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1 **SUBSIDENCE CONTROL ON RIVER MORPHOLOGY AND**
2 **GRAIN SIZE IN THE GANGA PLAIN**

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13

14 **ABSTRACT.** The Ganga Plain represents a large proportion of the current foreland
15 basin to the Himalaya. The Himalayan-sourced waters irrigate the Plain via major river
16 networks that support approximately 10% of the global population. However, some of
17 these rivers are also the source of devastating floods. The tendency for some of these
18 rivers to flood is directly linked to their large scale morphology. In general, the rivers
19 that drain the east Ganga Plain have channels that are perched at a higher elevation
20 relative to their floodplain, leading to more frequent channel avulsion and flooding. In
21 contrast, those further west have channels that are incised into the floodplain and are
22 historically less prone to flooding. Understanding the controls on these contrasting river
23 forms is fundamental to determining the sensitivity of these systems to projected climate
24 change and the growing water resource demands across the Plain. Here, we present a new
25 basin scale approach to quantifying floodplain and channel topography that identifies
26 areas where channels are super-elevated or entrenched relative to their adjacent
27 floodplain. We explore the probable controls on these observations through an analysis
28 of basin subsidence rates, sediment grain size data and sediment supply from the main
29 river systems that traverse the Plain (Yamuna, Ganga, Karnali, Gandak and Kosi
30 rivers). Subsidence rates are approximated by combining basement profiles derived
31 from seismic data with known convergence velocities; results suggest a more slowly
32 subsiding basin in the west than the east. Grain size fining rates are also used as a proxy
33 for relative subsidence rates along the strike of the basin; the results also indicate higher
34 fining rates (and hence subsidence rates for given sediment supply) in the east. By
35 integrating these observations, we propose that higher subsidence rates are responsible
36 for a deeper basin in the east with perched, low gradient river systems that are relatively
37 insensitive to climatically driven changes in base-level. In contrast, the lower subsidence
38 rates in the west are associated with a higher elevation basin topography, and entrenched
39 river systems recording climatically induced lowering of river base-levels during the
40 Holocene.

41

42

INTRODUCTION

43
44 Many of the rivers of the Ganga Plain are prone to abrupt switching of channel courses
45 (avulsion) causing devastating floods over some of the most densely populated regions on the
46 globe. The Kosi River that drains central Nepal and discharges onto the Ganga Plain of Bihar
47 State has a well-documented history of frequent channel avulsion and flooding (Wells and
48 Dorr, 1987). During 2008, a single channel avulsion event resulted in a temporary eastward
49 shift of the Kosi River by tens of kilometres where the channel breached its eastern levee
50 resulting in extensive flooding (Chakraborty and others, 2010; Sinha and others, 2005, 2013,
51 2014). Similarly, levee failures and channel avulsion resulted in catastrophic flooding of the
52 Indus Plain of Pakistan in 2010 and the displacement of at least 10 million people (Syvitski
53 and Brakenridge, 2013). The nature and frequency of channel avulsion is also a first-order
54 control on alluvial stratigraphy, defining the geometric distributions of channel and floodplain
55 deposits (Bridge and Leeder, 1979; Slingerland and Smith, 2004). In the Ganga Plain, the
56 distribution of Quaternary channel sands and floodplain muds determines groundwater
57 pathways and associated arsenic pollution (Shah, 2007). Given the significance of floodwaters
58 and groundwater pathways in the Ganga Plain, documenting and understanding variations in
59 the morphology of river channel and floodplain systems represents a research priority,
60 particularly in light of changes in monsoon intensity, glacial meltwater discharge and the water
61 demands of a growing population (Fleitmann and others, 2007; Immerzeel and others, 2010).

62 Systematic variations in the large-scale morphology of the river systems are recognised
63 across the extent of the Ganga foreland basin (fig. 1) (Sinha and others, 2005). Rivers of the
64 east Ganga Plain are characterised by shallow aggrading channels that frequently avulse and
65 flood, whilst those in the west are characterised by degrading systems with incised channels
66 and extensive areas of badland topography. In the east Ganga Plain, numerous channel
67 avulsions and random switching of the loci of fan lobe aggradation has resulted in a net

68 westward migration of >113 km of the Kosi River over the surface of its mega-fan during the
69 last two centuries (Wells and Dorr, 1987; Chakraborty and others, 2010). Palaeochannels are
70 well preserved across much of the surface of the Kosi and Gandak fans (Sinha and others,
71 2014), reflecting the dynamic and mobile nature of these systems. In the west Ganga Plain, the
72 Ganga River is described as a braided channel within a narrow incised valley with exposed
73 cliffs extending 15-30 m above the modern channel in parts (Shukla and others, 2001; Gibling
74 and others, 2005; Shukla and others, 2012). Numerous phases of incision and aggradation are
75 documented within both the Yamuna and Ganga valleys where distinct geomorphic surfaces
76 and facies associations are preserved in exposed valley walls (Shukla and others, 2001, 2012;
77 Gibling and others, 2005; Tandon, 2006).

78 In order to understand the controls on the variations in river morphology along the
79 Ganga Plain, we need to consider a range of possible scenarios. As rivers exit mountain ranges,
80 they commonly evolve into broad alluvial systems where river morphology (channel pattern,
81 geometry, gradient) is typically determined by water and sediment discharges, sediment grain
82 sizes, basin subsidence rates and vegetative patterns (fig. 2) (van den Berg, 1995; Dade and
83 Friend, 1998; Dade, 2000; Duller and others, 2010; Marr and others, 2000; Allen P.A. and
84 others, 2013). In addition, first-order predictions from various studies (for example Paola and
85 others, 1992a; Robinson and Slingerland, 1998; Duller and others, 2010; Allen P.A. and others,
86 2013) are that downstream grain size trends are also controlled by sediment supply and
87 subsidence rate, with increased sediment supply reducing fining rates, and increased basin
88 subsidence increasing fining rates as a result of enhanced rates of deposition or aggradation
89 promoting selective deposition in the proximal region of the basin. Grain size fining trends
90 impact the location of the gravel-sand transition (Dubille and Lavé, 2014), and variations in
91 river morphology (Dade and Friend, 1998).

92 This paper initially quantifies the basin-wide variability in incision and aggradation of
93 the river systems across the Ganga Plain from digital topography using a swath based technique
94 to map relative elevation of channels above or below their floodplains. The implications are
95 that the lateral variations in incision versus aggradation should be recorded in the underlying
96 basin stratigraphy, and that the relative contributions of sediment derived from the western and
97 eastern Himalaya to the Ganges-Brahmaputra Delta are likely to be affected by the relative
98 efficiency of sediment transport and bypass across the Ganga Plain. In addition to quantifying
99 the relief along the valleys of the rivers, we also generate new basin-wide data on subsidence
100 rates and grain size fining rates from the proximal foreland basin near to the mountain front.
101 We finally discuss and analyse these data in context of the observed patterns in incision and
102 aggradation of the river systems across the Ganga Plain.

103

104 A challenge when determining longer-term (millennial) controls on fluvial morphologies is to
105 differentiate signals driven by shorter-term stochastic variations in climate or tectonic activity
106 in the upstream catchment (Benda and Dunne, 1997a; Tucker and Slingerland, 1997; Leeder
107 and others, 1998). Forward models have simulated the effects of varying parameters such as
108 sediment flux and basin subsidence over different timescales relative to the equilibrium time
109 period of the basin, defined as the period required for streams within the basin to attain a steady-
110 state profile (Paola and others, 1992a; Heller and Paola, 1996; Robinson and Slingerland, 1998;
111 Marr and others, 2000). In a system as large as the Ganga Plain, potential short-term ($<10^4$
112 years) controls on sediment flux and grain size could be linked to climatic changes in
113 precipitation patterns, glacial discharge and extreme storm events or earthquakes. In contrast,
114 subsidence rates, which are controlled by topographic loading and the flexural response and
115 subduction velocity of the underlying lithosphere (Sinclair and Naylor, 2012) are unlikely to
116 vary at these timescales. Below, we discuss current knowledge and data available on the

117 potential controlling parameters on the large-scale morphology of rivers across the Ganga
118 Plain: sediment flux, basin subsidence and sediment grain size of rivers across the Ganga Plain.

119

120 *Current Constraints on Sediment Flux, Basin Subsidence and Sediment Grain Size across the*
121 *Ganga Plain*

122 Our current understanding of sediment flux into the Ganga foreland basin is based principally
123 on suspended sediment data from gauging station networks, but the spatial coverage of these
124 data is restricted (Blöthe and Korup, 2013). Advances in detrital cosmogenic radionuclide
125 (CRN) analysis have allowed ^{10}Be concentrations to be measured in modern river sediments,
126 allowing approximation of average denudation rates from the source catchments over
127 timescales of thousands of years (Vance and others, 2003; Lupker and others, 2012; Godard
128 and others, 2014). The published CRN data give an indication of how sediment flux delivered
129 to the foreland basin varies spatially between the major river systems that drain the Himalaya.
130 The mean erosion rates of ~ 1 mm/yr derived from these data can be used to infer the timescales
131 over which the rates are averaged, < 1000 years in this case based on the reduction in CRN
132 production rates with depth. Sediment fluxes calculated from ^{10}Be concentrations measured
133 from the modern river sediment reveal marginally lower fluxes at the western end of the Ganga
134 basin (fig. 3, table 1). However, estimates vary by up to a factor of three between sampling
135 years of a single river, highlighting the difficulty in accurately quantifying sediment flux to the
136 foreland basin using this approach (Lupker and others, 2012). Until we have a better
137 understanding of the controls on the variability in ^{10}Be concentrations, it remains difficult to
138 quantify spatial variations in millennial-scale sediment supply rate from Himalayan
139 catchments. Similarly, longer term erosion rates estimated from bedrock mineral cooling ages
140 of the Greater Himalaya Sequence along the strike of the range do not suggest a significant

141 west to east variation in erosion rates, although rates further east are marginally (~0.5 mm/yr)
142 higher (Thiede and Ehlers, 2013). Denudation rates over the past 4 Myr vary between ~1-2.5
143 mm/yr across the Greater Himalayan Sequence within the Ganga basin, but there are large
144 uncertainties with these data (Thiede and Ehlers, 2013). Furthermore, erosion rates in the
145 Greater Himalaya are thought to be relatively high in comparison to the Lesser Himalaya (Lavé
146 and Avouac, 2001), and as such, denudation rates derived from thermochronology studies in
147 this region do not represent catchment averaged rates. The timescales over which these
148 denudation rates have been averaged may also be too large to reflect spatial patterns in modern
149 or sub-millennial sediment fluxes to the Ganga Plain, and should not be interpreted as
150 comparable rates to those derived from ^{10}Be concentrations.

