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Channel-levee complexes and sediment flux of the upper Indus Fan

Taylor Landry Berlin

Louisiana State University and Agricultural and Mechanical College

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A Thesis

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The Department of Geology and Geophysics

by

Taylor L. Berlin
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ABSTRACT

The Indus Fan with a volume of 4.5 million km³ is the second largest submarine fan on Earth, only behind its neighbor to the east The Bengal Fan. It formed off the passive margin of Pakistan-India in the northern Arabian Sea. One of the more important aspects of the Indus Fan is its mostly complete sediment record of what has been eroded from the western Himalayan and Karakoram Mountains which act as the main source for the Indus River system. Since the initiation of the Himalaya about 50 Ma, sedimentation rates have fluctuated. This study attempts to calculate a new sediment budget for the Indus Fan and compare channel-levee complex architecture to periods of high and low sediment fluxes. Due to the quality of the 2D seismic available, multiple components of the channel-levee architecture were able to be interpreted which allowed for the reconstruction of how each complex built over time. The results of this study suggest peak sedimentation occurred during the late Miocene to early Pliocene. Multiple data sources support this period of peak sedimentation was caused by the onset of global cooling that began around 3 Ma. The climate change did not allow for fluvial and glacial systems to reach equilibrium. These results differ with previous work in that a steady increase in sedimentation rates were calculated to occur up until the late Miocene. In the early Pliocene, sedimentation rates started to decrease again till recent time.
INTRODUCTION

Channel-levee complexes reflect the nature of sedimentary processes active on a submarine fan. The geometry and architectural elements that are able to be identified in seismic data aid in the effort to accurately describe the morphology of these structures. In the case of the Indus Fan determining the effect climatic variation in Asia and tectonics have on erosion of the Western Himalayan and Karakoram Mountains along with the development of channel-levee complexes over the past 65 Ma may be possible.

Erosion in the modern Himalaya is thought to be controlled in some degree by the amount of monsoonal precipitation (Métivier et al., 1999, Galy and France-Lanord, 2001, Clift et al., 2010). It is well known that during periods of more intense monsoon activity that a positive increase in sedimentation rates might occur (Goodbred and Kuehl, 2000; Clift et al., 2008). Previous work by Métivier et al. (1990), Rea (1992), Burbank et al. (1993) and Clift et al., (2002) each have produced their own sedimentation models. The results of those studies differ slightly due to location biases and/or limiting data sets.

Improved seismic data over the past twenty years has led to the improved mapping and interpretation of the Upper Indus Fan. Studies by McHargue and Webb (1986), Kolla and Coumes (1987), Amir and Kenyon (1996) and Deptuck et al. (2003) have built off each other by using what the previous authors have done but improving or fine tuning their models or interpretations with improved seismic data or coverage. The previous work also differs in that some focus on the larger feeder canyons like in McHargue and Webb (1986) and Kolla and Coumes (1987) while others look at the smaller channel-levee complexes that extend out past the canyons (Amir and Kenyon, 1996; Deptuck et al., 2003).

The goal of this paper is to better understand how the sediment flux into the ocean is affected by climatic variation in Asia and by the erosion of mountains during the Cenozoic. Also to further extend the mapped area and description of channel-levee complexes in the upper Indus Fan. If we can't define the offshore deposited volumes properly then it is impossible to mass balance with erosion and sedimentation. The mapping and compaction correction that are performed are aimed at understanding the sediment flux to the ocean through time. Future models of how tectonics and climate affect erosion can be tested with the representative record of sedimentation in the Arabian Sea calculated in this paper because fan sediment is a proxy for erosion.
BACKGROUND

INDUS FAN

The construction of the Indus Fan is primarily due to the deposition of sediment by turbidity current processes (Naini and Kolla, 1982; Kolla and Coumes, 1984). It has formed off the passive margin of Pakistan-India. The Indus River system transported the sediment from the Western Himalayan and Karakoram ranges to the coast where sediment is either deposited in the Indus Delta, which covers 8,000 km² (McHargue and Webb, 1986), or the delta is bypassed and sediment makes its way to the shelf edge and/or the abyssal plain. Historically there have been three different submarine canyons that have allowed sediment to bypass the shelf and be deposited on the abyssal plain during periods of lower sea level (McHargue and Webb, 1986; Clift et al., 2002; Deptuck et al., 2003). As the turbidity current reaches the abyssal plain a decrease in slope and confinement decreases the velocity of the flow and allows for the deposition of sediment as lobes. During periods of high sediment flow the initiation of erosional channel-levee systems can occur to further transport sediment away from the continental slope.

