Sediment fluxes and buffering in the post-glacial Indus Basin

P. D. Clift*† and L. Giosan‡

*Department of Geology and Geophysics, Louisiana State University, Baton Rouge, LA, USA
†South China Sea Institute of Oceanology, Chinese Academy of Sciences, Guangzhou, China
‡Department of Geology and Geophysics, Woods Hole Oceanographic Institution, Woods Hole, MA, USA

ABSTRACT

The Indus drainage has experienced major variations in climate since the Last Glacial Maximum (LGM) that have affected the volumes and compositions of the sediment reaching the ocean since that time. We here present a comprehensive first-order source-to-sink budget spanning the time since the LGM. We show that buffering of sediment in the floodplain accounts for ca. 20–25% of the mass flux. Sedimentation rates have varied greatly and must have been on average three times the recent, pre-damming rates. Much of the sediment was released by incision of fluvial terraces constructed behind landslide dams within the mountains, and especially along the major river valleys. New bedrock erosion is estimated to supply around 45% of the sedimentation. Around 50% of deposited sediment lies under the southern floodplains, with 50% offshore in large shelf clinoforms. Provenance indicators show a change of erosional focus during the Early Holocene, but no change in the Mid–Late Holocene because of further reworking from the floodplains. While suspended loads travel rapidly from source-to-sink, zircon grains in the bedload show travel times of 7–14 kyr. The largest lag times are anticipated in the Indus submarine fan where sedimentation lags erosion by at least 10 kyr.

INTRODUCTION

Marine sediments represent the longest and most complete archives of continental environmental evolution, and they have been used to reconstruct past patterns, rates of erosion and changing environments at the time of deposition. These in turn can be used to assess the influence that climate or tectonics have over continental environments. However, if the marine record is to be utilized to its full potential, we must first constrain the transport of clastic sedimentary particles from continental sources to the marine sink.

To what extent are sediments deposited in the deep ocean representative of the erosion in the source regions at the time of their sedimentation? Clastic sediments do not relocate instantly into the sea after erosion in mountains, but their transport to the ocean could vary in efficiency and speed, spanning from days or weeks, to $>10^5$–$10^6$ years, because of storage and reactivation en route. Indeed, it has been argued that many Asian river systems have maintained a relatively constant mass flux to the ocean over at least the past 2 Ma because of sediment buffering in the floodplain (Métivier & Gaudemer, 1999). While this has been disputed over timescales $>10^8$ years (Clift, 2006), there is presently little control on how effective this buffering process might be on shorter timescales.

In this study, we present a new sediment budget for the post-glacial Indus River system in order to quantify the degree of buffering in a large drainage system and to see how the deep-sea sediment record might be used to understand erosion in the mountain sources. We use sedimentary provenance data to assess the speed of sediment transport from the source ranges to the ocean and further to the deep-sea. We focus on the post-Last Glacial Maximum (LGM) period because the sediment stored on the shelf clearly sits on a glacial unconformity, thus allowing ready estimation of the total deposited volume. This is also an appropriate time period for estimating eroded volumes onshore because a number of incised fluvial terraces in the floodplains (Giosan et al., 2012a) and mountains are known to date from ca. 10 ka (Bookhagen et al., 2006; Hewitt, 2009). It has been recognized that the Early Holocene was a period of rapid sediment flux to South Asian deltas (Goodbred & Kuehl, 2000; Giosan et al., 2006a), but it is not clear to what extent this sediment was freshly eroded bedrock rather than older fluvial sediments reworked from the floodplains and terraces in the mountains. While some sediment was released by retreating glaciers after the LGM, this affected the central and eastern Himalaya more than the Indus watershed because of the relatively dry climate of this western region (Owen et al., 2008). Hedrick
et al. (2011) noted that while those regions north of and in the rain shadow of the Greater Himalaya experienced only moderate glacial advance at the LGM, there was significant glacial advance in the more monsoonal regions within and south of the Greater Himalaya. Retreat of those glaciers would have released significant sediment quantities that could then be transported to the ocean.

The Indus River basin has a strongly erosive monsoonal climate over parts of the drainage, as well as steep topography and well-defined areas of rapid rock uplift within the source mountains (e.g., Nanga Parbat, south Karakoram metamorphic domes (Zeitler et al., 1993; Maheo et al., 2004)). In addition, there is a strong rainfall gradient across the mountains from wet in the south to the dry regions of Ladakh, Karakoram and Tibet in the north (Bookhagen & Burbank, 2006) (Fig. 1). The sediment sources in the Himalaya and Karakoram are very diverse in composition, and thermochronologic ages that make large-scale sediment provenance quantification simpler than in many drainages (Clift et al., 2004; Garzanti et al., 2005). Crucially, it is known that the origin of the sediment reaching the ocean has changed sharply since the LGM (Clift et al., 2008). The Indus is a sediment-productive river, which used to supply at least 250 Mt year \(^{-1}\) to the Arabian Sea prior to modern damming (Milliman & Syvitski, 1992), although other estimates emphasize a larger range from 100 to 675 Mt year \(^{-1}\) (Ali & De Boer, 2008). Modern and relict wide floodplains flank the Indus and stretch >1300 km from the mountain front to the delta, providing ample potential opportunity for storage and reworking.