151 Basin subsidence histories across the Indo-Gangetic Plain requires multiple, well
152 documented wells with good stratigraphic resolution (Allen P.A. and Allen J.R., 2013), but
153 these types of data are not available for this region (Burbank and others, 1996). Therefore, for
154 this study, we calculate the tectonic forcing of subsidence using the depth to the crystalline
155 basement that underlies the Siwalik succession derived from the Seismotectonic Atlas of India,
156 prepared by the Geological Survey of India (Narula and others, 2000), integrated with the local
157 subduction velocities (Sinclair and Naylor, 2012).

158 Grain size data from the principal Himalayan rivers are not available, and are therefore
159 a key component of the new data presented in this study. We note that detailed downstream
160 grain size fining trends have been analysed from smaller Himalayan rivers (Dubille and Lavé,
161 2014) that drain the foothills termed 'Piedmont Rivers' (Sinha and Friend, 1994). The grain
162 size data show a clear transition from gravel to sand in the rivers at approximately 8-20 km
163 from the mountain front. Given the order of magnitude increase in catchment size and likely
164 sediment supply from the larger rivers that drain the high mountains of the Himalaya, it is

165 reasonable to predict an increase in distance from the mountain front to the gravel-sand
166 transition for the main rivers presented here.

167 REGIONAL CONTEXT

168 The Himalayan foreland basin formed as a result of the ongoing collision between the Indian
169 and Eurasian plates, where crustal thickening generates high topography that is supported by
170 the flexural rigidity of the underlying lithosphere (Lyon-Caen and Molnar, 1985; Flemings and
171 Jordan, 1989; Burbank and Beck, 1991; Burbank, 1992; Brozovic and Burbank, 2000). Along
172 the strike of the mountain range, variations in lithospheric rigidity and basement faulting are
173 believed to have modulated both basin width and large-scale patterns of subsidence (Burbank
174 and others, 1996). The Himalayan orogen is split into four major structural units that run
175 broadly parallel from west to east (fig. 1). These units are from south to north: the Neogene
176 Siwalik Group, the Proterozoic Lesser Himalayan Sequence, the Proterozoic-Ordovician
177 Greater Himalayan Crystalline Complex and the Proterozoic to Eocene Tethyan Himalayan
178 Sequence (Yin, 2006). These lithological units are bound by major faults, the most active of
179 which is the Main Frontal Thrust (MFT). The MFT is the most southerly structure, situated
180 between the Siwalik Group and the foreland basin, and absorbs approximately 80% of the
181 $\sim 21 \pm 1.5$ mm/yr convergence) between India and south Tibet in central Nepal (Lavé and
182 Avouac, 2000).

183 Sediment generated by the erosion of the Himalayan mountain range accumulates in
184 the foreland basin. The thickness of the basin fill reduces progressively with distance from the
185 mountain front, consistent with asymmetric subsidence caused by thrusting of the overlying
186 orogen (Burbank and Beck, 1991; Burbank, 1992; Yin, 2006). The basin fill is dominated by
187 the Neogene Siwalik Group and the pre-Miocene Rawalpindi Group (Burbank and others,
188 1996). The Siwalik Group comprises thick molasse deposits formed by the erosional products
189 of the Lesser and Higher Himalaya (for example Kumar and others, 2004). Thin-skinned

190 tectonics associated with the MFT have incorporated these poorly consolidated molasse
191 deposits in the hanging wall of frontal structures, forming the Siwalik Hills which represent
192 the youngest and southernmost topography of the Himalaya (Mugnier and others, 1999). The
193 foredeep basin (*sensu* DeCelles and Giles, 1996) lies immediately south of the Siwalik Hills,
194 forming the Indo-Gangetic Plain. Immediately inboard of the thrust front are several wedge-
195 top basins, locally termed ‘Duns’ that act to buffer the sediment delivery to the foredeep
196 (Densmore and others, 2015). In comparison to the confined bedrock channel both upstream
197 and downstream of the Dun, the laterally unconfined and lower gradient surface of these Dun
198 valleys has promoted sediment deposition during periods of heightened sediment export from
199 the mountains, producing a thick alluvial valley fill. Dun valleys of direct relevance to this
200 study are the Chitwan and Dehra Dun valleys where the Gandak, Ganga and Yamuna rivers
201 flow through prior to passing the MFT and exiting onto the Plain (fig. 1).

202 The Ganga Plain (and henceforth Plain) forms the central third of the Indo-Gangetic
203 Plain and covers an area of 250,000 km², whilst the drainage area of the entire Ganga basin is
204 in excess of 1,060,000 km² (Singh, 1996). The hydrology of rivers draining the basin is
205 dominated by the Indian Summer Monsoon (ISM), when over 85% of the annual rainfall falls
206 between June and September (Sinha and Friend, 1994; Tandon, 2006), producing broad peaked
207 annual hydrographs. Along strike gradients in precipitation have been identified using passive
208 microwave data (Bookhagen and others, 2005; Anders and others, 2006) where catchments in
209 the east typically experience more Indian summer monsoon precipitation than those in the west.
210 A strong north-south precipitation gradient has also been identified as a result of orographic
211 enhancement of precipitation, where the heaviest rainfall is induced by the first significant
212 topography encountered by southerly air masses originating from the Bay of Bengal
213 (Bookhagen and others, 2005; Bookhagen and Burbank, 2006; Anders and others, 2006).

214 Apatite fission track ages from across the entire Himalaya do not reveal any systematic change
215 in exhumation rates along the strike of the range (Thiede and Ehlers, 2013).

216 Sediment carried into the foreland basin that is not immediately deposited, typically
217 sand and finer material, can continue downstream via the Ganga, Brahmaputra and Indus rivers
218 ultimately reaching the sea where it accumulates in the Bengal and Indus fans. This fraction
219 represents up to ~90% of the total sediment load exported from the Himalaya (Lupker and
220 others, 2011). Sediment trapped within the foreland basin is deposited across vast alluvial fans
221 that are separated by broad interfan or interfluvial areas that are drained by foothill or Plain fed
222 rivers (Jain and Sinha, 2003; Sinha and others, 2005). These interfan parts of the basin are filled
223 primarily with sediments eroded from the frontal Siwalik range, and sediments derived and
224 reworked locally from the Plain (Sinha and others, 2005).

225 The rivers feeding the Plain can be divided into mountain, foothill and Plain fed (Sinha
226 and Friend, 1994). Mountain-fed rivers originate from large source areas within the Himalayan
227 orogen, typically with a glacial source. Foothill or 'Piedmont' rivers have relatively small
228 catchment areas of 20-2500 km² (Dubille and Lavé, 2014) and drain the interfluvial region
229 between alluvial fans created by sediment deposition of the much larger mountain fed rivers.
230 Plain fed rivers repetitively rework sediment deposited by the mountain and foothill fed rivers
231 (Sinha and others, 2005). Grain size measurements in central Nepal across a number of interfan
232 or foothill fed channels have documented a rapid gravel-sand transition occurring ~8-20 km
233 downstream of the mountain front (Dubille and Lavé, 2014). This same rapid transition is
234 consistent with vertical grain size measurements taken from the Siwalik molasse exposed in
235 the frontal Himalayan folds (Dubille and Lavé, 2014). Grain size fining rates have not been
236 documented for the mountain fed rivers.

237

METHODOLOGY

Topographic Analysis

238
239 Effective mapping of channel elevations relative to their adjacent alluvial fan surface reveals
240 spatial variations in both aggradation and incision of active fluvial systems. Existing
241 approaches to identify regions where channels are perched above their adjacent floodplain, or
242 ‘super-elevated’ (Bryant and others, 1995), are typically limited to linear elevation transects
243 across target alluvial fans using digital elevation models (DEMs) (for example Sinha and
244 others, 2005; Chakraborty and Ghosh, 2010; Chakraborty and others, 2010). A number of
245 limitations arise from this approach: (i) the approach is limited in its spatial resolution as each
246 transect only records elevation across a small portion of the fan, which may not necessarily be
247 coincident with areas of highest avulsion risk; (ii) the orientation of the transects does not
248 directly reflect the geometry of either the channel or fan system; (iii) differentiating data noise
249 from geomorphic features such as channel levees that are often comparable in amplitude
250 (Chakraborty and others, 2010), requiring significant degrees of smoothing to pick out first
251 order features of the alluvial fan system (Chakraborty and Ghosh, 2010). Sinha and others
252 (2014) addressed the first of these issues by taking a series of profiles following parallel linear
253 transects at 2 km spacing down the Kosi fan, permitting an assessment of changes in channel
254 super-elevation along the length of the alluvial fan; however, the spatial resolution is still
255 limited to the transects themselves, and suffers from the same problems relating to transect
256 orientation and noise outlined above. Noise reduction could potentially be achieved using
257 swath profiles which provide a means of increasing the signal-noise ratio, and should highlight
258 characteristics of the along-profile topography (Telbisz and others, 2013; Hergarten and others,
259 2014). More recent generalised swath profile methods permit the use of arbitrary, non-linear
260 baselines, such as river courses, enabling the unbiased characterisation of river valley
261 morphology, but averaging along the length of a stream reach, reducing resolution (Hergarten
262 and others, 2014).

