation of the complete bedform with extensive bottomset, Permian Bainmedart
1625 Coal Measures, Prince Charles Mountains, Antarctica (from Fielding, 2006). Palaeocurrent
1626 direction is parallel with the cliff towards left.

1627 **Fig. 5.** Experimentally produced antidune and chute and pool laminae show features similar to
1628 field examples in Figures 2-4. (a) Flow parallel and (b) flow perpendicular view of antidune
1629 laminae displaying a combination of convex-up, scour and fill and planar lamina (from
1630 Alexander et al., 2001). (c) Flow parallel and perpendicular view of experimental laminae
1631 produced under changing flow conditions from chute and pool to antidunes and to plane beds,
1632 displaying an upward transition from scour and fill features to convex-up lamina and to planar
1633 lamina (from Alexander et al., 2001).

1634 **Fig. 6.** Comparison of upper flow regime (UFR) and high deposition rate (HDR) sedimentary
1635 structure abundance in modern (1, 4, 8) and ancient (2-3, 5-7) variable-discharge river deposits,
1636 to dominantly cross stratified deposits (9) characteristic for rivers of perennial discharge zone.
1637 Note that some variable-discharge rivers successions are dominated by UFR structures with over
1638 70%, whereas others only contain around 50%. (a) Proportion of sedimentary structure groups in

1639 each river deposit. (b) Comparison by each sedimentary structure group. (c) Pie charts of facies
1640 distribution by individual river systems.

1641 **Fig. 7.** High deposition rate structures are also extremely common in seasonal river deposits.
1642 Climbing ripples commonly occur together with upper flow regime structures. (a) Climbing
1643 ripples with sub- to supercritical angle of climb, Triassic Moenkopi Fm., USA (from Banham
1644 and Mountney, 2014). Climbing ripples, in (b) Eocene Wasatch Fm., Uinta Basin, USA, and (c)
1645 Rio Colorado, Altiplano, South America (from Donselaar et al., 2013). (d) Climbing dune cross
1646 strata, Eocene Wasatch Fm., Uinta Basin, USA. Palaeocurrent direction is parallel with the cliff
1647 towards left. (e) In gradational planar laminated sandstone, each lamina is multiple millimeters to
1648 centimeters thick, and has gradational or graded boundaries, Eocene Wasatch Fm., Uinta Basin,
1649 USA. (f) Gradational planar laminae that consist of graded gravel and sand particles, Triassic
1650 Moenkopi Fm., USA (from Banham and Mountney, 2014).

1651 **Fig. 8.** Examples of channel sandstones dominated by climbing-ripple cross lamination. (a)
1652 Climbing-ripple cross laminated channel sandstone, bounded by erosion surfaces (thick lines)
1653 displays multiple internal erosion surfaces (thin lines), Eocene Wasatch Fm., Uinta basin, USA.
1654 Palaeocurrent direction is obliquely to the left (b) Climbing-ripple cross laminated channel
1655 sandstone with mud interlayers, Rio Colorado, Altiplano, South America (from Donselaar et al.,
1656 2013). (c) A measured section and photo of dominantly climbing-ripple cross laminated channel
1657 fill sandstone, Jurassic Moenave Fm., USA (from Olsen, 1989). Palaeocurrent direction is
1658 obliquely into the cliff.

1659 **Fig. 9.** In-channel mud layers commonly occur in monsoonal and subtropical rivers. (a)
1660 Desiccated mud veneer on the edge of a ripple-marked sand bar, as observed a few days after the
1661 Laingsburg flood in 1981, Karoo South Africa (Stear, 1985). (b) Similar mud drape on the edge

1662 of a ripple-marked bar in the Permian Lower Beaufort Fm., Karoo, South Africa (Stear, 1985). In
1663 both cases, channel is to the right and paleo-flow into the photo. Decimeter-thick pedogenically
1664 modified mud layers in lateral accretion sets (c) Eocene Green River Formation, Uinta Basin,
1665 USA; and (d) Eocene-Oligocene Potoco Fm., Andes (Hampton and Horton, 2007). In both cases
1666 paleo-flow is into the photo. (e) Desiccated and broken-up mud drapes at the bottom of
1667 seasonally dry Rio Colorado, Altiplano, South America (Donselaar et al., 2013). Arrow shows
1668 current direction. Desiccated curled (f) and broken-up (g) mud drapes in deposits of Wadi El
1669 Arish, Egypt (Sneh, 1983).