**ERODED VOLUMES**

Eroded and incised fluvial terraces are recognized across the northern half of the Indus floodplains, as well as in
the mountains, and constitute a potentially significant
core source of sediment for the river. Reworking of sediment
stored in these terraces is likely a major source of sedi-
ment to the Indus since the LGM (Fig. 1). Giosan et al.
(2012a) noted that the alluvial plain in Punjab shows inci-
sion by the modern Indus and its tributary rivers forming
valleys only 10–20 m deep, but tens of kilometres wide
(Fig. 2). The valleys are separated from each other by
elevated plateau regions or interfluves. Optically stimu-
lated luminescence (OSL) dating of the sediments at the
top of the interfluves has revealed that sedimentation was
ongoing until ca. 10 ka and that the valleys have thus
been cut after that time (Giosan et al., 2012a). A high-
resolution digital elevation model of the region reveals
that the incision is deepest closest to the mountain front
and decreases southward, becoming insignificant close to
the final confluence of the Indus and the Sutlej Rivers
(Figs 1 and 2). By assuming that the northern floodplains
were accretionary and had a smooth topography prior to
incision, we can estimate how much material has been
evacuated from the valleys. Using maximum and mini-
imum estimates of the total area of the incised valleys
(Fig. 2) and assuming a linear reduction in depth from
north to south, we estimate that the amount of sediment
reworked further south since 10 ka is ca. 925 km$^3$ if we
use the volume under the upper presumed interfluve sur-
face in Fig. 2. A minimum estimate can be derived using
the lower interfluve surface, which implies erosion of ca.
370 km$^3$ (Table 1). The upper figure for the eroded vol-
ume assumes that relief on the interfluves is largely the
result of fluvial erosion, whereas the lower figure esti-
mates only the main valley erosion of the Indus and its
Himalayan tributaries assuming that the interfluve relief
is caused by aeolian remodelling. These two estimates
also give a measure of the uncertainty for this source of
sediment, although we do expect that the larger value is
closer to reality because active dunes are mostly limited
to the Thal Desert (the Indus–Jhelum–Chenab inter-

Fig. 2. Shaded topographic map of the northern half of the Indus River floodplains showing the modern courses of the major rivers
and the ‘interfluve’ terraces between them. Optically stimulated luminescence dating has shown that these terraces formed up to 10 ka
and has been incised. Transects across the floodplain show that the river now occupy incised valleys. We calculate the maximum and
minimum limits for eroded sediment as the regions under the dotted lines interpreted as a smooth surfaces at 10 ka. Topography is
from NASA’s Shuttle Radar Topography Mission via GeoMapApp™. Stars show the age control points from Giosan et al. (2012b).

© 2013 The Authors
Basin Research © 2013 John Wiley & Sons Ltd, European Association of Geoscientists & Engineers and International Association of Sedimentologists
fluve) and in the Thar Desert, which is largely south of our study area (the interfluve south of the Sutlej–Indus courses; Fig. 2).

Assessing the volumes eroded from the mountains is more complex and several sectors need to be considered separately if a realistic figure of remobilized sediment is to be achieved for this region. We divide the mountains into four sectors based on climate and geology. One sector covers the Greater and Lesser Himalayan ranges that are influenced by heavy summer monsoon rains. These are expected to be more sediment productive and potentially more tectonically active than those areas to the north that lie effectively on the western edge of the Tibetan Plateau and that experienced major uplift earlier in the Palaeogene (Harris, 2006). The main tributaries that join the Indus from the east are largely sourcing their sediment in this Himalayan regime (Fig. 3). The trunk Indus, the Zanskar and the northern parts of the Sutlej River drain a separate region lying in the rain shadow of the Himalaya. We infer similar conditions for the western parts of the Kabul River, and those streams draining the Karakoram and Hindu Kush (i.e. the Shyok, Hunza, Gilgit, Swat and Chitral Rivers) because these regions are also not heavily monsoonal in character. The third major division lies around the Nanga Parbat massif (Fig. 3) where it has been recognized that especially strong landsliding has caused large-scale damming of the Indus in the past 20 kyr (Korup et al., 2010). We follow the study of Hewitt (2009) in defining this region to be 100 km around the peak of Nanga Parbat itself, including the major rivers passing around the peak to the north and east, as well as the Shigar River (Fig. 3). A separate estimate was made for a fourth sector in which sediment is stored behind landslide rock dams in the Karakoram. Hewitt (1998) mapped a series of large dams and the extent of lake terraces behind them in Baltistan, east of the Hunza River (Fig. 3). Not all the lakes were completely filled with sediment at the point that the dam was breached, but we use the estimates of the breached height to calculate how much material may have been reworked, at least from those dams defined by Hewitt (1998) (Table 1). The