263 We present a new, spatially distributed method to map patterns of fluvial incision and
264 aggradation across alluvial fan systems that addresses the above issues. The premise of this
265 method is that when a channel is elevated relative to its floodplain or adjacent fan surface, the
266 adjacent surface will lie below the elevation of the channel; when incised, the adjacent surface
267 will have a higher elevation relative to the channel. Therefore, by mapping every location
268 within the DEM to the closest point in the channel, it is possible to assess the relative elevation
269 of the channel compared to the rest of the fan.

270 In order to produce maps of channel super-elevation, we use a swath-based method,
271 similar to that developed by Hergarten and others (2014) to construct generalised swath profiles
272 using curvi-linear baselines.

273 The first step in our procedure is to extract the trunk channel on the alluvial fan from
274 the DEM. For this work, river networks were extracted from a 90 m resolution Shuttle Radar
275 Topography Mission (SRTM) DEM using ESRI ArcMap v10.1, using a steepest-descent flow
276 routing algorithm. Channel elevations along these river networks represent the elevation of the
277 water surface in 2000, the time of the SRTM data capture. The root mean squared error (RMSE)
278 of these data in mountainous regions is ± 7.75 m, while in less mountainous regions, the RMSE
279 of the SRTM is ± 14.48 m (Amans and others, 2013). Given that the flow stage will be highly
280 variable through the year, there may be a small impact on these results, although this is likely
281 to be within the RMSE error of the DEM. As such, the use of these data should provisionally
282 be limited to the interpretation of very large scale patterns or where relief exceeds the RMSE
283 of the data ($\sim \pm 10$ -15 m).

284 These trunk channels are subsequently used as a baseline along which we generate an
285 80 km wide swath. This swath determines the region in which we map the relative super-
286 elevation of the trunk channel, and in other applications can be modified for other river systems
287 as required. Within this neighbourhood, we iterate through every pixel, p_i , and map it to the

288 nearest point in the trunk channel baseline, following Hergarten and others (2014); DEM pixels
289 for which the closest point on the baseline is at either of the termini are excluded. The elevation
290 difference between the fan surface and the nearest point on the trunk channel is then calculated,
291 with the resultant swath revealing spatial variations in the elevation of the fan surface relative
292 to the closest point in the active channel (fig. 5). Negative values indicate areas of the fan that
293 are lower in elevation to the closest point in the trunk channel (the channel is perched above
294 the neighbouring fan surface); these areas are shaded red on the swath in figure 5. Conversely,
295 where the trunk channel is entrenched, elevations on the neighbouring fan surface are greater
296 than the closest point in the trunk channel; the more entrenched portions of the swath are
297 coloured in blue on figure 5. Areas of the swath which are at a similar elevation to the channel
298 are shaded in yellow, and areas more than 100 m above the channel are in purple, which
299 typically represents mountainous topography.

300 Channel lengths extending from the Himalayan mountain front (defined as the most
301 southerly area of notable relief) to the Ganga trunk stream were extracted for each river.
302 Longitudinal profiles of each river were also extracted from the DEM from which slope values
303 averaged over a 10 km moving window were then calculated. Normalised channel steepness
304 (k_{sn}) was also calculated at the fan apex using a reference concavity of 0.5, to allow comparison
305 of channel gradients independently of upstream catchment area (Wobus and others, 2006; Allen
306 G.H. and others, 2013). Similar profiles were constructed from the surface of the adjacent
307 floodplains, or valley tops where channels were entrenched in the west Ganga Plain. These
308 profiles followed transects that were broadly parallel to the channel.

309

310 *Basin Subsidence*

311 The distribution of sediment deposition, or the spatial apportioning of sediment sorting, at a
312 given distance (x) can be expressed as:

$$R(x) = \frac{r}{q_s} \quad (\text{eq 1})$$

313 where r is the rate of sediment deposition, and q_s is sediment flux (Paola and Seal, 1995). R
 314 can be made non-dimensional such that:

$$R^*(x^*) = C_0 \frac{r(x)}{q_s(x)} L \quad (\text{eq 2})$$

315 where R^* is the non-dimensional function that describes the distribution of deposition, C_0 is the
 316 volumetric sediment concentration on the bed, r is the rate of sediment deposition, q_s is the
 317 sediment flux, and L is the length of the depositional stream (Paola and Seal, 1995). Assuming
 318 that the rate of sediment deposition is controlled by the rate of tectonic subsidence, defined as
 319 $\sigma(x)$, $R^*(x^*)$ can also be expressed as:

$$R^*(x^*) = (1 - \lambda_p) L \frac{\sigma(x)}{q_s(x)} \quad (\text{eq 3})$$

320 where L is the length of the depositional stream, $q_s(x)$ is the rate of decay in sediment flux
 321 downstream ($L t^{-1}$), λ_p is the sediment porosity and x^* is the normalised longitudinal location
 322 along a deposition system of length L (Duller and others, 2010). If R^* determines how sediment
 323 sorting is apportioned spatially, most sorting will occur at the upstream end of the system where
 324 the greatest proportion of sediment is deposited (Duller and others, 2010). The rate of down-
 325 system grain size fining can also be described by the fractional Exner sediment mass balance
 326 (Paola and Seal, 1995), also incorporating tectonic subsidence through the R^* function:

$$\frac{df}{dx^*} = f \left[R^* \left(1 - \frac{1}{J} \right) \right] - \frac{1}{J} \frac{dJ}{dx^*} \quad (\text{eq 4})$$

327 where f is the fraction of a given sediment size in the deposit and $J = p/f$, where p is the fraction
 328 of a given sediment size in the transporting system (Duller and others, 2010). This predicts a

329 correlation between subsidence rates, sediment grain size fining rate and hence river
330 morphology.

331 The methodology for calculating the tectonic forcing of subsidence of the surface near
332 the mountain front uses new maps of the depth to crystalline basement derived from seismic
333 data combined with known shortening rates (Stevens and Avouac, 2014). The approach doesn't
334 use the depth to basement, but instead utilises the gradient of the basement nearest to the
335 mountain front (Sinclair and Naylor, 2012). By reconstructing the gradient of the basement of
336 the subducting slab (θ) and combining it with known convergence velocities (V_{con}) between
337 the Ganga Plain and the Himalaya (Stevens and Avouac, 2015), we can derive the vertical
338 velocity which determines the modern subsidence velocity at the surface (V_{sub}) at point x using:

$$339 \quad V_{sub}(x) = V_{con} \tan \theta(x) \quad (\text{eq 5})$$

340 This tectonic forcing of surface lowering (subsidence) at the mountain front remains steady as
341 long as the following remain constant: 1) the mean distribution and magnitude of topography;
342 2) the density structure of the mountain range; 3) the convergence velocity between the
343 subducting lithosphere and the distributed load of the range, and 4) the gradient of the
344 subducting lithosphere. Within these parameters, the most likely to vary at a high spatial and
345 temporal scale is the distribution of topography, as thrust units are accreted at the front of the
346 range. Analogue and numerical experiments from thrust wedges indicate that fluctuations in
347 frontal accretion versus internal thickening of the wedge can result in punctuated topographic
348 growth at a timescale characterised by the length of accreted thrust sheets divided by the
349 convergence velocity (Hoth and others, 2007; Naylor and Sinclair, 2007). For the Himalayas,
350 typical spacing of thrust units are approximately 12 km, which when divided by a mean
351 convergence velocity of 18 km/Myr yields a timescale of probable topographic variations of
352 0.66 Myr.

353 Additionally, the rate of stratigraphic onlap of the Siwalik Group onto the basement of
354 the foredeep is 19 ± 5 km/Myr which is comparable to the convergence velocity, suggesting
355 these parameters have been in steady state for the recent history of the thrust wedge and
356 foreland basin system (Lyon-Caen and Molnar, 1985; Mugnier and Huyghe, 2006). Based on
357 these arguments, we do not envisage the tectonic forcing of subsidence to have varied
358 significantly for at least the last 100,000 years.

359 For this study, the time interval of interest is the period over which the present
360 morphology of the river systems of the Ganga Plains is defined; this interval may be
361 approximated by the topographic relief of the fluvial system divided by the sediment
362 accumulation rates. In this case, the local relief of the incised and super-elevated channel
363 systems is up to 30 m. Holocene sedimentation rates for the proximal basin are of the order of
364 1 mm/yr (Sinha and others, 1996). Based on these rates, we propose that the time interval of
365 interest in determining the basin's surface morphology is approximately 30,000 years.
366 Consequently, we see no reason to consider that subsidence rates have varied at any given
367 location in the basin during the development of the present-fluvial morphology across the
368 Ganga Plain.

369 The long-term ($>10^6$ yr) and recent convergent velocity between the subducting plate
370 and the Himalayan topography can be approximated from the stratigraphy of the foreland basin,
371 and modern GPS data respectively. As outlined above, stratigraphic sequences observed in
372 deep well and seismic data imply convergence rates of between ~ 10 - 20 mm/yr over the past
373 15 to 20 Ma from these data (Lyon-Caen and Molnar, 1985). Contemporary GPS data (Feldl
374 and Bilham, 2006; Stevens and Avouac, 2015) have demonstrated along strike differences in
375 modern India-Tibet convergence rates. Rates in the eastern Himalayan arc are typically 18-20
376 mm/yr compared to 12-15 mm/yr in the west. The tectonic displacement of fluvial terrace
377 surfaces in central Nepal (Lavé and Avouac, 2000) and northwest India (Wesnousky and

378 others, 1999) further support a systematic east to west decrease in convergence rates with
379 estimates of 21.5 ± 1.5 mm/yr and 11.9 ± 3.1 mm/yr, respectively.