1670 **Fig. 10.** In-channel mud deposits are characteristically interbedded with sandy channel fill
1671 deposits. (a) Longitudinal section of a braid bar complex of the Ganga River showing three bar
1672 depositional events (A, B, and C). These events are separated by erosional contacts, mottled,
1673 deformed tens of meters wide mud horizons, and are produced during high magnitude floods.
1674 Water overtops during the flow of peak monsoon discharge (Singh et al., 2007). (b) A few
1675 meters wide mud horizons at the top of sandy flood deposits in a braid bar of the Ganga River
1676 (Shukla et al., 2001). (c) Mud layers in lateral accretion sets thicken downward towards the
1677 lower parts of foresets and in bottomsets of the large-scale cross strata, Green River Fm., Uinta
1678 Basin, USA. Palaeocurrent direction is obliquely into the cliff. (d) Tens of meters wide mud
1679 horizons at the top of flow-perpendicular flood deposits in River Gash, Sudan. The sedimentary
1680 structures in the flood units indicate upward decrease in flow velocity (Abdullatif, 1989).

1681 **Fig. 11.** Thick soft-sediment or intra-basinal clast conglomerates occur in some seasonal river
1682 deposits. (a) Several meters thick mud and caliche clast conglomerate in the Oligocene-Miocene
1683 Petaca Fm., Andean Foreland (from Hulka and Heubek, 2010). A 4 meters thick mud clast
1684 conglomerate that is disorganized within the lower 2 meters (b) and stratified within the upper 2

1685 m (c), Eocene Wasatch Fm., Uinta Basin, USA. Palaeocurrent direction is into the photo. (d)
1686 Mud clast conglomerate at a channel base that is disorganized in the lower part and cross
1687 stratified in the upper part, Eocene Wasatch Fm., Uinta Basin, USA. Palaeocurrent direction is
1688 obliquely to the left. (e) Overview and (f) detail of a crudely stratified lens-shaped mud-clast
1689 conglomerate, Eocene Green River Fm., Uinta Basin, USA. Palaeocurrent direction is into the
1690 photo. (g) A meter thick, laterally extensive conglomerate at a channel base, Eocene Wasatch
1691 Fm., Uinta Basin, USA. Palaeocurrent direction is obliquely to the left.

1692 **Fig. 12.** Bank erosion and failure during UFR flow conditions and falling stages of floods are
1693 especially significant for formation of soft-sediment clast conglomerates in seasonal rivers. (a) A
1694 soft-sediment clast conglomerate (Fci) grades into planar laminated sandstone (Fh and Fm) in
1695 deposit from one flood event. The pebble-sized clasts consists of calcrete, silt and sand, Permian
1696 Organ Rock Fm. (from Cain and Mountney, 2009). (b) Shear failure caused by bank
1697 undercutting, the Brahmaputra River (from Coleman, 1969). (c) Bowl-shaped failure caused by
1698 porewater overpressure and subaqueous flow, the Brahmaputra River (from Coleman, 1969). (d)
1699 Bench cut into a channel bank mantled with fallen blocks and carbonate clasts, Rupen River,
1700 India (Gibling and Tandon, 1997). (e) Erosional bank margins and carbonate clast conglomerate
1701 at the channel base (c), Rupen River, India (Gibling and Tandon, 1997). (f) Undercut channel
1702 bank mantled with an intra-formational conglomerate, the Big Thomson River after 2013 flood
1703 (photo by Rich Abrahamson from The Coloradoan, archive.coloradoan.com). (g) Dendritic
1704 tributive pattern formed after Colorado flood in 2013, by channel bank flowage due to porewater
1705 overpressure (from <http://www.firstcoastnews.com>). River is to left in this image.

1706 **Fig. 13.** Soft-sediment deformation structures are common in some seasonal rivers. Examples of
1707 convolute bedding overlying (a) a thin zone of climbing ripples and planar laminations, and (b)

1708 planar laminae with extreme distortion above the convolute laminae, the Brahmaputra River
 1709 (from Coleman, 1969). (c) Convolute bedding above planar laminae, Bijou Creek 1965 flood
 1710 deposit (from McKee et al., 1967). (d) Convolute bedding (Fc) in the Ganga River deposits
 1711 (Singh and Bhardwaj, 1991). (e) Rare convolute bedding above planar laminations in Eocene
 1712 Wasatch Fm., Uinta Basin, USA. (f) Overturned cross strata overlain by sigmoidal or humpback
 1713 cross strata (Cxd), Jurassic Kayenta Fm., USA (from North and Taylor, 1996). (g) Overturned
 1714 cross strata in the Jurassic Kayenta Fm., USA (from Bromley, 1991).