<table>
<thead>
<tr>
<th>Erosion sources</th>
<th>Area of incision (km²)</th>
<th>Volume eroded (km³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Floodplains</td>
<td>92 545</td>
<td>925 370</td>
</tr>
<tr>
<td>Rumbak drainage</td>
<td>162</td>
<td>0.077 0.051</td>
</tr>
<tr>
<td>Volumes eroded outside main valleys</td>
<td>510 609</td>
<td>242 161</td>
</tr>
<tr>
<td>Volumes in Kirthar and Sulaiman Ranges</td>
<td>195 402</td>
<td>93 62</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Length (km)</th>
<th>Maximum</th>
<th>Minimum</th>
</tr>
</thead>
<tbody>
<tr>
<td>Indus in Ladakh</td>
<td>70</td>
<td>7</td>
</tr>
<tr>
<td>Indus upstream of Gilgit</td>
<td>670</td>
<td>71 38</td>
</tr>
<tr>
<td>Indus around Nanga Parbat</td>
<td>220</td>
<td>601 323</td>
</tr>
<tr>
<td>Trunk Indus – Himalaya realm</td>
<td>425</td>
<td>53 15</td>
</tr>
<tr>
<td>Shyok River</td>
<td>270</td>
<td>33 11</td>
</tr>
<tr>
<td>Hunza River</td>
<td>225</td>
<td>27 9</td>
</tr>
<tr>
<td>Shigar River</td>
<td>93</td>
<td>254 137</td>
</tr>
<tr>
<td>Gilgit River</td>
<td>160</td>
<td>20 7</td>
</tr>
<tr>
<td>Zanskar River</td>
<td>180</td>
<td>22 7</td>
</tr>
<tr>
<td>Swat Valley</td>
<td>280</td>
<td>34 11</td>
</tr>
<tr>
<td>Chitral Valley</td>
<td>236</td>
<td>29 10</td>
</tr>
<tr>
<td>Kohistan River</td>
<td>98</td>
<td>12 4</td>
</tr>
<tr>
<td>Nanga Parbat River</td>
<td>165</td>
<td>450 243</td>
</tr>
<tr>
<td>Trunk</td>
<td>290</td>
<td>36 10</td>
</tr>
<tr>
<td>Sutlej – Himalaya realm</td>
<td>230</td>
<td>29 8</td>
</tr>
<tr>
<td>Trunk Sutlej – Tibetan realm</td>
<td>213</td>
<td>27 8</td>
</tr>
<tr>
<td>Trunk Jhelum tributary</td>
<td>169</td>
<td>21 6</td>
</tr>
<tr>
<td>Trunk Chenab</td>
<td>367</td>
<td>46 13</td>
</tr>
<tr>
<td>Chenab Tributary</td>
<td>155</td>
<td>19 5</td>
</tr>
<tr>
<td>Trunk Ravi</td>
<td>217</td>
<td>27 8</td>
</tr>
<tr>
<td>Trunk Beas</td>
<td>176</td>
<td>22 6</td>
</tr>
<tr>
<td>Trunk Kabul – East Branch</td>
<td>300</td>
<td>38 11</td>
</tr>
<tr>
<td>Trunk Kabul – West Branch</td>
<td>400</td>
<td>51 14</td>
</tr>
<tr>
<td>Total</td>
<td>1929</td>
<td>908</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Volume (km³)</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maximum</td>
<td>Minimum</td>
</tr>
<tr>
<td>Total Himalayan rivers</td>
<td>252</td>
</tr>
<tr>
<td>Total Karakoram, Kohistan and Hindu Kush</td>
<td>1122</td>
</tr>
</tbody>
</table>
uncertainties on these numbers are high if poorly defined, but it is clear that damming and rock slides are important sediment buffers in the Karakoram, and that major volumes of sediment have been stored in the mountain valleys. We use an uncertainty of ±50% for the eroded volumes in these rock-dammed valleys, following the logic for the large Himalayan rivers outlined below.

In contrast to the sediment stored behind dams in the major river valleys, we recognize that sediment is also stored within the tributary valleys to these large trunk rivers, as well as on the wider landscape in the upper parts of the drainage. In an attempt to quantify how much sediment may be released from these areas, we consider the drainage that covers ca. 160 km² located SW of Leh and south of the Indus River in Ladakh (Figs 3 and 4). We do not choose the Rumbak drainage for any particular reason and indeed aim to use it as a typical example of a catchment in the rain shadow of the Greater Himalaya. It was selected for study based on its moderate size that allowed it to be surveyed across most of its extent and because it is relatively accessible. Comparison of the Rumbak with other valleys in Ladakh suggests that it is typical in its form and
location. In the future, surveying of wider areas will test whether the Rumbak is indeed representative as we argue. The wider region is made up of a series of smaller catchments of this type, and this is therefore a reasonable starting point for making a budget in this part of the Indus basin. Surveying of this drainage revealed the presence of an incised terrace system running all the way from the confluence with the Indus to the upper reaches (Fig. 5a). While the incision is greatest towards the base of the drainage, this reduces to zero in the upper reaches. We used field measurements of the incision to calculate the volume of sediment removed since the incision of the terrace, which is presumed to be post-glacial, based on comparison with other Himalaya terraces (Table 2).

Table 2. Estimates of the sediment volumes eroded from the individual side valleys within the Rumbak drainage used to determine sediment storage within the upper part of the Indus source regions. See Fig. 4 for location

<table>
<thead>
<tr>
<th>Streams</th>
<th>Length of segment (km)</th>
<th>Distance from confluence (km)</th>
<th>Maximum incision (km)</th>
<th>Maximum channel width (km)</th>
<th>Volume eroded (km³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>15.20</td>
<td>0.00</td>
<td>0.0500</td>
<td>0.0500</td>
<td>0.0190</td>
</tr>
<tr>
<td>2</td>
<td>5.30</td>
<td>9.30</td>
<td>0.0194</td>
<td>0.0388</td>
<td>0.0020</td>
</tr>
<tr>
<td>3</td>
<td>3.40</td>
<td>10.90</td>
<td>0.0141</td>
<td>0.0283</td>
<td>0.0007</td>
</tr>
<tr>
<td>4</td>
<td>4.20</td>
<td>5.10</td>
<td>0.0332</td>
<td>0.0664</td>
<td>0.0046</td>
</tr>
<tr>
<td>5</td>
<td>9.60</td>
<td>1.00</td>
<td>0.0467</td>
<td>0.0934</td>
<td>0.0209</td>
</tr>
<tr>
<td>6</td>
<td>6.10</td>
<td>7.90</td>
<td>0.0240</td>
<td>0.0480</td>
<td>0.0035</td>
</tr>
<tr>
<td>7</td>
<td>5.20</td>
<td>5.60</td>
<td>0.0316</td>
<td>0.0632</td>
<td>0.0052</td>
</tr>
<tr>
<td>8</td>
<td>5.85</td>
<td>8.70</td>
<td>0.0214</td>
<td>0.0428</td>
<td>0.0027</td>
</tr>
<tr>
<td>9</td>
<td>2.20</td>
<td>12.50</td>
<td>0.0089</td>
<td>0.0178</td>
<td>0.0002</td>
</tr>
<tr>
<td>10</td>
<td>2.20</td>
<td>11.00</td>
<td>0.0138</td>
<td>0.0276</td>
<td>0.0004</td>
</tr>
<tr>
<td>11</td>
<td>1.90</td>
<td>7.20</td>
<td>0.0263</td>
<td>0.0526</td>
<td>0.0013</td>
</tr>
<tr>
<td>12</td>
<td>1.70</td>
<td>8.50</td>
<td>0.0220</td>
<td>0.0441</td>
<td>0.0008</td>
</tr>
<tr>
<td>13</td>
<td>2.45</td>
<td>5.10</td>
<td>0.0332</td>
<td>0.0664</td>
<td>0.0027</td>
</tr>
<tr>
<td>Total</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.0641</td>
</tr>
</tbody>
</table>

Fig. 5. Field photographs showing examples of terracing in the Indus basin (a) in the Rumbak drainage (see Figs 3 and 4) where the terrace is around 20 m high, and (b) the Pang gorge, south of the more plains in Ladakh (location on Fig. 3). Terrace here shows ca. 250 m of relief.