380 Models calibrated against gravity data have also indicated that the flexural rigidity of
381 the Ganga Basin varies along strike of the basin (Lyon-Caen and Molnar, 1985; Jordan and
382 Watts, 2005). Jordan and Watts (2005) demonstrated that the central Himalayan foreland
383 basin, which relates to the west Ganga Plain, has a higher effective elastic thickness (~ 70 km)
384 compared to regions in the east and west (~ 30 -50 km).

385 The gradient of the basement beneath the proximal foreland basin (θ) is measured using
386 the depth to basement plots of the Ganga basin derived from depth converted reflection seismic
387 data (fig. 4A). The dip of the basement beneath the mountain front has been calculated using
388 the average gradient of the first 30 km of each profile basin-ward of the mountain front, thus
389 reflecting a control on basin subsidence velocities of the proximal basin. Six cross-sections of
390 the foreland basin have been generated from these plots and second order polynomial equations
391 and curves have been fitted through the data to extend the cross section to a point beneath the
392 mountain front to account for increasing rates of subsidence close to the mountain front
393 (Sinclair and Naylor, 2012). A range of V_{sub} values have been calculated along the course of
394 each river using variable V_{con} values to assess the impact on V_{sub} estimates. The variable
395 convergence rate estimates used are based on Stevens and Avouac (2015) with values of
396 13.3 ± 1.7 mm/yr for the Yamuna and Ganga, 18.5 ± 1.8 mm/yr for the Sharda and Karnali and
397 20.2 ± 1.1 mm/yr for the Kosi and Gandak. Previously mentioned convergence rate estimates
398 from Wesnousky and others (1999) and Lavé and Avouac (2000) have also been included in
399 this analysis.

400

Grain Size

401 Extensive coarse gravel bars dominate the bed of the major rivers of the Ganga Plain as they
402 exit the mountain front. During the low-flow season (October-May), a considerable portion of
403 the channel bed is accessible. If it is assumed that equal mobility conditions (Parker and Toro-
404 Escobar, 2002) are attained during monsoon flows, allowing full reworking of gravel bar
405 material, then the gravel deposits visible during this period should represent bedload
406 transported and deposited during the preceding monsoon (Attal and Lavé, 2006). Equal
407 mobility implies that the grain size distribution of the annual transported yield is finer than that
408 of the gravel in the armoured surface exposed at low flow, and similar to that beneath the
409 armoured layer in the subsurface (Parker and Toro-Escobar, 2002). Grain size measurements
410 were taken from ~30 to 50 km upstream of the mountain front down to the gravel-sand
411 transition of each of the Kosi, Gandak, Sharda, Ganga and Yamuna rivers. Ideally,
412 measurements would have been carried out at regular intervals but sampling was restricted by
413 access and in-channel structures. Where large engineered dams (barrages) have been
414 constructed to divert water into channels for irrigation were present, samples were taken at
415 least 1-2 km upstream or downstream of the structure to minimise localised hydrodynamic and
416 trapping effects, this being the distance over which the influence of the barrage appeared to
417 dissipate.

418 Grain size measurements were taken of both the surface and subsurface material using
419 photographic and volumetric analysis, respectively, to account for the effects of surface
420 coarsening (Dietrich and others, 1989; Parker, 1990). Samples were restricted to parts of the
421 bar which appeared recently reworked with imbricated and sub-rounded to rounded gravel
422 (clearly fluvial in origin). Gravel size variations were observed down the length of the gravel
423 bars so sites were chosen in the centre of the coarsest fraction for consistency. At each site, 5
424 – 10 photos were taken of the channel bed to use for photo counting. Particle sizes were
425 measured from each photo by overlaying a regular numeric square grid with 100 nodes, and

426 measuring the intermediate b -axis of each pebble beneath the nodes (Attal and Lavé, 2006;
427 Whittaker and others, 2011). Due to the coarse nature of much of the gravel bars, larger pebbles
428 were often covered by multiple grid intersections. Consistent with the sampling method of Attal
429 and others (2015), pebbles covering n grid intersections were counted n times. This premise is
430 based on Kellerhals and Bray's analysis (1971) using a voidless cube model, although it is
431 noted that this method results in over estimation of D_{84} values (Attal and others, 2015). Results
432 from each photo at a given site were combined to create a single grain size distribution at each
433 sampling location.

434 Volumetric subsurface measurements were taken using techniques documented by a
435 number of studies (Attal and Lavé, 2006; Whittaker and others, 2010; Dubille and Lavé, 2014).
436 Surface material was first removed from the site location (to a depth equal to the size of the
437 largest pebble) and 100-300 kg of material was excavated and sieved through a series of 1, 2
438 and 4 cm square mesh sieves. Pebbles larger than 8 cm were individually weighed, and the
439 weight of each fraction was recorded. For pebbles with b -axis greater than 8 cm, a
440 representative diameter was calculated by assuming that the pebble was roughly spherical and
441 had a density of 2650 kg m^{-3} (Whittaker and others, 2010). A well-mixed representative sample
442 of ~ 1 kg of the fraction < 1 cm was sieved using a 1 mm sieve, from which a ratio was
443 calculated and applied to the whole < 1 cm fraction.

444 The presence of boulders on some gravel bars meant that the recommendation that the
445 largest clast represents $< 5\%$ of the sample mass was not always fulfilled (Church and others,
446 1987). For both surface and subsurface measurements, the effects of excessively large pebbles
447 on D_{84} and D_{50} measurements was assessed using the same method outlined in Attal and others
448 (2015). This process involves the removal of the largest clast from the distribution and
449 recalculating the D_{84} and D_{50} values. This process was then repeated but with the addition of a
450 large clast, similar in mass to the largest clast recorded within that sample. The recalculated

451 D_{84} and D_{50} values with the removal or addition of the largest clast are plotted as upper and
 452 lower error bars on subsurface volumetric samples. Due to the large number of measurements
 453 obtained for each surface sample, following the same procedure on surface grain size
 454 distributions produced minimal variation ($< 5\%$) in D_{84} and D_{50} values. Instead, a more
 455 conservative approach was taken, as outlined in Whittaker and others (2011), where an error
 456 margin of $\pm 15\%$ was applied to account for subjective bias when measuring the intermediate
 457 axis of each pebble beneath the grid node. This margin of 15% was estimated by Whittaker and
 458 others (2011) based on the differences in grain size distribution from repeat sampling of the
 459 same photo.

460 The position of the gravel-sand transition was also mapped for each river, by noting the
 461 point at which exposed deposits were near exclusively sand ($> 95\%$). In some instances, small
 462 patches of gravels were present but represented a very small proportion ($\sim 1-5\%$) of the bed
 463 fraction based on visual observations.

464 Downstream fining rates of the gravel fraction along each river were calculated using
 465 Sternberg's exponential function of the form:

466

$$467 \quad D_x = D_0 e^{-\alpha x} \quad (\text{eq 6})$$

468 where D_0 is the predicted input or initial characteristic grain size in the system (such as D_{84}),
 469 α is the downstream fining exponent and x is the distance downstream (Sternberg, 1875). Linear
 470 functions have also been fitted to account for alternative fining patterns observed in the
 471 literature (Rice, 1999; Whittaker and others, 2011):

472

$$473 \quad D_x = D_0 - \beta x \quad (\text{eq 7})$$

474 where β is the dimensionless fining rate (grain size reduction/km). The two key processes that
 475 are commonly seen to control downstream fining rates in fluvial systems are (1) the selective

476 transport and deposition of particles and (2) abrasion of particles where larger particles are
477 broken down by mechanical processes (Paola and others, 1992b, 1992a; Ferguson and others,
478 1996; Rice and Church, 2001; Attal and Lavé, 2006; Fedele and Paola, 2007; Duller and others,
479 2010). The effect of pebble abrasion is considered negligible in this instance, as the lithology
480 of gravel bars in all rivers was dominantly quartzite, suggesting that grain size fining by
481 abrasion is likely to be similar across all systems. Any differences in grain size fining will
482 likely reflect spatial variations in the grain size distribution of sediment delivered to the Plain
483 from the Himalaya, sediment flux, the spatial distribution of basin subsidence, and local
484 hydraulic and topographic effects (Paola and others, 1992a; Robinson and Slingerland, 1998;
485 Fedele and Paola, 2007; Duller and others, 2010; Whittaker and others, 2011).

486

487

RESULTS

488

Topographic Analysis

489 Along the strike of the mountain front, results of the swath profile analysis are consistent with
490 previous findings (Gibling and others, 2005; Sinha, 2005; Sinha and others, 2005) where the
491 degree of channel entrenchment was found to increase from east to west (fig. 5). In the far west
492 of the Ganga Plain, both the Yamuna and Ganga rivers are clearly entrenched within well-
493 defined broad valleys that are incised into the surface of their respective alluvial fans. These
494 incised valleys narrow with distance downstream of the mountain front from ~20 km to < 1 km
495 at a point immediately upstream of the Ganga and Yamuna confluence at Allahabad (fig. 5B).
496 Close to the mountain front, the valley sides are ~ 30 m high and reduce to 10 – 20 m by ~80
497 km downstream. Lateral incision into valley walls by large meander loops are clearly preserved
498 in the lower half of the Ganga River. The Sharda and Karnali rivers converge at Mahsi, ~100
499 km downstream of the mountain front, to form the Karnali system which is also known as the
500 Ghaghara River in India (fig. 5B). Both tributaries of the Karnali River flow down a well-

501 defined incised valley up to 40 km in width. Downstream of the Sharda and Karnali confluence,
502 the river course turns more sharply to the east and the incised valley loses definition as the
503 degree of entrenchment into the fan surface is reduced. Much of the surrounding floodplain is
504 of a comparable elevation (within 10 m) to the active channel here. Further east, the Gandak
505 and Kosi rivers show minimal signs of entrenchment on the surface of their respective alluvial
506 fans. Much of the surrounding floodplain is of a similar or lower elevation, most notably on
507 the Kosi River (Chakraborty and others, 2010; Sinha and others, 2013; Sinha and others, 2014).