1715 **Fig. 14.** In-channel vegetation and vegetation-induced sedimentary structures. (a) Photo and (b)
 1716 line drawing of two channel-fills (CB1 and CB2) with a variety of in situ trees and vegetation-
 1717 induced sedimentary structures, Pennsylvanian-Permian Sydney Mines Fm., Nova Scotia (from
 1718 Allen et al., 2014). The approximate base of the lower channel body (CB 1). The base of CB 1 is
 1719 at ground level is heterogeneous with abundant mudstone (grey) partings which overlie hollows
 1720 that were infilled with sediment following decay of upright tree trunks. Note the downturned
 1721 beds associated with many of these mud-filled hollows. CB 1 becomes coarser grained
 1722 upsection, although lateral transitions into mudstones are common. The chaotic nature of
 1723 bedding is due to plant-sediment interactions. The mud layer at the top of CB 1 has been
 1724 extensively pedogenically modified. CB 2 exhibits a similar geometry and architecture to CB 1.
 1725 Palaeocurrent is approximately out of the photograph. (c) Centroclinal cross strata, (d) upturned
 1726 bedding around a former tree, and (e) three broken *Calamites*, Pennsylvanian Sydney Mines Fm.,
 1727 Nova Scotia (from Allen et al., 2014). (f) Upright cordaitalean (*Dadoxylon*) tree, and (g)
 1728 complete *Cordaite*s leaves in Pennsylvanian Tynemouth Creek Fm., Canada (from Bashforth et
 1729 al., 2014). (h) Small in situ tree fossil encased in channel sandstone, showing tilting in down-
 1730 paleoflow direction (to right) and upstream-trailing roots (to left) in Pennsylvanian Maritimes

1731 basin (from Fielding et al., 2009). (i) Linear *M. argentea* grove forming the nucleus of a linear
1732 bar, and (j) plant litter against lower foreset of dune near the low-stage channel in Burdekin
1733 River, Australia (from Fielding et al., 2009).

1734 **Fig. 15.** In-channel bioturbation indicates temporary channel bed exposure and fluctuating water
1735 table. Deep insect nests (a and b) and *Camrogyma* traces (c), Eocene Wasatch Fm., Uinta Basin,
1736 USA. (d) Vertical traces in Miocene Tariquia Fm., Andean foreland (from Uba et al., 2005). (e)
1737 *Ancorichnus* traces on an accretion surface, Eocene Wasatch Fm., Uinta Basin, USA. Close up
1738 (f) and overview (g) of in-channel pedogenic modification on an accretion surface, Eocene
1739 Wasatch Fm., Uinta Basin, USA. Photo (h) and line drawing (i) of in-channel bioturbation on
1740 heterolithic accretion surfaces, Paleogene Ebro basin, Spain (from Saez et al., 2007). la – lateral
1741 accretion.

1742 **Fig. 16.** Deposition during flood events generates distinct flood deposits. In conglomeratic
1743 channel fills, the single flood deposits tend to consist of erosionally based conglomerate to
1744 sandstone units. Examples from (a) Devonian Snehvide Fm., Greenland (from Kelly and Olsen,
1745 1993); (b) Miocene-Pliocene Guandacay Fm., Andean Foreland (from Hulka and Heubek, 2010);
1746 (c) photo and (d) line drawing of Miocene-Pliocene Guandacay Fm., Andean Foreland (from
1747 Uba et al., 2005); (e) monsoonal Tista River, India (from Chakraborty and Gosh, 2010). (f)
1748 Example of sandy planar laminated flood deposits bounded by erosion surfaces (Ss), with a basal
1749 conglomerate (Ch), monsoonal Tista River, India (from Chakraborty and Gosh, 2010). (g) Flood
1750 deposit with scour and fill to planar laminated sandstones, based by a mud-clast conglomerate,
1751 Eocene Wasatch Fm., Uinta Basin, USA. Palaeocurrent direction is obliquely to the left.