We estimated 0.064 km³ (64.1 x 10⁶ m³) of erosion over the 160 km² of the Rumbak Valley, and we used this value to extrapolate how much material might be held in similar systems across the entire mountainous parts of the Indus catchment, including the Kirthar and Sulaiman.
Ranges in the west. These latter ranges are not part of the Himalaya, and they experience a drier climate than the monsoonal southern face of the Himalaya. Until they can be surveyed themselves, we use the sediment yield from the Rumbak Valley because this is located in the Himalaya rain shadow and is also subject to a more arid environment. The amount of material derived from such settings appears to be much smaller than other sources and is estimated here at 279 km$^3$, although clearly the uncertainties of not more than 20% and a standard deviation equivalent to ±56%. Other significant sources of uncertainty to the eroded volumes in other parts of the Himalaya, which have not been the subject of detailed study, are the height of the terraces above the river and the age of the terraces, as the length and width of the river valleys are reasonably easily measured. Based on the existing study of terraces, we estimate height uncertainties of not more than 20% and a similar value for the age, although terraces dating at 8–10 ka are found where studies have been performed in the monsoonally affected Himalaya (Bookhagen et al., 2006; Dortch et al., 2008).

We apply a value of 80.4 m$^3$ km$^{-1}$ of incision to the entire length of the Beas, Ravi, Chenab and Jhelum Rivers, as well as their major tributaries in order to estimate the possible released volumes. Because incision of terraces is known to be high immediately into the frontal Siwalik Hills in Nepal (Lavé & Avouac, 2001), there is no reason to consider a reduced incision adjacent to the floodplains. We further apply this rate to the Himalayan parts of the Sutlej (i.e. those south of the Himalayan rain shadow), the Indus downstream of the Nanga Parbat region and the eastern parts of the Kabul River that lie in the wetter regions east of the Khyber Pass (Fig. 3). Unlike the Rumbak drainage, we assume a constant terrace incision along the length of the valley between the range front and the northern side of the Greater Himalaya. We follow the earlier studies in considering that this terracing has resulted from landsliding and damming of the river followed by back-filling (Bookhagen et al., 2006). The common occurrence of landsliding in the Himalaya (Dortch et al., 2008) and the record of localized lake sediment (Phartiyal et al., 2005) testifies to this being an active and important process in controlling sediment flux in that part of the river basin.

We estimate sediment storage along the trunk Indus in Ladakh using a different data set because of the contrasting drier climate and less tectonically active geological setting here that might result in different conditions compared with the monsoonal Sutlej. Figure 8 shows a transect along the Indus gorge between Leh and Khalsi (Fig. 3) with a number of terrace levels identified at a range of heights above the Indus stream. We note that because no single terrace extends all the way along the 70 km survey, we infer that the valley has been filled locally by landslides damming the river, especially in

![Schematic diagram showing how the volumes of eroded sediment are calculated. (a) In areas where a 10 ka or other post-glacial terrace are known the area of sediment removed since the start of incision can be used to calculated volumes along a river. (b) In a side valley, such as the Rumbak drainage, the depth of incision is normally at a maximum where the stream reaches the trunk river (at point Z) and reduces to zero at the head of the valley (point Y).](image-url)
regions where the gorge is narrow and prone to blocking. Although landslides are common, they are usually rapidly breached and destroyed, but when large rock landslides block the river, then the possibility exists for the formation of long-lived lakes and their subsequent infilling by sediment (Costa & Schuster, 1988). That such events have occurred along the upper Indus is testified by the presence of laminated lake deposits interbedded with fluviol and alluvial fan sediments in the Indus Valley around Leh (Phartiyal & Sharma, 2009). \(^{14}C\) dating of these sediments shows that they roughly predate the LGM and were formed between 31 and 51 ka, during a period of stronger monsoon. We thus know that Terrace B (Fig. 8) was likely incised prior to the LGM. Terrace A, B and C are 30 ka or older, allowing us to assume that the lower terraces (i.e. D–I) post-date the glacial maximum and provide an estimate of the amount of sediment released from the trunk Indus in Ladakh (Table 3). Although better age control would be required for a more precise estimate, our approach here will provide a first-order approximation for the amount of sediment reworked during the Holocene. We estimate that volume to be 5.69 km\(^3\) over the 70-km transect, allowing us to calculate how much material could have been stored in the Indus Valley upstream of the Nanga Parbat region. Although the region of study is reasonably well defined in terms of

Fig. 7. Shaded topographic map of a section of the Sutlej River showing the variation in storage capacity along a stretch of the mainstream. Sections show that depending on the valley geometry significant differences in storage capacity exist. Section D corresponds to the study area of Bookhagen et al. (2006).
volume and to a lesser extent in age, we assign an uncertainty of ±30% to the total estimate, reflecting along strike possible variability in storage capacity. This value is less than that assigned to the Sutlej because the surveyed region of the Indus spans ca. 70 km, rather than being focused in a restricted zone as was the case with the Sutlej. This is considerably less than the sediment storage capacity of the Himalayan Sutlej; 0.081 km$^3$ km$^{-1}$ compared with 0.771 km$^3$ km$^{-1}$ (Bookhagen et al., 2006). We also apply this ‘rain shadow’ rate to those lengths of the Kabul River west of the Khyber Pass, as well as to the Shyok, Hunza, Gilgit, Swat, Chitral, Kohistan and Zanskar Rivers because they all lie north of the Himalayan range crest and thus are generally much drier (Bookhagen & Burbank, 2006).