508 The Kosi channel currently occupies the western margin of its alluvial fan where the
509 channel bed is marginally elevated with respect to the surface of the central area of the fan.
510 This pattern is most apparent in the upper ~80 km of the fan where much of the floodplain on
511 the east bank is relatively lower in elevation, in some cases by up to nearly 10 m, than the active
512 channel (fig. 5A). Whilst still within the RMSE of the DEM, this observation appear consistent
513 with independent observations. In 2008 the Kosi River breached its eastern embankment at
514 Kusaha, Nepal (Sinha and others, 2009; Chakraborty and others, 2010). Much of the avulsion
515 belt occupied the depressed area identified as lower in elevation in the SRTM data that were
516 captured several years earlier in 2000. For the remaining length of the fan, the Kosi channel
517 sits at a very similar elevation to that of the fan surface.

518 This west to east gradient also extends beyond the Ganga Basin into the wider Indo
519 Gangetic Plain. East of the Ganga Plain, tributaries of the Brahmaputra River appear similar in
520 nature to the Gandak and Kosi rivers where channels are either at a similar elevation or
521 marginally super-elevated relative to their surrounding floodplain (fig. 5C). Further west, rivers
522 in the Indus basin show similar characteristics to those in the west Ganga Plain where active
523 channels are laterally constrained in broad incised valleys (fig. 5A). Unlike the Ganga Plain
524 however, these valleys widen with distance downstream and the degree of entrenchment

525 appears lower at 10-20 m. This is interpreted as a contrast in dominant controls on channel
526 morphology between the Indus and Ganga basins.

527 Longitudinal river profiles and 10 km averaged slope values extracted from SRTM data
528 show that the Yamuna, Sharda and Karnali rivers exhibit elevated slope values relative to rivers
529 further east within the first 40 km downstream of the mountain front (fig. 6). The Ganga
530 however appears to exit the mountain front with a marginally lower gradient than the other
531 west and central Ganga Plain rivers. In the east Ganga Plain the Gandak and Kosi rivers are
532 lower in gradient, with maximum values of ~ 0.0015 m/m in the first 10 km, rapidly decreasing
533 to ~ 0.0005 m/m by 20 km downstream of the mountain front. The Kosi maintains a more
534 consistent and initially lower gradient down the length of its fan, attaining a maximum value
535 of ~ 0.001 m/m. By 40 km downstream, all of the channel gradients converge at ~ 0.0005 m/m
536 and fluctuate between 0 – 0.001 m/m for the remainder of the profile (fig. 4B). By normalising
537 channel gradient for upstream catchment area (k_{sn}), similar patterns are displayed where
538 systems in the west and central Ganga Plain are typically steeper at the fan apex, with k_{sn} values
539 of 200-300 (fig. 6). Whilst both 10 km averaged slope and k_{sn} values appear to be influenced
540 by some noise along the first 10 km of the Gandak and Kosi profiles, k_{sn} values in the east
541 Ganga Plain appear slightly lower (150-250). With the exception of noise in the Gandak profile,
542 there were no evident knickpoints that were larger in magnitude than the RMSE (~ 15 m) of the
543 data.

544 Comparing the average gradient of the Ganga and Kosi channels to their adjacent fan
545 surfaces, we see that the Ganga fan surface is steeper than the active channel (fig. 7). This is
546 more pronounced at the fan apex where the degree of channel entrenchment is also greatest. In
547 contrast, the surface gradient of the Kosi fan is comparable to the gradient of the active channel,
548 and an absence of significant channel entrenchment is also clearly highlighted.

549

Basin Subsidence

550 The depth to basement plots (fig. 4B) demonstrate an along strike variation in the geometry of
551 the Ganga basin, as has also been recognised previously by Singh (1996). In the east Ganga
552 Plain, the basin is deeper (5000-6000 m) and relatively narrow at ~200 km. The basement has
553 a steep or even convex, distal edge. Further west, the basin widens beneath the Sharda and
554 Karnali rivers. Generally, the basin is shallower here but there are isolated basement lows such
555 as on the Sharda section where the basin reaches 6000 m near to the mountain front. In the far
556 west, the basin is notably shallower at 3000-4000 m and again narrows to ~200 km wide. These
557 variations in depth to basement at the mountain front broadly correlate with the variations in
558 flexural rigidity of the downgoing lithosphere (Jordan and Watts, 2005), with lower rigidities
559 correlating with greater basin depth at the mountain front.

560 Results indicate that the highest average subsidence velocities (V_{sub}) at the mountain
561 front are located in the east of the region near the Kosi fan, with rates of 1.6 ± 0.6 mm/yr. Further
562 west average subsidence rates decrease to 1.4 ± 0.4 mm/yr beneath the Gandak, 0.4 ± 0.2 mm/yr
563 beneath the Karnali, 0.8 ± 0.2 mm/yr beneath the Ganga, and 0.3 ± 0.4 mm/yr beneath the
564 Yamuna. V_{sub} estimates are generally comparable across all but the Kosi and Gandak systems
565 which are notably higher. When these calculated subsidence estimates are compared to
566 documented short term sedimentation rates across the Ganga Plain, values are comparable. An
567 average sedimentation rate of ~ 0.08 (± 0.19) mm/yr for the entire Ganga floodplain has been
568 calculated from chemical mass balance equations (Lupker and others, 2011). Sedimentation
569 rates would be expected to increase exponentially from the cratonic to orogenic margin of the
570 basin however (Flemings and Jordan, 1989), which is consistent with sedimentation rates
571 documented closer to the mountain front. Sedimentation rates of 0.62-1.45 mm/yr have been
572 calculated from radiocarbon dating of organic material in northern Bihar upstream of the axial
573 Ganga channel between the Kosi and Gandak rivers, averaged over a time period of 700-2500

574 yr (Sinha and others, 1996). The comparison of calculated long term subsidence rates of
575 1.6 ± 0.6 mm/yr beneath the Kosi fan with the short term sedimentation rates of 0.62-1.45 mm/yr
576 suggest the system is broadly in balance, with subsidence slightly outpacing sediment
577 accumulation in this part of the basin.

578 *Grain Size*

579 Grain size distributions from sites closest to the mountain front have been compared across
580 each of the sampled channels and are found to be comparable between systems (fig. 8).
581 Subsurface grain sizes documented close to the mountain front on the Yamuna are generally
582 finer than other sites, where a D_{84} value of 66 mm was recorded compared to values ranging
583 between 146-248 mm for the Kosi, Sharda and Ganga. This is attributed to the upstream barrage
584 near Faizabad (~3 km upstream). Compared to similar barrages located close to the mountain
585 front on the Ganga, Sharda and Gandak rivers, a much larger proportion of flow is diverted
586 into extensive canal networks at the Yamuna barrage, resulting in severely reduced flows
587 downstream in the natural channel. During low flow conditions, parts of the Yamuna channel
588 are entirely dry. It seems reasonable to interpret that a greater proportion of coarser material is
589 trapped upstream or very close to the barrage, where there is insufficient discharge to rework
590 or mobilise the coarsest fraction. The D_{84} of the subsurface Gandak sample was also found to
591 be relatively fine (83 mm) compared to the Kosi, Sharda and Ganga samples, which is likely a
592 function of the upstream Chitwan Dun. The coarse fraction of the sediment load is likely to be
593 deposited at the upstream edge of the Dun, where bedrock channels emerge onto the low
594 gradient alluvial surface of the Dun. This is consistent with grain size measurements taken
595 within the Dun which show an overall fining and narrowing of the grain size distributions (fig.
596 9), where the main source of sediment into the channel is restricted to seasonal inputs by
597 ephemeral channels draining the surface of piedmont alluvial fans comprised of Upper Siwalik
598 Group conglomerates (Kimura, 1999; Densmore and others, 2015). Hillslope processes are

599 largely absent in these piedmont catchments where the primary source of sediment is from
600 recycled Siwalik deposits, resulting in a much narrower input grain size distribution into the
601 main Gandak channel. Subsurface D_{84} and D_{50} values measured on the Gandak within the
602 Chitwan Dun vary by ~50 mm, compared to values of 100-200 mm upstream of the Dun (fig.
603 9).

604 In general, there is a strong correlation between subsurface and surface grain size
605 measurements in terms of relative change between values down each profile. Whilst there is a
606 clear surface coarsening visible, local changes in subsurface grain size are also reflected in
607 surface grain size measurements, adding confidence to this sampling approach. This trend is
608 more apparent in the coarser (D_{84}) fraction of the sediment load (fig. 10). The position of the
609 gravel-sand transition and downstream fining rates in each channel shows a considerable west
610 to east variation (fig. 11), irrespective of the grain size distribution of the supplied sediment
611 (fig. 8). The mapped position of the gravel-sand transition relative to the mountain front on
612 each river suggests that gravel progrades further into the basin for rivers in the central and west
613 Ganga Plain (fig. 11). For the Gandak and Kosi rivers in the east Ganga Plain, the gravel-sand
614 transition was documented within 20 km downstream of the mountain front. The gravel-sand
615 transition observed at the Yamuna, Ganga and Sharda rivers was notably further downstream
616 at ~38, 28 and 45 km respectively. The gravel-sand transition was also not found to be an abrupt
617 transition; in most instances a zone of ~2-5 km was noted where the bed was predominantly
618 sand but large patches (up to 25% of the total bed fraction) of gravels were present, although
619 these patches reduced in extent downstream. The position of the gravel-sand transition relative
620 to long channel profiles (fig. 6) is coincident with a break in channel slope, where a steeper
621 channel gradient exists upstream of the transition. This break in slope is less apparent in the
622 east Ganga Plain on the Gandak and Kosi profiles, which may be explained by the noise in the
623 DEM from which the long channel profiles were extracted. It seems more probable that any

624 change in gradient associated with the gravel-sand transition is not as pronounced in the east
625 Ganga Plain due to the gradients of these channels being lower overall. Upstream of the gravel-
626 sand transition on the Sharda, Ganga and Yamuna, channel gradient and the absolute elevation
627 of channels exiting the mountain front are also greater than for the Gandak and Kosi (fig. 12).