1752 **Fig. 17.** Examples of flood deposits. (a) Stacked planar-laminated flood deposits, bounded by
1753 sharp surfaces and generally lacking mud drapes. The lowermost unit is based by a mud-clast

1754 conglomerate, Eocene Wasatch Fm., Uinta Basin, USA. Palaeocurrent direction is obliquely into
1755 the cliff. (b) Two stacked flood deposits bounded by erosion surfaces. The lower deposit
1756 transitions from convex-up low-angle laminae to climbing ripples, and the second from scour
1757 and fill structures to climbing ripples, Eocene Wasatch Fm., Uinta Basin, USA. Two examples of
1758 flood units from Kosi river with (c) planar laminated to climbing-ripple laminated and (d) scour
1759 and fill to planar to climbing ripple laminated sandstones that is partially truncated by next event
1760 (from Singh et al., 1993). Paleo-flow direction is towards left in c and obliquely into the diagram
1761 in d. (e) Scour and fill to planar laminated flood deposit, Eocene Wasatch Fm., Uinta Basin,
1762 USA. (f) Deposit of 1965 flood in Bijou Creek, CO (from McKee, 1967), and (g) of Palmer
1763 River, Lake Eyre basin, Australia (Croke et al., 1998), both showing a transition from planar to
1764 climbing-ripple laminated sandstone. (h) Repeated transitions from planar (Sh) to climbing ripple
1765 laminations (Sr) in Permian Beaufort Fm., Karoo, South Africa (Stear, 1985).

1766 **Fig. 18.** Examples of flood deposits. (a) Stacked, laterally continuous planar-laminated flood
1767 deposits of Tista River, India (from Chakraborty and Gosh, 2010). (b) Flood deposits of Wadi El
1768 Arish, Egypt (from Sneh, 1983). (c) Deposits of 1965 flood in Bijou Creek, CO. Note that there
1769 are two units deposited from a single flood due to the occurrence of two flood waves (from
1770 McKee, 1967).

1771 **Fig. 19.** Examples of flood units and macroforms. (a) Thickness of individual flood deposits in
1772 lateral accretion sets in the lower part of the channel fill is on decimeter scale, in contrast to the
1773 overlying flood units that are 1-2 m thick. Each flood deposit is draped by mud, Eocene Green
1774 River Fm., Uinta Basin, USA. Palaeocurrent obliquely from left to right. (b) Extremely thick,
1775 erosionally bounded flood deposits that consist of stacked scour and fill, convex-up low-angle
1776 and planar laminated sandstones are laterally continuous for more than 100 m, Eocene Wasatch

1777 Fm., Uinta Basin, USA. Palaeocurrent from left to right. (c) A few m thick flood units that are
 1778 laterally continuous for more than 100 m, Eocene Wasatch Fm., Uinta Basin, USA.

1779 Palaeocurrent obliquely into the cliff. (d) Thick flood units that display finer-grained (very fine
 1780 sand to silt) upper parts that are climbing-ripple laminated, Cretaceous Piceance basin, USA.
 1781 Palaeocurrent obliquely into the cliff.

1782 **Fig. 20.** Examples of macroforms. (a) Low angle accretion sets (1-9) in Ganga River deposits,
 1783 covered by mottled silty sand (A). Thick arrows show inclinations of the accretion sets and thin
 1784 arrows palaeocurrent directions from cross strata (from Shukla et al., 2001). (b) Dry Burdekin
 1785 river bed at low stage, covered by dunes, plane beds, large gravelly antidunes (dark traces in
 1786 lower center), and linear groves of *Melaleuca argentea* (from Fielding et al., 2009). (c) Flat bed
 1787 morphology of the dry Dikala River, Kobo basin, Ethiopia (from Billi, 2007). (d) Development
 1788 bars similar to transverse bars of linguoid type (center) in River Gash, Sudan as flood stage falls.
 1789 However, these bars are partially erosional features, as seen by the erosional edges. (e) The same
 1790 location, but at lower flood stage. Flow is towards the observer (north), and channel width is
 1791 about 100 m (from Abdullatif, 1989).