The total eroded volume is estimated at 1752–4104 km$^3$ (Table 1). The regional breakdown shows that the largest single source of sediment is the peri-Nanga Parbat region (32–40%), with other major contributions distributed between the Karakoram–Hindu Kush (21–27%) and the northern floodplains (21–23%) (Fig. 9). Overall, uncertainties are estimated at ±40% reflecting the relatively well-defined volumes in the floodplains, but the less well-known values from the large river valleys, or eroded from the high-mountain topography.

### DEPOSITED VOLUMES

The Indus alluvial plain gradually becomes depositional south of the confluence with the Punjabi tributaries and shows maximum aggradation near the modern channel belt (Giosan et al., 2012a). The resulting landscape is a unique fluvial mega-ridge of subdued relief extending from the lower Punjab to the Arabian Sea. Under the deposits of the mega-ridge, post-glacial sediments fill up the Indus’ Pleistocene incised-valley system (Kazmi, 1984; Giosan et al., 2006a).
We estimated the volumes stored by the incised valley and the mega-ridge as irregular prisms (Fig. 10) with average widths of 100 and 150 km for the Indus alluvial plain and the delta, respectively. We employed average gradients for the modern alluvial plain (7 cm km$^{-1}$) (Giosan et al., 2012a) and the Pleistocene incised-valley profile (25 cm km$^{-1}$) (Giosan et al., 2006a) starting from a deposit thickness of 90 m at the coast. The depth assumption is supported by interceptions of the valley bottom by our drill sites at the coast (Figs 1 and 11) at Keti Bandar as well as inland (Jati and Gul at ca. 70 km from the coast and Matli at ca. 170 km from the coast). These values should provide a minimum estimate of the valley fill volume with the incised valley extending ca. 250 km from the present coast. Then, using Kazmi’s interpretation of an incised-valley extending over 700 km upstream, we obtained a maximum volume estimate. Based on our drill core data, we assume Kazmi’s view of an incised valley that is very wide and deep with a low gradient that allows the valley to reach far inland. This view is supported by available data so far but will need to be confirmed by future chronologies of the infill in the Sindh plain. However, the Indus incised valley as such envisioned is well within the empirical range for other large palaeovalley dimensions (Blum et al., 2013). Although the proposed flanking of the incised valley by patches of previous highstand deposits (Kazmi, 1984) is in line with some existing models for palaeovalleys (see e.g. Blum et al., 2013), this is not supported by our Matli site where Holocene sediments overlie Eocene limestone at ca. 60 m depth rather than being entirely Pleistocene (Fig. 11) (Giosan et al., 2006b). Nevertheless, to allow for a first-order uncertainty estimate, we calculated minimum and maximum stored volumes in the alluvial plain by assuming rectangular and triangular cross sections for the incised-valley system, respectively (Table 4; Fig. 10a). Within these parameters, together, the alluvial plain and the delta have stored between ca. 2100 and 3600 km$^3$ of post-glacial sediment.

In front of the Indus delta, post-glacial Indus sediments have been stored within the Indus canyon and as mid-shelf clinoforms on both sides of the canyon (Fig. 12) (Giosan et al., 2006b; Limmer et al., 2012). A third clinoform has developed along the Kutch coast with most sediment probably advected eastward alongshore from the Indus delta and redeposited on the mid-shelf by the offshore-directed tidal jet at the Gulf of Kutch mouth (Giosan et al., 2006a). However, it is possible that the Kutch clinoform may only be a surficial feature as the incised valley of the Indus may have not extended so far east. We calculated the maximum volumes stored within the clinoforms below the modern shelf bathymetry down to ~90 m water depth assuming a lowstand erosion of the shelf to that depth (Table 4). This assumption is supported by the depth and width of the incised valley at the coast (Fig. 11) as well as by available seismic profiles (Giosan et al., 2006b; Limmer et al., 2012).

Our recent survey of the Indus canyon shows a transition at ca. 800 m water depth from an active depositional region in the upper canyon to a meandering channel morphology indicative of intermittent turbidity flow (Clift et al., 2013). Sediments recovered from the upper canyon document extremely high rates of sedimentation, while the lower Indus canyon shows instead that turbidite sedimentation ended shortly after 6.9 ka (Clift et al., 2013). Previous coring in the canyon dated a cessation of turbidite sedimentation after 11 ka ca. 75 km from the canyon head (Prins et al., 2000). We calculated the
volume of the sediment prism accumulated in the upper Indus canyon using a 4 km average width, the present average slope of the canyon to 800 m water depth and a predepositional lowstand profile extending from 800 m water depth to 90 m subsurface at the coast (Fig. 10c). Overall, the maximum total volume of sediment stored offshore in the canyon and clinoforms is ca. 2800 or 1900 km$^3$ when we do not consider the Kutch clinoform, values that are of the same order as the Indus inland storage (Table 4). First-order uncertainties vary between 55% for the inland part of the system and 35% offshore.

**MINERAL TRANSPORT TIMES**

The transport times of mineral grains need to be constrained for the Indus system if different phases are to be used to reconstruct erosion patterns in the mountains that can be correlated with changes in climate. Available U-series dating of sediment in the Ganges River, for example, suggests long transport times of ca. 100 kyr for sediments travelling between the range front and the confluence with the Brahmaputra near the delta (ca. 1650 km) (Chabaux et al., 2006; Granet et al., 2007), although residence times were estimated to be shorter ca. 25 kyr for finer-grained sediment (Granet et al., 2010) and most recently the cosmogenic dating argues for sediment transfer times of ca. 1400 years in the Ganges (Lupker et al., 2012) consistent with the similar short residence times deduced in the Andean foreland (Wittmann et al., 2011). Furthermore, different mineral phases will have different lag times depending on their density, shape and grain size. Radiometric dating of mica by $^{39}$Ar-$^{40}$Ar methods and zircons by U-Pb methods in the
Red River has highlighted how different mineral groups may indicate quite different provenances for the same river sediment samples (Hoang et al., 2009, 2010). In the Red River example, the discrepancy was explained as reflecting the slower passage of zircon grains through the river compared with the mica or clay fractions. Zircon is documented to be concentrated in the bedload of large Himalayan rivers (Garzanti et al., 2010; Lupker et al., 2011).