628 Fining rates were generally comparable across the Yamuna, Ganga and Sharda rivers
629 (table 3 and fig. 13). For each site, r^2 values determined using each model were also near
630 identical suggesting that the rate of exponential decay is very low (table 3). Using the linear
631 decay model, fining rates of 1.31-4.75 mm/km were observed for D_{84} values across the
632 Yamuna, Ganga, Sharda and Gandak channels whilst a rate of 10.5 mm/km was obtained for
633 the Kosi (fig. 11). This same increase in fining rate is also apparent in the D_{50} fraction, where
634 rates increase from 0.83-1.24 mm/km across systems in the west and central Plain, to 3.21
635 mm/km along the Kosi. Comparable spatial differences in fining exponents (α) obtained from
636 the exponential model were also found and are presented in figure 12. The relatively low fining
637 exponent on the D_{84} fraction of the Gandak is likely to reflect upstream deposition of the
638 coarsest fraction of the sediment load within the Chitwan Dun. As previously discussed, the
639 grain size distribution exiting the Dun is much narrower (that is the D_{84} and D_{50} values are
640 quite similar) on the Gandak relative to other systems (fig. 9). The D_{84} and D_{50} values are also
641 lower than comparable sites at the mountain front in other systems.

642

643

DISCUSSION

644 Topographic analysis of the west Ganga Plain has highlighted the degree of channel
645 entrenchment in the surface of the Yamuna and Ganga fans, compared to the relatively subdued
646 surfaces of the Gandak and Kosi fans further east. Subsidence velocity estimates and
647 downstream grain size fining rates have been found to be highest in the east Ganga Plain, where
648 fan gradients are typically less steep and the gravel-sand transition is found closer to the

649 mountain front. In the west Ganga Plain, the basement depth of the basin is notably lower than
650 the east Ganga Plain, which when combined with known convergence velocities, suggests that
651 the west Ganga Plain is subsiding less rapidly. Assuming basement gradient and convergence
652 velocity yield a reasonable proxy for recent subsidence rates, then clear along strike variations
653 in subsidence rate exist across the Ganga basin. These variations arise from differences in the
654 elastic thickness of the underlying lithosphere (Lyon-Caen and Molnar, 1983, 1985; Jordan
655 and Watts, 2005; Jackson and others, 2008) and/or local inherited morphological variations in
656 the underthrust Indian basement (for example Lash, 1988) combined with varying convergent
657 velocities. This is also consistent with published modelling results that have suggested the
658 equivalent elastic thickness of the lithosphere is lower beneath the east Ganga Plain than the
659 west (Jordan and Watts, 2005). This then raises the question of how, or whether, spatially
660 variable subsidence can be expressed in the surface morphology of the Ganga Plain.

661 *What are the Timescales of the Controlling Processes?*

662 Various modelling studies have suggested that the relative impact of increased and decreased
663 subsidence rates, sediment flux, water supply and gravel fraction on basin
664 stratigraphy/response is strongly dependent on the timescale over which these variations occur
665 (for example Paola and others, 1992a; Heller and Paola, 1996; Duller and others, 2010). To
666 determine whether a forcing is slow or rapid, an equilibrium response time (T_{eq}) is calculated
667 using the square of the basin length divided by the basin diffusivity (Paola and others, 1992a).
668 For the Himalayan foreland basin, a T_{eq} of 2 Myr (± 1 Myr) has been calculated (Heller and
669 Paola, 1992). Variations in parameters that occur over a timescale lower than T_{eq} are
670 subsequently termed as rapid, and those higher than T_{eq} as slow. Along strike of the orogen,
671 variations in the position of the gravel-sand transition on the Yamuna, Ganga, Sharda, Gandak
672 and Kosi rivers are consistent with long term (> 1 Myr) patterns of subsidence across the basin
673 where lower subsidence rates in the west result in a more distal gravel-sand transition than

674 regions experiencing higher rates of subsidence in the east Ganga Plain, where a greater
 675 proportion of sediment is trapped in the proximal part of the basin due to a greater volume of
 676 accommodation being generated (Paola and others, 1992; Marr and others, 2000).

677 *What are the Spatial Characteristics of the Controlling Processes?*

678 Previous works have also suggested that conditions imposed by local subsidence rates could
 679 modulate the gradient of large alluvial fan surfaces over millennial timescales (Allen P.A. and
 680 others, 2013). Longitudinal profiles of the Gandak, Sharda, Ganga and Yamuna rivers reveal a
 681 distinct break in slope at the gravel-sand transition (fig. 6). The transport coefficients of sand
 682 and gravel differ by a factor of ~ 10 (Marr and others, 2000) which, in combination with bed
 683 slope, determine the total flux of sediment at a point on the fluvial surface:

$$684 \quad q_s(x, t) = -v \frac{\partial z(x,t)}{\partial x} \quad (\text{eq 8})$$

685 where q_s is sediment flux, v is the transport coefficient and z is the surface elevation. The
 686 transport coefficients of gravel and sand (v) are reported as 0.01 and 0.1 $\text{km}^2 \text{yr}^{-1}$, respectively.
 687 These transport coefficient values incorporate a number of independently known or
 688 quantifiable variables including water discharge, Shields stress, dimensionless sediment flux
 689 and sediment porosity (Marr and others, 2000). At the gravel-sand transition, an increase in
 690 transport coefficient associated with a change from a gravel-bed to sand-bed river may occur;
 691 however, the associated reduction in channel slope would also be expected to reduce sediment
 692 flux along the profile. Analogue modelling of gravel bed channels has also suggested that a
 693 reduction in both total sediment flux and grain size, as a result of upstream deposition, will
 694 reduce the required transport capacity of the channel downstream. The progressively finer and
 695 smaller sediment load could therefore remain in transport within a channel with a lower
 696 gradient (Paola and others, 1992b). This is consistent with our observations across the Ganga
 697 Plain, where a relatively distinct change in channel slope is associated with the gravel-sand

698 transition (fig. 4B). Interestingly, the positions of the gravel-sand transitions on the Gandak
699 and Kosi rivers are directly comparable to those observed in smaller foothill or ‘Piedmont’
700 rivers (~8 to 20 km downstream of the mountain front). The catchment area of these Piedmont
701 rivers ranges from ~25 – 350 km² (Dubille and Lavé, 2014) whilst the Gandak and Kosi
702 catchment areas are an order of magnitude larger at ~31,000 km² and ~50,000 km², respectively
703 (table 2). The gravel-sand transition on the Gandak, which lies ~100 km west of the foothill
704 systems considered by Dubille and Lavé (2014), was noted at ~20 km. The transition on the
705 Kosi, which lies ~100 km east of their study area, was noted at ~13 km. This suggests that the
706 distance that gravel progrades out from the mountain front is not strongly dependent on
707 upstream catchment area, and therefore unlikely to be dependent on absolute sediment flux,
708 given the dramatically different catchment areas of the foothill and mountain catchments.
709 However, the abrupt change in slope associated with the gravel-sand transition is not a constant
710 feature across these smaller Piedmont rivers. A less abrupt change in channel slope associated
711 with the gravel-sand transition was observed in a number of smaller foothill-fed systems
712 draining the Gandak-Kosi interfan area (Dubille and Lavé, 2014). In this instance, the subdued
713 break in slope was attributed to the relative high proportion of sand relative to gravel
714 transported by the channel at the mountain outlet, where a steep channel gradient was still
715 needed to transport the large proportion of sand downstream of the transition. Where coarse
716 gravels or conglomerate made up less than ~30% of the sediment load, no apparent break in
717 slope was observed at the transition. Whether this same relationship scales up to the larger
718 mountain fed systems has not been examined in detail.

719 The most dominant cause of a rapid reduction in grain size associated with the gravel-
720 sand transition in aggrading systems has been attributed to selective sorting (for example Paola
721 and others, 1992b; Ferguson and others, 1996), where downstream fining by selective sorting
722 is enhanced by bedload sedimentation (Rice, 1999; Dubille and Lavé, 2014). Poorly sorted

723 gravel mixtures and bimodal gravel inputs have been modelled to yield similar fining
724 characteristics (Paola and others, 1992b), suggesting that rapid fining by selective deposition
725 at the gravel-sand transition is insensitive to input sediment grain size distributions. Our current
726 understanding of sediment flux to the Ganga basin is based on a synthesis of published fluxes
727 calculated from ^{10}Be concentrations measured in modern river sediments (table 1). The
728 variability within these data do not allow any robust conclusions to be drawn regarding spatial
729 variations in the long-term sediment supply rate from the Himalayan catchments to the Ganga
730 basin. A more thorough understanding in the observed variability in these ^{10}Be concentrations
731 should be a target of future studies to better understand the role of sediment flux on these
732 systems. Spatial variations in discharge from the Ganga catchment into the Plain could also
733 contribute to the observed morphological signal. Whilst comparable discharge data are not
734 available across these systems, the upstream catchment area of these systems yields an
735 appropriate substitute, where numerous studies have shown a close correlation between these
736 two variables (for example Knighton, 1998). In general, where larger catchment areas are
737 observed in the east, larger discharges would also be expected (table 2) which is consistent
738 with gauged measurements where available (Sinha and others, 2005). Annual precipitation
739 estimates from the Tropical Rainfall Measuring Mission (TRMM) between 1998-2001 across
740 the Himalaya have further suggested that precipitation is typically higher in catchments feeding
741 into the east Ganga Plain (Anders and others, 2006). Interestingly, for a given sediment supply,
742 increased rates of water supply have been modelled to correspond with advancing gravel fronts
743 (Paola and others, 1992a). This doesn't appear to be a factor in the Ganga system where
744 catchment area (and presumably discharge) are greatest in the east Ganga Plain, and the gravel
745 sand transition is found in its most proximal position. Given that it appears that the position of
746 the gravel-sand transition is independent of variations in input grain size distributions, upstream

747 catchment area and sediment flux, it is suggested that longer term patterns of subsidence rate
748 are a governing control on the grain size transition in the modern Ganga Plain.