1792 **Fig. 21.** Avulsions that are common in rivers with seasonally variable discharge generate highly
 1793 amalgamated tabular-looking channel zones (a), where channel return frequency is high, Eocene
 1794 Wasatch Fm., Uinta Basin, USA. Black lines indicate some of the channel bases. Palaeocurrent
 1795 is obliquely to the left into the cliff. In areas with lower channel return frequency, flood units
 1796 characteristically contain floodplain deposits with numerous tabular splay sands, overlain by
 1797 channel fills (yellow arrows), Eocene Green River Fm., Uinta Basin, USA (b); Cretaceous
 1798 Williams Fork Fm., Piceance Basin, USA (c). Palaeocurrent into the cliff both in b and c.

Fig. 22. Examples of large fluvial fans (fluvial megafans). (a) Google Earth image of the Kosi megafan, built by frequent Kosi river avulsions. The avulsions are demonstrated by mapping of paleo-channel positions (b), as depicted in historical maps on LANDSAT image. Red shaded area shows August 2008 avulsion (from Chakraborty et al., 2010; see also Fig. 25). (c) Mosaic of Landsat TM4/5 images in 1990, band 7 and (d) interpretation of the Brazilian Pantanal and the megafans that feed the Paraguay River. (1) Paraguay, (2) Cuiabá, (3) São Lourenço, (4) Taquari, (5) Taboco, (6) Aquidauana and (7) Nabileque (from Zani et al., 2012).

Fig. 23. There is considerable variability in the number of the facies and architectural characteristics present in the dataset in Table 1. (a) Graph illustrating the number of datasets where each characteristic has been recognized. Some features are present in almost 100% of the datasets, whereas others only in 20%. (b) Histogram showing that most commonly the datasets in Table 1 report the presence of 5-7 characteristics or recognition criteria, 89% of the datasets report 5 or more.

Fig. 24. Examples of channel lithosomes of rivers strongly affected by highly seasonal monsoon precipitation, as a function of discharge flashiness (proportion of high-magnitude flood), sediment supply, and the presence of in-channel vegetation. (a) Amalgamated lithosomes with thick flood units reveal the flashiest systems with high peak discharge, very sharp flood hydrographs with especially rapid rise and decline, and deposition during high-magnitude floods only. (b) Lithosomes with variable flood units also signify that deposition only occurred during high-magnitude floods, but with less sharp flood hydrographs and preservation of some waning-stage deposits. (c) Lithosomes with flood units and lower flow regime deposits indicate that a smaller proportion of the total sediment budget was transported during the peak flood, and cross strata and ripples formed, and barforms and channels were able to migrate more systematically

during less intense phases of flood or at non-flood discharges. (d) Lithosomes with in-channel vegetation are internally more complex due to the interaction of vegetation with flow (modified from Fielding et al., 2009). UFR – upper flow regime; HDR – high deposition rate.

Fig. 25. Images of Kosi River flood in 2008 and Indus River flood in 2010. Both these catastrophic floods were considerably amplified by barrage construction with poor understanding of monsoonal river morphodynamics. Moderate Resolution Imaging Spectroradiometer (MODIS) images of NASA's Terra satellite of Kosi River from (a) pre-flood conditions in August 8, (b) during the flood on August 24, and (c) September 2, and (d) Dartmouth Flood Observatory map from September 2 (from <http://earthobservatory.nasa.gov>). Moderate Resolution Imaging Spectroradiometer (MODIS) images of NASA's Terra satellite of Indus River from (e) pre-flood conditions on July 16, (f) during the flood on September 20, (g) pre-flood conditions on August 18, 2009, and (h) during the flood August 17, 2010 (from NASA 2010).

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(a) Antidune lamina, flow parallel view

(b) Antidune lamina, flow perpendicular view

(c) Chute and pool lamina

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River deposits with a high proportion of upper flow regime and high deposition rate structures:

1 Kobo Basin, Ethiopia (Billi, 2007). 2 Eocene Wasatch Fm., Uinta Basin, US (Plink-Bjorklund and Birgenheier, 2013). 3 Jurassic Kayenta Fm., US (North and Taylor, 1996). 4 Karoo, South Africa (Stear, 1985). 5 Cretaceous Williams Fork Fm. (McDowell and Plink-Bjorklund, 2013). 6 Eocene Wasatch and Green River Fm., Uinta Basin, US (Plink-Bjorklund and Birgenheier, 2013). 7 Eocene Wasatch and Green River Fm., Uinta Basin, US (Plink-Bjorklund and Birgenheier, 2013). 8 River Gash, Sudan (Abdullatif 1989).