Indus Fan deposition is thought to have initiated during the late Oligocene to early Mioocene (Qayyum et al., 1997; Clift et al., 2001). This is after the onset of Himalayan uplift during the middle Eocene (Molnar and Tapponnier 1975; Sahni and Kumar 1974; Dewey and Bird 1970). Deposition occurs on the western passive margin of the Indian subcontinent which formed when seafloor spreading started to separate the Seychelles platform from India at the end of the Late Cretaceous (McKenzie & Sclater 1971; Naini and Talwani 1983; Minshull et al., 2008). The fan covers $1.20 \times 10^6$ km² and extends 1,500 km out into the Arabian Sea. It is second in size only to the Bengal Fan located on the opposite side of India. The fan is bounded to the south by the Carlsberg Ridge (see Fig. 1), to the east by the Chagos-Laccadive Ridge and to the west by the Owen and Murray Ridges. The largest fan thicknesses occur in the proximal fan where greater than 10 km of sediment has been deposited (Clift et al., 2002; Naini and Kolla, 1982).

The Indus fan is separated into three different sections, the upper, middle and lower fan (Amir et al., 1996; Kolla and Coumes, 1987). The upper portion of the fan extends from the foot of the continental slope at about 1,000 m water depth to about 3,300 m water depth with the steepest gradient of 1.500 (.11°). The middle fan extends to about 3,900 m depths with an average gradient of 1.700 (.08°). The lower fan has an average gradient of about 1:1000 (0.05°) and extends to depths down to 4,600 m.

Multiple canyon complexes have helped supply Indus River sediment to the Indus Fan throughout its history. The three outlined in figure 2 have been well described by McHargue and Webb (1986), Kolla and Coumes (1987) and Normark and Carlson (2003). The Indus Trough forms more landward on the shelf where the three canyons merge. The canyons are labeled 1-3, 1 being the oldest and 3 the Indus Canyon being the youngest and forming during the
Pleistocene (Kolla and Coumes, 1987). According to Kolla and Coumes (1987), relative ages were assigned to the canyon complexes on the basis of stratal relationships between seismic reflectors. The eastward migration of the canyon complexes can be attributed to multiple factors. Sea level change, Coriolis force, eastward migration of the Indus river and depocenters (Snelgrove, 1967), uplift of the Murray Ridge and channel meander (Komar, 1969).

INDUS RIVER

The Indus River, 2,900 km long, currently drains an area of approximately $1.10^6$ km$^2$ with a sediment discharge of 450 million tons per year before damming compared 50 million tons per year after (Milliman et al., 1984). The fifth largest sediment load in the world (Wells and Coleman, 1984) can be attributed to the drainage of barren, poorly consolidated sediment consisting of glacial and fluvial-reworked detritus that eroded from rapidly uplifting mountain ranges with high relief in the Himalayan and Karakoram region (Giosan et al., 2006; Milliman et al., 1984). The warmer temperatures and the arrival of the monsoon season during the summer months leads to peak discharge of sediment. A combination of seasonal melting of Himalayan glaciers and increased runoff caused by large amounts of rainfall are the key factors to modern day erosion and transportation of sediment to the Indus River.