Suspended sediments tend to be transported quickly from source to the ocean, but the grains of the river bedload would be expected to move more slowly and may undergo significant transport only during the strongest flood events (Williams, 1989; Syvitski et al., 2000). Bedload not only moves more slowly than suspended load, but stepwise with potentially very significant storage intervals in fluvial bars and floodplain.

Here we compare the evolution in bulk sediment Nd isotopes with the changing zircon age populations in the same sediments cored at Keti Bandar at the river mouth. Nd is a useful sediment proxy because it is controlled by the average age of the crust from, which it is eroded and is generally not affected by weathering processes (Goldstein et al., 1984). A shift in $\varepsilon_{ND}$ to more negative values...
between 14 and 9 ka was interpreted by Clift et al. (2008) as reflecting stronger erosion of older, more continental rocks compared with younger more primitive ones during the Early Holocene (Fig. 13). Such a trend would indicate a shift of erosion away from arc-like rocks in the Indus suture zone and towards the range front of the Himalaya, as these are composed of deformed but very old Indian plate rocks. Today, the strongest monsoon rains are in the Lesser Himalaya, and the change in $\varepsilon_{Nd}$ values is consistent with a climatic trigger for that shift, not least because the speed of the change precludes tectonic explanations and drainage capture can be ruled out in this case. Nd isotopes are dominated by clay minerals in the fine-grained fraction (Goldstein et al., 1984) and because these usually travel rapidly as suspended load it is not surprising that the change in $\varepsilon_{Nd}$ values is synchronous with the general increase in summer monsoon strength over the same time period (Fleitmann et al., 2003; Herzschuh, 2006).

What is more difficult to explain is why the subsequent reduction in summer monsoon rains is not paralleled by a matching rise in $\varepsilon_{Nd}$ values as the erosion of the Lesser Himalaya weakened. The lack of a response suggests a buffer in the system. We suggest here that the stability of $\varepsilon_{Nd}$ values analysed at the Indus delta since 7 ka reflects a dominant reworking of sediment since that time, and that although the discharge from the Indus tributaries now

Fig. 12. Shaded bathymetric and topographic map of the Indus delta area showing the major depositional areas estimated in our budget. Map is contoured between 200 m water depth and 200 m elevation above sea level at 20-m intervals. Topography is from NASA’s Shuttle Radar Topography Mission via GeoMapApp.

Fig. 13. Diagram showing changing sediment provenance compared with evolving continental climate in SW Asia. (a) The GISP2 ice core climate record (Stuiver & Grootes, 2000), (b) the intensity of the SW monsoon traced by speleothem records from Qunf and Timta Caves (Fleitmann et al., 2003; Sinha et al., 2005) in Oman and by pollen (Herzschuh, 2006) from across Asia (black line), (c) Nd isotopic evolution of sediments from the Indus delta and Indus canyon (Clift et al., 2008) and (d) changing zircon age populations from the delta at Keti Bandar (Clift et al., 2010).

© 2013 The Authors
Basin Research © 2013 John Wiley & Sons Ltd, European Association of Geoscientists & Engineers and International Association of Sedimentologists
favours a stronger erosion from the Karakoram and Kohistan compared with that in the Early Holocene (Alizai et al., 2011), this signal has yet to be communicated to the delta, at least in terms of Nd isotopes (Fig. 14). Indeed, the most recent change in $e_{\text{Nd}}$ values near the delta is a decrease from $-15$ to $-16$, the opposite of the change expected from the water discharge (Fig. 13). We interpret this most recent change to be linked partly to reworking in the floodplain and also to the effect of synthetic damming on the Indus since the 1930s (Inam et al., 2007).

U-Pb dating of zircon sand grains can be used to assess the provenance of the bedload dominated fluvial fraction, and it has been shown that this method is particularly suitable to the Indus because of the diversity of ages in the potential source terrains and the documented variation the age spectra of zircons in the tributaries (Alizai et al., 2011). We note that while the zircon populations seen at the delta change with time in a way that is consistent with the variation in Nd isotopes (i.e. with more Karakoram–Trans-himalayan grains ($<300$ Ma) at the LGM compared with the Holocene), these changes are not synchronous for the two proxies. Figure 13 highlights the fact that while $e_{\text{Nd}}$ values had fallen to a minimum by 9 ka and stabilized at a new low value of $-13$ to $-14$ after 7 ka, the zircon populations remain relatively constant over that entire time period. The most obvious change in zircon ages occurs sometimes after 7 ka, when the proportion of Lesser Himalayan zircons ($1500–2300$ Ma) rose from 22% to 40%, at the same time that Karakoram–Trans-himalayan grains ($<300$ Ma) dropped from 37% to 22%.

We propose that the two provenance proxies are out of phase with one another because they represent different parts of the sediment load. While the bulk sediment Nd signature is dominated by sediment transported in the suspended load, especially clays and mica, the zircons are necessarily part of the bedload because of the high density ($ca. 4.6$ g cm$^{-3}$ compared with 2.75 for illite or $2.68$ g cm$^{-3}$ for plagioclase) and coarse grain size (>100 $\mu$m required for the laser dating method). The platiness of phyllosilicates including micas means that these are preferentially carried in the suspended load compared with the more rounded aspect of zircon grains. We suggest that the time difference between the change in Nd and that seen in zircons reflects the transport time of the zircon in the bedload, that is, ca. 7–14 kyr. This number is significantly less than the 100 kyr estimated by Granet et al. (2007), although the course of the Ganges and the Indus have comparable lengths: ca. 1300 km from the end of the Sutlej gorge to the Indus delta and ca. 1650 km from the end of the Ganges gorge to the confluence with the Brahmaputra. Our estimate is longer than the $^{10}$Be-derived rates of Lupker et al. (2012), although because that 1400 year estimate is based on a quartz-dominated mineralogy, it might be expected to be shorter than the zircon travel time.