River deposits with dominant cross strata:

9 North Horn Fm., Uinta Basin, US (Plink-Bjorklund and Birgenheier 2013).

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Legend:

	mud curly		convoluted
	mud drapes		climbing ripples
	mud debris		ripples
	massive		planar cross stratified
	pebbles		trough cross stratified
			undulated
			horizontally stratified

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August 8, 2008

September 2, 2008

August 24, 2008

Dartmouth Flood Observatory Map September 2, 2008

Figure 25 continued
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Table 1. Summary of datasets included in this review.

x present; - absent; D dominant; A abundant; M minor; LA-lateral accretion, DA-downstream accretion; VA vertical accretion; D/A – dominant in proximal and abundant in distal part; M/A – minor in proximal and abundant in distal part; FM – fluvial megafan; TF – terminal fan; TDS – terminal distributary system; FF – fluvial fan; FDS – fluvial distributary system; DFS –distributive fluvial system.
* - ancient systems with independent evidence for palaeoclimatic conditions.

River system	Sedimentary structures			Lithology		Biogenic features		Architecture			References	
	Upper flow regime	High deposition rate	Soft-sediment deformation	Mud-clast conglomerate	In-channel mud	In-channel traces, desiccation	In-channel vegetation	Flood units	Macro-forms	Avulsions	Fluvial fan	
<u>Modern and Quaternary examples:</u>												
Monsoonal												
Tista River	D/A	A	M	M	x		-	x	x	D	FM	Chakraborty and Ghosh, 2010
Kosi River	A	A	A	A	x	x	x	x	x	D	FM	Singh et al., 1993; Chakraborty et al., 2010
Baghmati River	A	A			x		-		x	D	-	Jain and Singha, 2003; 2004
Ganga river	A	A	M/A	M	x	x	x	x	x	D	FM	Singh and Bhardwaj, 1991; Shukla et al., 2001; Singh et al., 2007
Plio-Pleistocene Siwalik	A	A		A	x				VA	D	FM	Thomas et al., 2002; Kumar et al., 2003
Mio-Pleistocene Siwalik	A	A	x	A	x	x			DA	D	FM	Kumar et al., 1993; Willis, 1993; Nakayama and Ulak, 1999
Quaternary Lake Eyre Basin	A				x							Croke et al., 1998
Sub-humid												
Burdekin River	A	A	x	x	x	x	x		-		-	Fielding et al 2009
Semi-arid												
River Gash	A				x			x	DA	D	TF	Abdullatif, 1989
Kobo Basin	D				M		-	x	LA			Billi, 2007
Karoo	D	A			x		-	x	-	D	TDS	Stear, 1985
Late Quaternary, Lake Eyre Basin	D	A			x	x	x		VA			Croke et al., 1998
Arid												
Rio Colorado	-	D			A	x	-	x	LA	D	FF	Donselaar et al., 2013
Wadi El Arish	A	A	M	x	x		-	x				Sneh, 1983
Late Quaternary Sinai	A	A	M	x	x		-	x				Sneh, 1983
Lake Eyre Basin	A	A	-	M				x	DA			Williams, 1971
Lake Eyre Basin	x	A			A	x	x		LA		FDS	Fisher et al., 2008

Ancient examples:**Monsoonal**

Miocene Siwalik*	A	A		x	-	-		VA DA LA	D	FM	Zaleha, 1997
Paleocene-Eocene Lake Eyre Basin*	A			x						FF	Croke et al., 1998; Greenwood et al., 2003
Cretaceous Williams Fork*	A	x		x	x	-	x	DA LA	D	FM	Plink-Bjorklund et al., 2014

Sub-humid

Eocene Green River Fm.*	A	A	x	x	x	x	x	DA LA	D	FM	Plink-Bjorklund and Birgenheier, 2013; Plink-Bjorklund et al., 2014
Pennsylvanian Maritimes basin*	A	A	x	x	x	x	x	-		-	Fielding et al., 2009; Allen et al., 2011
Lower Pennsylvanian Cumberland basin*	A		x		x	x	x		D	FM	Bashforth et al., 2014
Pennsylvanian Permian, Nova Scotia*	D	A	x		x	x	x	VA DA			Allen et al., 2014
Mississippian- Permian, Cumberland B.*	D	A			x	x	x	VA M: DA LA			Allen et al., 2013
Lower Pliocene Chaco foreland*	A	A	x	M	x		x	DA M: LA	D	FM	Uba et al., 2005