REGIONAL GEOLOGIC SETTING

The Arabian Sea is located in the northwestern Indian Ocean where numerous types of plate boundaries occur such as a passive margin, subduction zone, mid-ocean ridge and transform plate boundary (Figure 1). The Indus Fan is the dominant sedimentary feature of the Arabian Sea and is located at the triple point between Arabian, Eurasian and Indian Plates. Northwest of the fan the Eurasian Plate is over thrusting the Arabian Plate forming the Makran Accretionary Prism. The transform boundary between the Arabian and Indian Plates extends from the Owen Ridge (Owen Fracture Zone) north to the Murray Ridge and onshore to the Chaman Fault (Fournier et al., 2008). Southwest of the fan, the spreading center between the Arabian and African Plates forms the Carlsberg Ridge. Located east of the fan a series of peaks rising above the seafloor that have been buried by fan sedimentation is the Laxmi Ridge. This topographic high, according to Talwani and Reif (1999) and Naini and Talwani (1982), is a continental block bounded by oceanic crust on either side. The Chagos-Laccadive Ridge to the southeast of the fan is the last structural high which is made up mostly of volcanic rocks that are linked to the Deccan ‘hot-spot’ during the northeastward movement of India (Avraham and Bunce 1977; Chaubey et al., 2001).

STUDY AREA

2D Seismic lines from two different seismic surveys are located off the coast of Pakistan, south southwest of Karachi and cover an area of about 6,000 km$^2$ (Figure 2). All but two seismic lines were collected by Total in 2007. The
two lines used in the depth conversion and decompaction explained later were collected by British Petroleum in 2007. The grid is approximately 180 km from the coast and 45 km from the shelf edge on the upper Indus Fan. Spacing between each line is about 5 km. The water depth ranges from 1,500 m to almost 2,100 m.

Figure 1. Location map with the main morphological features of the Arabian Sea. The dashed white lines representing plate boundaries. The dashed red line outlines the edge of the Indus Fan. 2D seismic lines used for this study shown in red. Makraan Accretionary Prism (MA). DSDP drill site 222 marked by •. Edited from Calves et al., (2008).

UPPER FAN CHANNEL-LEVEE COMPLEXES

Channels are extended negative relief features that are formed/maintained through the transportation and deposition of sediment by turbidity-current flows over a relatively long period of time. The two main types of channels observed on submarine fans, including the Indus Fan, are large leveed valleys (originates from a larger canyon upstream) whose primary role is to act as a sediment feeder system to smaller distributary channels (Mutti and Normark, 1991; Amir and Kenyon, 1996; Posamentier and Kolla, 2003). These unleveed distributary channels are the second main type of channels and help to further
transport sediment to the distal fan where eventually sediment is deposited as depositional lobes. This study focuses on larger channels that are bounded by levees due to the location of the seismic grid on the upper fan and lack of seismic coverage over the middle and lower Indus Fan.

Four types of channel-levee complexes were identified using the channel-fill types proposed by Normark (1970), Mutti and Normark (1991) (erosional, depositional, and depositional/erosional or mixed) and the architectural elements described in Pickering et al. (1995), Clark and Pickering (1996), and Deptuck et al., (2003). These channels help show the variation in channel-levee architecture on the upper portion of the Indus Fan. As stated before the upper fan starts from the edge of the continental slope in about 1,000 m water depth and extends offshore to about 3,300 m water depth (Amir et al., 1996; Kolla and Coumes, 1987). Other elements of channel-levee complexes that are interpreted and described in this study include outer levees, inner levees, erosional bases, continuous high amplitude reflectors (C-HARs), discontinuous high amplitude reflectors (D-HARs), channel-fill high amplitude reflectors (CF-HARs), continuous low amplitude reflectors (C-LARs), discontinuous low amplitude reflectors (D-LARs) and reflection free zones. Examples and descriptions of each of these types of reflectors will be described and illustrated below.
CHANNEL-FILL TYPE

Whether a channel is erosional, depositional or a combination of the two each will have its own unique type of channel-fill deposits. Channel-fill deposits characteristic of an erosional system are basal, coarse-grained residual facies that are overlain by fine-grained, possibly channel-abandonment deposits seen in figure 3B labeled as CF-HARS which will be defined later (Normark, 1970; Mutti and Normark, 1991; Catterall et al., 2010). At the base of the channel-levee complex a distinct erosional surface can be identified that will continue to migrate deeper vertically rather than aggrade (McHargue and Webb, 1986). The vertically deeper migrating channel was unable to be identified in this study. Depositional channel-fill deposits mostly consist of parallel layers of coarse-medium grained sediment that infilled the channel after it’s or original phase of activity as a sediment transport pathway (Figure 3D) (Hamilton; 1967; Normark, 1970; Mutti and Normark, 1991). Also, no erosion of older sediment occurs below the channel-levee complex. If a combination of the two previously described channel-deposits is present then a channel is deemed to be both erosional and depositional (Figure 3B). Determination of channel morphology for this type of channel-levee complex can be much more complex than the previous two. Typically the channel originates as being erosional and transitions to depositional (Normark, 1970; McHargue and James, 1986; Mutti and Normark, 1991).