**DISCUSSION**

The sediment budget components presented above are clearly subject to significant uncertainties, but it is nonetheless noteworthy that the total sediment volumes deposited, 4050–6725 km$^3$, are only moderately greater than the volumes eroded 1752–4104 km$^3$. This means that much (26–99%, average 54%) of the sediment deposited in the Indus delta, on the Indus shelf and within the Indus canyon since the LGM can be explained in terms of reworking from pre-existing sediments stored in the alluvial plains, rather than from newly generated sediment from bedrock erosion. Because it seems unlikely that erosion would simply cease for any significant length of time, our result implies that the products of new erosion may be stored in terraces or landslides and that erosion is much slower in the absence of extensive glaciers. Our budget reconstruction shows that the floodplains are one of the most important depocentres and sediment sources, but
the volumes stored and released are just 21–23% of the total transport. The concept of sediment flux to the ocean being buffered by the floodplains alone clearly does not apply to the Indus system (Métivier & Gaudemer, 1999).

The 4050–6725 km$^3$ of sediment deposited since ca. 14 ka (the age of the base of the post-glacial delta as dated at Keti Bandar; Clift et al., 2008) corresponds to an average rate of 506–841 Mt year$^{-1}$, assuming an average density of 1.75 g cm$^{-3}$. Considering that the 725–2500 km$^3$ stored under the Sindh alluvial plain post-dates 14 ka, the average rate of mass delivery would be 325–1120 Mt year$^{-1}$. We note that this is considerably greater than the 250 Mt year$^{-1}$ estimated as predamming flux by Milliman & Syvitski (1992) but lies in the range from 100 to 675 Mt year$^{-1}$ estimated by Ali & De Boer (2008). Even in a more conservative estimate with 20 ka as the lower limiting age for the budget, the average rate stays high, at 420 Mt year$^{-1}$. This implies a large degree of temporal variation in sediment flux: low rates in more recent times would have to be balanced by higher rates in the early Holocene, and this would be consistent with the observation of the delta prograding seawards from 14 to 8 ka despite the rapidly rising sea level of that time (Giosan et al., 2006a). Such a pulse mirrors the observation of such an event in the Bengal delta (Goodbred & Kuehl, 2000b) and further argues against major sediment buffering in the net flux to the ocean.

The eroded material derived from the Nanga Parbat region is one sediment flux anomaly that emerges from our budget: while sediment eroded from that region accounts for 32–40% of the total erosional side of the budget, only 3% of zircons can be clearly tied to Nanga Parbat at the delta (Clift et al., 2010). This apparent contradiction is related to the definition of Nanga Parbat. The very young (<12 Ma) U-Pb ages of zircons that are considered characteristic of the massif are found within the gneisses and granites in the core of the uplift (Zeitler & Chamberlain, 1991; Zeitler et al., 1993). However, the region of heaviest landsliding and sediment storage is more widely defined within 100 km of the peak (Hewitt, 2009) and thus encompasses areas of the Kohistan Arc and the southern Karakoram that are experiencing some additional rock uplift centred around the peak. These areas may not be experiencing the degree of uplift and exhumation seen at Nanga Parbat itself, but they are also clearly more strongly uplifted and subject to seismic shaking that other regions in the mountain source regions. If this wider definition of the Nanga Parbat syntaxis is used, then the apparent discrepancy with the eastern Namche Barwa syntaxis is significantly reduced. Petrology and geochemical indicators suggest that ca. 35% of the Brahmaputra sediment downstream of Namche Barwa is eroded from that massif (Garzanti et al., 2004), far more than the values normally assigned for the western syntaxis at 3% based on zircons (Clift et al., 2010) and 6% based on heavy minerals (Garzanti et al., 2005).

We note that there is a reasonable agreement between the estimated sources of the eroded sediments and the zircon populations seen at Thatta and Keti Bandar. Our eroded budget suggests that 21–27% of the total sediment load is being derived from the Karakoram/Hindu Kush and another 32–40% from ‘greater Nanga Parbat’ (including the Kohistan Arc), a total of 53–67%. This proportion is significantly higher than the 24% of zircons with U-Pb ages less than 300 Ma seen at Thatta (Clift et al., 2004), but closer to the 44% of grains dated at <300 Ma from the LGM sediments at Keti Bandar (Clift et al., 2010). Zircons with this <300 Ma age range are considered typical of erosion from Karakoram, Kohistan and Nanga Parbat sources. Given the zircon transport time, we would expect those sediments eroded from terraces in the mountains since 10 ka to have now reached the delta and be represented by the modern sample from Thatta. However, Alizai et al. (2011) noted that the trunk Indus is much less rich in Zr, and thus zircon, compared with most of the Himalayan tributaries. Indus sands yielded only 18 ppm in Zr compared with as much as 63 ppm in the Jhelum. These concentrations are presumed to be representative of the zircon load in the modern rivers, yet we note that the Zr concentrations are much less than the 162 ± 47 ppm reported from the delta by Clift et al. (2010). The analysed sands from the delta may have been enriched in zircon as a result of hydraulic sorting concentrating these heavy minerals in the samples taken, although the concentration is within the range of 152–238 ppm reported by Hu & Gao (2008) for the continental upper crust. Alternatively, hydraulic sorting may have caused dilution of heavy minerals in the samples analysed by Alizai et al. (2011). In the absence of alternative ways of correcting for zircon abundance in the different streams feeding the delta, we use the ratios of Zr in the different streams as a proxy for weighting their concentrations.