Semi-arid

Pliocene Caspian basin	D	A	A	A	x	x	-	x		D	TFS	Hinds et al., 2004
Oligocene- Miocene Chaco foreland*	A	A	x	A		x	-	x	-		-	Uba et al., 2005; Hulka and Heubek, 2010
Miocene Tariquia Fm*	A	A	x	A	x	x	-	x		D	FM	Uba et al., 2005
Miocene Ebro Basin, Spain*	D	A		x	x	x	-	x	DA LA	D	FDS	Nichols and Hirst, 1998; Nichols and Fisher, 2007
Miocene Kong Pa, Vietnam*	A	A		x	x					D	FF	Nielsen et al., 2007
Eocene-Oligocene Potoco Fm.*	A	A	A	x	A		-	x	-	D	FF	Hampton and Horton, 2007
Eocene Wasatch Fm.*	D	D	-	A	M	A	-	x	DA	D	FM	Plink-Bjorklund and Birgenheier, 2013; Plink-Bjorklund et al., 2014
Eocene Green River Fm.*	D	A	-	x	x	x	-	x	DA LA	D	FM	Plink-Bjorklund and Birgenheier, 2013; Plink-Bjorklund et al., 2014
Paleogene Ebro basin*	D	A			x	x	-		LA		FF	Saez et al., 2007
Cretaceous	A	A		x	x		-	x	-	D	TF	Rhee and Cough, 1993

Pxonghae Fm Jurassic Kayenta Fm.*	D	A	x	A	M	-	-	R: DA LA VA LA DA LA	D	TF	Bromley, 1991, North and Taylor, 1996	
Jurassic Moenave Fm.*	D	A		N	x	-	-		D		Olsen, 1989	
Triassic Chinle Fm.*	D	A	x	x	x			x	D	FF	Deluca and Eriksson, 1989; Trendell et al., 2013	
Triassic Moenkopi Fm.	A	A		A	x			x			Banham and Mountney, 2014	
Triassic, Marocco*	D	A	-	M	x	x	-	x			Mader and Redfern, 2011	
Triassic, Nova Scotia*	A	A		x	x		-		DA LA	DFS	Leleu and Hartley, 2010; Leleu et al., 2010	
Permian-Jurassic South Africa	D	x	x	x	x			x		FF	Turner, 1986	
Permian Organ Rock Fm.*	A	A	x	x	x			x	D	TFF	Cain and Mountney, 2009	
Permian Beaufort Fm., South Africa*	D	A		A	x		-	x	VA LA	D	Stear, 1985, Wilson et al., 2014	
Permian Motur Fm., India*	x				x				LA	D	Ray and Chakraborty, 2002	
Permian Uralian foreland basin*	A				x				LA	D	TF	Newell et al., 1999
Pennsylvanian Permian, Nova Scotia	D	A*			x	x			VA			Allen et al., 2014
Mississippian- Permian, Cumberland B.	A				A	x	-		M: LA			Allen et al., 2013
Devonian Trentishoe Fm.	D	A	A	M	x			x	VA LA DA			Tunbridge, 1981
Devonian Gun Point Fm.	A	A					-			TF		Sadler and Kelly, 1993
Devonian Greenland, Spitsbergen, England, Ireland*	A	A					-			TF		Kelly and Olsen, 1993

Table 2[Click here to download Table: Table 2.docx](#)**Table 2.** Comparison of monsoonal and subtropical rivers with highly seasonal discharge to perennial rivers.