749 Where gravels prograde farthest downstream in the rivers of the central and west Ganga
750 Plain, channel gradients are found to be steeper close to the mountain front. In the east Ganga
751 Plain, lower channel gradients are observed upstream of the gravel-sand transition, where the
752 relative change in channel gradient across the gravel-sand transition is also less pronounced.
753 Channel gradient measurements derived from SRTM DEM elevations (fig. 6) are not of
754 sufficient spatial resolution or quality to compare with grain size measurements obtained as
755 part of this study. However, given the lack of obvious pattern in grain size distributions
756 measured at the mountain front (fig. 8) and relatively subtle changes in channel gradient at the
757 gravel-sand transition identified in the east Ganga Plain compared to the west (fig. 6), it seems
758 improbable that differences in grain size can account for the along strike variations in channel
759 gradient and grain size fining rates. Other possible interpretations of the variation in channel
760 gradient are that profiles in the east experience higher subsidence rates at the mountain front,
761 resulting in the gravel-sand transition being closer to the mountain front. Late Holocene
762 sedimentation rates of 0.62-1.45 mm/yr (Sinha and others, 1996) on the Ganga Plain are
763 comparable or slightly lower than subsidence velocity estimates beneath the Kosi and Gandak
764 Rivers of 1.6 ± 0.6 and 1.4 ± 0.4 mm/yr, respectively. Comparable information of sedimentation
765 rates in the west Ganga Plain are not available. Alternatively, if there has been greater sediment
766 flux in the west, then the channel may have experienced a greater degree of backfilling. Without
767 evidence for the latter, we suggest higher differential subsidence in the east Ganga Plain as the
768 most probable mechanism. Relative differences in fining exponents of the gravel fraction
769 downstream of the mountain front are consistent with along strike variations in subsidence,
770 where gravels in the Kosi River have a fining exponent two to three times greater than systems
771 in the west Ganga Plain (fig. 13). Whilst the Gandak River has a relatively high subsidence

772 velocity estimate, this same pattern in fining exponent is not as apparent and has been attributed
773 to the buffering role of the upstream Chitwan Dun. Based on these observations, we interpret
774 that spatial variations in subsidence rates play a controlling role in along strike variations in
775 the longitudinal profiles of these rivers and grain size fining rates. However, spatially variable
776 subsidence rates alone do not explain the entrenchment of the western rivers.

777 *Climate and Signal Preservation*

778 Top down changes in sediment and water discharges must have also influenced these systems
779 (Sinha and others, 2005; Wobus and others, 2010). The seasonal nature of water and sediment
780 delivery to the Ganga Plain is highly sensitive to variations in the strength of the Indian summer
781 monsoon during which ~80% of the annual flow is discharged. From marine isotope stage 3
782 into the Last Glacial Maximum (LGM), a combination of low insolation and strong glacial
783 conditions are thought to have significantly weakened the Indian summer monsoon and
784 regional precipitation (Goodbred Jr., 2003; Gibling and others, 2005). This is also reflected in
785 much lower runoff values interpreted from proxy records of palaeosalinity and $\delta^{18}\text{O}$ in the Bay
786 of Bengal during the LGM (Cullen, 1981; Duplessy, 1982). Following the LGM, a variety of
787 proxy records have suggested there was a widespread increase in precipitation, particularly
788 after ~12 ka (Cullen, 1981; Goodbred Jr., 2003; Srivastava and others, 2003). How fluvial
789 systems react to these climate driven variations in water and sediment discharges is more
790 difficult to predict as both incision and aggradation can occur simultaneously within a
791 catchment in response to a single perturbation (Tucker and Slingerland, 1997).

792 Numerous studies have examined the relationship between climatic transitions and
793 phases of fluvial incision and aggradation (for example Tucker and Slingerland, 1997;
794 Goodbred Jr., 2003; Gibling and others, 2005; Srivastava and others, 2008; Wobus and others,
795 2010; Duller and others, 2012; Densmore and others, 2015). Modelling results from Marr and

796 others (2000) have suggested that a rapid increase (over timescales shorter than T_{eq}) in water
797 flux and/or decrease in sediment flux can result in proximal erosion of gravel and advance of
798 the gravel-sand transition. Rapid increases in sediment flux were also found to initiate an
799 increase in proximal channel gradient and retreat of the gravel front (Marr and others, 2000).
800 However, considerable variability in the geomorphic response generated by increased runoff
801 intensity has also been modelled by Tucker and Slingerland (1997) using a physically based
802 model of drainage basin evolution (GOLEM); significant variations in sediment flux were
803 found to result from relatively modest variations in surface runoff, highlighting the difficulty
804 in correlating a specific cause (climatic condition) to effect (geomorphic response). There are
805 also complexities regarding how climatically driven waves of incision and aggradation are
806 propagated downstream of the Himalaya into the Ganga Plain. The effects of stochastic forcing
807 on sediment supply to channel networks has been considered in previous studies (Benda and
808 Dunne, 1997a, 1997b), where the intermittent storage and release of sediment within a
809 catchment has been modelled to dramatically alter the sediment mass balance over thousand
810 year time scales (Blöthe and Korup, 2013). Using both modelling outputs and circumstantial
811 field evidence, unsteady sediment supply was found to affect channel morphology through the
812 generation of sediment waves and transient phases of aggradation (Benda and Dunne, 1997b).
813 If sediment transport through the catchment acts as a non-linear filter and buffers climatic
814 signals (Jerolmack and Paola, 2010; Blöthe and Korup, 2013), it raises the question of what
815 magnitude and wavelength of climatic forcing is capable of being recorded in the sedimentary
816 record of the Ganga Plain?

817 Thermo-luminescence dating of quartz sands and radiocarbon dating on shell and
818 calcrete materials preserved in the upper 2-8 m on the Ganga-Yamuna interfluvial yield ages
819 between 6-21 ka (summarised in Srivastava and others, 2003); these ages suggest that this was
820 when the modern Ganga and Yamuna channels were last connected to the interfluvial floodplain

821 surface (Srivastava and others, 2003; Gibling and others, 2005). Such a situation is consistent
822 with climatic fluctuations associated with the end of the LGM and subsequent strengthening of
823 the Indian summer monsoon at ~11-7 ka (Goodbred Jr., 2003), which could have initiated
824 widespread incision of channels into their respective mega-fans across the Ganga Plain. A
825 corresponding increase in sediment delivery to the Bengal basin was also noted between ~11-
826 7 ka, which translates to a mean sediment load of more than double current load estimates
827 derived from late Holocene deposits in the basin (Goodbred and Kuehl, 2000). Estimates for
828 sediment remobilisation during the early Holocene by incision across the Plain only account
829 for ~2-25% of the total volume of sediment deposited into the basin during this period,
830 suggesting that sediment flux exported from the Himalaya must have been considerably
831 elevated (Goodbred Jr., 2003). Crucially, these observations in the Bengal basin imply that a
832 wide scale climatic perturbation was rapidly propagated down the full length of the Ganga
833 system. Whether this signal was locally amplified by reworking of vast deposits of stored
834 sediment within the Himalaya (for example Blöthe and Korup, 2013) is unknown.

835 A downstream reduction in valley width and channel entrenchment identified on the
836 Yamuna, Ganga and Karnali systems is consistent with a top down wave of incision, most
837 likely initiated by a climate-induced increase in water or relative decrease in sediment
838 discharges during the early Holocene (Tucker and Slingerland, 1997; Goodbred and Kuehl,
839 2000; Goodbred Jr., 2003; Wobus and others, 2010). However, this signal is not apparent in
840 the east Ganga Plain where channels show minimal signs of entrenchment. Whilst some
841 aggradation is thought to have occurred during the late Holocene, these rates are not thought
842 to have been sufficient to infill earlier valley incision (Goodbred Jr., 2003). It therefore seems
843 unlikely that the east Ganga Plain underwent any significant phase of incision, such as that
844 experienced in the west Ganga Plain. This is consistent with well data drilled from the Kosi
845 mega-fan (Singh and others, 1993) that suggested that the Kosi River has maintained a

846 relatively mobile braided channel throughout the Holocene, which migrated across much of the
847 surface of the mega-fan depositing a gravelly sand to fine sand unit.