Using the modern water discharge values of the main eastern Himalaya tributaries as a proxy for relative discharge, we can calculate a weighted average Zr concentration for sediment from these rivers of 37 ppm. This in turn allows us to calculate the predicted proportion of <300 Ma vs. Himalayan zircons in the mixed sands. If 59–61% of the sediment since the LGM were from the trunk Indus, then we would calculate that ca. 30% of the zircons would be <300 Ma in the mixed sediment. This is closer to the 24% of such zircons recorded by Clift et al. (2004) at Thatta and suggests that our eroded sediment budget lies within error of the predicted budget based on zircon ages. If the zircon concentration correction is in error, then there may be a significant mismatch between our sediment budget and the provenance of Holocene delta sands.

On millenial timescales, sediment flux rates and sources vary significantly, indicating that several buffers are active in the Indus sediment transport system. Although the flux must have been much higher earlier in the Holocene than in the pre-industrial recent, the composition of sediment varies in ways that makes simple linkage between erosion and climate records difficult to
make using marine sediments. This is particularly true if deep-sea fan sediments are considered because the flux of sediment to the final depocentre ceased during the Holocene as a normal consequence of rising sea level and following the classical sequence stratigraphic models of Vail et al. (1977). Prins et al. (2000) dated the youngest turbidite on the upper fan at 11 ka, while our coring of the Indus canyon itself shows that turbidite sedimentation in the lower parts ended shortly after 6.9 ka (unpublished data), during a period of rapid sea level rise and when canyon sedimentation would normally have finished. Conversely, sedimentation has been rapid both on the clinofoms of the Indus shelf and likely offshore the Rann of Kutch, as well as at the head of the Indus canyon where the most rapid recent sedimentation has been occurring (Giosan et al., 2006a). Presumably, if the sea level remains stable and sediment delivery continued at a high rate, then sediment flux to the deep canyon and submarine fan would be re-established in the near future (Burgess & Hovius, 1998). However, in most glacial cycles, sea level has not been as stable for as long as during the Holocene but instead fell again shortly after reaching a maximum (Suiver & Groots, 2000).

This scenario suggests that the sediment flux to the deep-sea would have been re-established on the fan after a moderate non depositional break, but the source of the sediment reaching the fan is primarily from reworking. Since 7 ka, sediment transported to the delta appears to be dominated by sediment reworked from terraces in intramountain tributary valleys with a subordinate but significant reworking of the lower alluvial plains. As sea level fell, erosion of the shelf clinofoms and incision of the neighbouring delta plain would have resulted in a mixed array of sources and ages for the sedimentary grains delivered to the submarine fan during the regression phase. It thus seems unlikely that detailed provenance work on fan turbidite sands could be correlated in any meaningful way with the monsoon climate record on millennial timescales or shorter because the initial erosion of these grains may have occurred during the LGM. An initial reworking phase from mountain terraces occurred in the Early Holocene. This material might then be deposited in the northern floodplains from where it may be reworked again in the latter parts of the Holocene. A lag of at least 10 ky would seem to be implied for sedimentary particles travelling from the peaks of the Karakoram to the bottom of the Arabian Sea. This is not significant if the record is to be used to look at erosion on tectonic timescales ($10^6$ year). If the goal is to look at erosion on millennial timescales, records on the shelf fronting the delta may be of greater use because the delta stratigraphy does show a clear change in provenance during the 14–9 ka period synchronous with the climate change, at least in the fast-travelling suspended sediment load. However, even in these deposits, further reworking of the lower alluvial plain masks any erosional change driven by a weakening monsoon in the upper parts of the drainage since 8 ka.

**CONCLUSIONS**

The sediment flux to the Arabian Sea from the Indus River has experienced significant variability after the LGM. We estimate that 4050–5675 km$^3$ of sediment has been deposited in the incised valley, subaerial delta, shelf clinofoms and the upper Indus canyon. About half of the stored sediment lies offshore on the shelf and in the canyon. While ca. 21–23% of the total depositional volume comes from incision of the upper alluvial plain after 10 ka, the bulk of the sediment is derived from the deep gorges around the Nanga Parbat syntaxis that account for 32–40% of the sediment flux despite comprising only ca. 5% of the area of the mountain sources in the Himalaya–Karakoram–Hindu Kush region. The Karakoram is also an important source, accounting for ca. 21–27% of the total sediment released. Sediment buffering in the mountains appears to be largely controlled by landsliding, which is in part controlled by climate (Bookhagen et al., 2005). Only 5% is eroded from terraces in major river valleys within the monsoonal regions of the Lesser and Greater Himalaya.

Our budget implies that primary weathering of bedrock supplied ca. 46% of the sediment reaching the delta since the LGM. Although climate seems to control the source and rate of sediment supply in the Early Holocene, strong reworking from terraces and floodplains during the Mid–Late Holocene means that changes in erosion patterns after ca. 8 ka are not recorded in the changing composition of delta sediments. We estimate that zircon grains in the bedload take between 7 and 14 ky to travel to the delta compared with shorter travel times for the suspended load. This further inhibits our ability to relate changes in provenance to climatic events as the bedload and suspended load travel times are decoupled and lag the initial climatic trigger.

**ACKNOWLEDGEMENTS**

PC thanks the Leverhulme Trust, the University of Aberdeen and NERC for support for the science project underlying this study. LG acknowledges support from the National Science Foundation and Woods Hole Oceanographic Institution. The authors thank colleagues at National Institute of Oceanography, Karachi for their help over many years of work on and offshore Pakistan and Bodo Bookhagen for advice on sediment storage in the Sutlej Valley. We thank editor Peter van der Beek and reviewers Eduardo Garzanti and Steven Goodbred for their helpful advice in improving the initial submission.

**REFERENCES**


© 2013 The Authors
Basin Research © 2013 John Wiley & Sons Ltd, European Association of Geoscientists & Engineers and International Association of Sedimentologists

Sediment flux in the Indus Basin


tuning of quaternary glaciation between the monsoon-influenced Greater Himalaya and the semi-arid Transhimalaya of Northern India. *Quatern. Int.* **236**, 21–33. doi:10.1016/j.quaint.2010.07.023