	Monsoonal and subtropical	Perennial	Forcing factors
Sedimentary structures	Dominant or abundant Froude supercritical and transcritical flow (UFR), and abundant high deposition rate (HDR) structures	Dominated by Froude subcritical flow structures such as cross strata	The high flow velocities, and the characteristically rapid rise, sharp peak and rapid decline of the flood hydrograph generate fast flow, increase sediment transport in suspension, and cause rapid deposition. Sustained Froude subcritical flow conditions and low deposition rates are needed for dune migration and cross strata formation.
	Soft-sediment deformation structures, such as convoluted bedding, water escape structures and overturned cross strata are abundant in some systems	Soft-sediment deformation structures less common and of smaller magnitude.	Pore fluid expulsion generated due to the rapid fall of water level during falling stage of the flood, or undercutting and steepening and porewater overpressure. Loading by overlying sediments in conditions of high deposition rates. Lateral shearing of water-saturated sediment under fast and heavily sediment-laden flow during peak flood. More common with high water table, and less common in well-drained settings.
Lithology	Common in-channel mud layers on accretion set boundaries, alternate with coarse deposits.	Where present, in-channel mud occurs as plugs due to channel abandonment	The characteristically rapid waning of high-magnitude floods.
	Thick soft-sediment clast conglomerates at channel bases and above internal erosion surfaces	Thin soft-clast trains or lags, a few grain diameters thick that line erosion surfaces	Erosion and channel bank undercutting during peak flood fast flow. Rapid lowering of water level during the falling stage of flood and associated bank collapse or flowage.
Bioturbation and biogenic structures	Terrestrial in-channel burrows, especially long vertical burrows at accretion set boundaries	In-channel traces absent. Burrowing from channel top down or on floodplain	Frequent and sustained complete or partial channel bed exposure between floods. Sustainably low and highly fluctuating water table.
	In-channel in-situ vegetation, ex-situ plant material and adpressions, roots, and vegetation-induced sedimentary structures	In-channel vegetation, roots and vegetation-induced structures absent. Roots occur only at the top of the channel fills and on floodplain	Colonization of the channel floor during periods of very low flow or lack of base flow in order for vegetation to become established. More common in sub-humid subtropics and monsoonal domain where arborescent vegetation is common. Vegetation-induced structures form as a result of hydrodynamic interactions between living vegetation and the sediment.
Macroforms	Scarcity of well-developed barforms or accretion sets. Dominant low-angle downstream and vertical accretion sets. Where lateral accretion sets occur, they transition into vertical accretion over very short distances. Bars form in modern systems during pre- or post-flood conditions, but have a low preservation potential	Barforms and accretion sets well developed. Lateral accretion sets of point bars, or a combination of downstream, lateral and upstream accretion in braid bars due to systematic bar migration.	Bar formation and migration is inhibited by (1) rapid discharge variations that enhance rapid deposition and aggradation, (2) high suspended sediment concentration, and (3) UFR flow conditions. Sustained lower flow regime conditions are needed for systematic bar migration. UFR conditions during the peak flood remobilize sediment and lower-flow regime bedforms are not preserved. The downstream accretion sets resemble very rapid deposition due to flow expansion in downstream direction from previous flood deposits, rather than bar migration.
Depositional units	Distinct, in places laterally extensive, depositional units bounded by erosion surfaces or by an erosion surface below and a mud drape above, reflect deposition from single flood. Upper boundary commonly shows desiccation, pedogenic modification, or bioturbation. Contain Froude supercritical and transcritical flow structures (UFR) only, or have abrupt transitions from UFR structures to climbing ripples or to a mud drape.	May also contain erosionally bound depositional units, but they are less distinct and dominated by lower-flow regime structures, such as cross strata. Transitions from higher to lower velocity structures are gradational rather than abrupt (e.g. from trough cross strata to planar cross strata and then to ripples). Lack bounding mud layers and in-channel bioturbation, desiccation and pedogenic modification	Erosion and deposition from fast flow, characteristically rapid decline in flow strength, with no low-stage reworking of the flood deposit. High preservation potential due to upper flow regime and suspension transport and rapid deposition, as contrasted to gradual and successive bedload migration. Desiccation during channel bed exposure
Avulsions	Frequently observed in modern monsoonal and subtropical rivers. Evidenced by the dominance of downstream and vertical accretion, the lack of well-developed macroforms, high degree of channel amalgamation, and splay-rich floodplain.	Occur in perennial rivers with high sediment load. Less frequent and more gradational.	Avulsions are common in monsoonal and tropical rivers due to frequent occurrence of floods. The high and highly variable flood discharges and the high sediment concentrations result in rapid channel bed aggradation and resultant superelevation that is a key threshold for occurrence of avulsions.
Fluvial fans	Commonly form fluvial fans, such as fluvial megafans, terminal fans, terminal distributary systems, fluvial distributary systems, and distributive fluvial systems	Do not build fluvial fans	There is a tendency for monsoonal domain and subtropical zone rivers to form fluvial fans, if they have large enough discharge and high enough sediment supply, and enter an unrestricted low-gradient basin, due to high deposition rates and frequent avulsions.