848 *Subsidence vs. Climate*

849 The Kosi exhibits low channel and fan gradients, flows over the most rapidly subsiding
850 portion of the basin and displays the highest sediment grain size fining rate. Where subsidence
851 rates are higher, proximal vertical sedimentation rates would also be expected to be higher.
852 This results in a smaller amount of sediment remaining in transport further downstream than
853 for a comparable sediment flux in a system experiencing a slower rate of subsidence. The
854 equilibrium gradient of the channel would therefore be expected to be lower where subsidence
855 rates are higher and/or where a greater proportion of the total sediment load is trapped in the
856 proximal basin, as a channel with a lower gradient should be able to convey the smaller
857 sediment load (Robinson and Slingerland, 1998). We hypothesise that patterns of incision and
858 aggradation on this timescale reflect differences in the sensitivity of these systems to climatic
859 forcing of sediment and water flux (Q_s and Q_w respectively), such as that experienced during
860 the early Holocene in response to increased strength of the Indian summer monsoon between
861 ~11-7 ka at the end of the LGM. The sensitivity of these systems to changes in Q_s and Q_w is
862 dependent on the gradient of the equilibrium channel under the new Q_s and Q_w values relative
863 to the subsidence controlled gradient of the wider fan surface, assuming that the channel was
864 not originally entrenched. If the revised equilibrium channel gradient is lower than the original
865 gradient of the alluvial fan, the channel will incise into the surface of the fan apex until a lower
866 channel gradient is attained, producing an incised channel. For a constant climatic forcing of
867 channel lowering along the strike of the Ganga Plain, incision will only occur where channel
868 lowering rates outpace subsidence which will inherently be more difficult to achieve where
869 subsidence rates are higher in the east Ganga Plain (fig. 14).

870

871 Based on upstream catchment areas and satellite-derived precipitation data, it seems
872 likely that systems in the east Ganga Plain experience higher discharges than those further west.
873 Discharge may play a key role in shaping the wider fan morphology, but the position of the
874 modern gravel sand transition is not consistent with these spatial variations in precipitation or
875 possible variations in sediment flux which could be related. Lower channel gradients in the east
876 could reflect these higher water discharges (van den Berg, 1995; Knighton, 1998), but would
877 fail to explain the triggering mechanism behind fan entrenchment in the west Ganga Plain.
878 However, the proximal position of the gravel-sand transition and low channel gradients
879 observed on the Kosi are consistent with the model results simulated under increased basin
880 subsidence rates (over timescales greater than T_{eq}).

881 Whilst absolute sediment fluxes to the basins are uncertain, approximately 90% of the
882 total flux is thought to bypass the basin (Lupker and others, 2011) which would suggest that
883 sediment availability does not limit these systems. Again, the proximal position of the gravel
884 sand transition relative to the mountain front further suggests that the majority of this bypassed
885 sediment is likely to be transported in suspension. Spatial variations in the amount of coarse
886 bedload exported into the Plain, and deposited upstream of the gravel sand transition, is
887 unknown. Whilst beyond the scope of this study, this does appear to be a potential factor that
888 could directly influence the morphology of these systems, as the entirety of this coarser
889 sediment fraction is retained within the Plain. Further work is needed to better constrain the
890 relative proportions of suspended load and bedload within the total sediment fluxes of these
891 systems. The long term morphology of rivers in the east Ganga Plain appears to be primarily
892 controlled by the relatively higher subsidence rates experienced in the eastern end of the basin.
893 Furthermore, these systems appear to have been insensitive to wide scale changes in regional

894 climate, such as that experienced at the end of the LGM, which initiated wide-spread incision
895 in the west Ganga Plain.

896 CONCLUSIONS

897 A modified swath profile analysis has been applied to topographic data across much of the
898 Himalayan foreland basin to characterise the broad nature of incision and aggradation over
899 much of the Indo-Gangetic Plains. In general, we find that the degree of channel entrenchment
900 increases from east to west across the Ganga Plain, and also decreases with distance
901 downstream. First-order subsidence velocity estimates suggest a more rapidly subsiding basin
902 in the east Ganga Plain with rates of up to 1.6 ± 0.6 mm/yr. Further west, subsidence velocity
903 estimates decrease to as little as 0.3 ± 0.4 mm/yr. Grain size fining rates are also found to closely
904 reflect these patterns of subsidence, with the highest fining rates observed in the east Ganga
905 Plain and lowest in the west. Furthermore, data currently available does not support a strong
906 west to east variation in sediment flux at the thousand year timescale. Assuming that ~90% of
907 sediment delivered into the foreland basin is bypassed downstream, it also seems more likely
908 that the relative fraction of bedload delivered to the basin, which is trapped upstream of the
909 gravel sand transition, may have a more direct role on channel morphology than the total
910 sediment flux. We propose that higher subsidence rates are responsible for a deeper basin in
911 the east with perched, low gradient river channels that are relatively insensitive to climatically
912 driven changes in base-level. In contrast, the lower subsidence rates in the west are associated
913 with a higher elevation basin topography, and entrenched river channels recording climatically
914 induced lowering of river base-levels during the Holocene.

915

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FIGURE CAPTIONS

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1229 Fig. 1. Study catchments, location of major Dun valleys and geology (from Yin, 2006) in the
1230 Ganga basin on a 90 m Shuttle Radar Topography Mission (SRTM) derived Digital Elevation
1231 Model (DEM).

1232 Fig. 2. Major controls on large scale channel morphology across the Ganga Plain. These
1233 controls include sediment flux, q_s , to the basin; the distribution of tectonic subsidence, $\sigma(x)$,
1234 across the basin; the spatial distribution of sediment deposition down-system, $R^*(x^*)$; sediment
1235 grain size fining rate, df/dx^* ; and basin subsidence velocity, V_{sub} , which is a product of the
1236 horizontal convergence velocity across the Himalaya, V_{con} , and dip of the basement beneath
1237 the mountain front, θ .

1238 Fig. 3. Sediment flux estimates derived from cosmogenic ^{10}Be concentrations (and errors) and
1239 suspended sediment concentrations. (*) Where no data were available for the Sharda
1240 catchment, catchment-averaged erosion rates derived from the adjacent Karnali catchment
1241 (from Lupker and others, 2012) were used to calculate sediment flux estimates.

1242 Fig. 4. (A) Depth to basement contours across the Ganga basin showing positions of basin cross
1243 sections (black line) for each river and (B) basin profiles constructed using depth to basement
1244 contours in proximity of the Yamuna, Ganga, Sharda, Karnali, Gandak and Kosi rivers. Data
1245 sources: 90 m SRTM DEM and Geological Survey of India.

1246 Fig. 5. Valley topography from swath profile analysis for the three major river basins across
1247 the Himalayan foreland basin from west to east; (A) Indus (B) Ganga and (C) Brahmaputra.

1248 Fig. 6. Longitudinal profiles, 10 km averaged slope and normalised channel steepness (k_{sn})
1249 values for major tributaries of the Ganga basin. k_{sn} values are shown by the thinner black line
1250 on the slope plots. Vertical lines represent the position of the mapped gravel-sand transition.

1251 Fig. 7. Absolute elevation and 10 km averaged slope values of the modern Ganga and Kosi
1252 channels and their adjacent fan surfaces at the fan apex. Fan surface profiles followed transects
1253 that were broadly parallel to the channel, either from the top of the valley side where channels
1254 were entrenched or within ~5 km of the modern channel.

1255 Fig. 8. Surface and subsurface grain size distributions of gravel bar sediment at the mountain
1256 outlet of each river.

1257 Fig. 9. D_{84} and D_{50} values along the Gandak River. A notable fining and overall narrowing of
1258 the grain size distribution is visible as the channel enters the Chitwan Dun, resulting in a much
1259 narrower grain size distribution being transported into the Ganga Plain.

1260 Fig. 10. Comparison of surface and subsurface measurements for the D_{84} (diamonds) and D_{50}
1261 (circles) values at each site across the Ganga Plain. There is a much stronger correlation
1262 between surface and subsurface values in the D_{84} values than D_{50} .

1263 Fig. 11. Downstream distance from the mountain front (MFT) to the gravel-sand transition
1264 (GST) and linear model fining rates on averaged surface and subsurface D_{84} (dashed line) and
1265 D_{50} (solid line) grain sizes.

1266 Fig. 12. Lateral variations in (a) outlet elevation, (b) 10 km average channel gradient and
1267 normalised channel steepness (k_{sn}) at fan apex and (c) proximal fan apex slopes, (d) channel
1268 entrenchment at the fan apex and (e) calculated subsidence velocity (V_{sub}) beneath the proximal
1269 foreland basin across the Ganga Plain.

1270 Fig. 13. Evolution of sediment grain size on gravel bars. Downstream fining exponents (α) for
1271 surface and subsurface averaged D_{84} and D_{50} values downstream of the mountain front for the
1272 Yamuna, Ganga, Sharda, Gandak and Kosi rivers. Error bars were calculated for surface
1273 samples by applying a $\pm 15\%$ error margin to account for subjective bias. Error margins on
1274 subsurface samples reflect the effects of the addition and removal of large clasts from the

1275 sample on D_{84} and D_{50} measurements. It should be noted that the scale of the horizontal axis is
1276 changing between plots.

1277 Fig. 14. Cartoon illustrating the role of variable subsidence rate on surface morphology across
1278 the Ganga Plain, in response to climate-driven variations in water and sediment discharge. The
1279 relative lowering of the surface between time steps t_0 (black line) and t_1 (red dashed line) is
1280 equivalent to a fall in base level, where the gradient of the fan surface is similar between
1281 surfaces. The rate of base level fall is controlled by subsidence in these scenarios where it is
1282 assumed invariant between the two time steps. A change in external forcing (sediment flux,
1283 discharge) leads to an adjustment (reduction in this instance) of the fan slope between t_0 and t_1
1284 (red solid line), which can be accommodated with net aggradation where subsidence rates are
1285 high (1) but requires vertical incision into the fan apex where subsidence rates are lower (2).

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TABLE CAPTIONS

1288 Table 1. Sediment flux estimates summarised from Blöthe and Korup (2013)

1289 Table 2. Catchment and grain size sample summary information

1290 Table 3. D_{50} and D_{84} grain size fining rate data using both exponential and linear equations.

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