CHANNEL AGGREGATION AND MIGRATION

As a channel meanders, the channel axis or thalweg migrates laterally in the direction of the outer bank. The channel edge opposite of the cut bank can stack adjacent to one another and form a series of shingled events inclined in the direction of channel migration (figure 4B). The channel edges are able to be interpreted as relics of previous channels that were able to be preserved after multiple cut and fill events. According to Deptuck et al. (2003) these shingled
channel edges will aggrade if the amount of incision between the channel edges is less than the thickness of sediment that accumulates within them. If the incision between channel edges is equal to or greater than the amount of deposition within, no vertical aggradation will occur. A variety of amalgamated and laterally migrated channels can be seen in figure 5.

Figure 4. A) Uninterpreted seismic cross section. B) Interpreted seismic cross section of 4A showing an erosional base, channel migration and shingled channel edges.

Figure 5. Graph showing width and depth changes of stacked channel complexes between lateral and vertical amalgamation of individual channel bodies. Edited from Clark and Pickering (1996).
INNER LEVEES

Inner levees are interpreted as muddy bench-like ‘terraces’ that are situated between the larger and better defined outer levee and the channel edge or channel-fill as seen in figure 6B. These features have been recognized on the Indus Fan by Von Rad and Tahir (1997), McHargue (1991) and on the Bengal Fan by Hubscher et al. (1997). Inner levees can form from both depositional and erosional processes. In a depositional process the inner levee can form from vertical aggradation of inner levee deposits resulting from the overbanking of under-fit channels. In this study inner levees are commonly in the form of quadrilateral squares, rectangles or parallelograms. Deptuck et al. (2003) noted that a sharp erosive surface is commonly found between outer and inner levees and also between inner levees and channel-axis deposits.

![Figure 6. A) An uninterpreted seismic cross section. B) An interpreted seismic cross section of 6A showing outer levees, inner levees and the erosive surfaces that separate them.](image)

OUTER LEVEES

Outer levees are thought to form from the overbanking or flow-stripping of predominantly fine-grained sediments during the passage of turbidity currents (Naini and Kolla, 1982; McHargue and Webb, 1986; Peakall, McCaffrey, and Kneller, 2000; Posamentier and Kolla 2003). More specifically the fine-grained sediments of overbank deposits are thin-bedded and current-laminated (Clark and Pickering, 1996). It is common for hemipelagic and pelagic mudstones to be laminated between the overbank deposits as well. The outer levees are wedge shaped and are thickest near the channel and thin away from the channel (Figure 6B). Interpreted seismic reflections tend to show an increased dip as the levee aggrades. Seismic reflections on the distal part of the outer levee can show internally developed downlap and toplap surfaces.

SEISMIC FACIES

The different elements of channel-levee complexes that were previously mentioned have seismic reflector characteristics that allow for them to be
interpreted. These characteristics consist of amplitude reflector strength, continuity/discontinuity, thickness and the spatial proximity to one another. The following section will describe how the different elements of channel-levee complexes are interpreted using seismic profiles.

High amplitude reflectors in general result from a high contrast in the velocities at which adjacent layers transmit acoustic waves (McHargue 1991). Continuous high amplitude reflectors (C-HARs) were used to mark the base and top of an individual channel-levee complex (Figure 7B). These continuous high amplitude reflectors would onlap onto older channel-levee complexes and are thought to be either distal overbank deposits that get deposited after channel abandonment or hemipelagic drape (Kolla and Coumes, 1987). It is not uncommon for these reflectors to pinch out on older channel-levees. Continuous high amplitude reflectors were also identified in the outer levees of most of the channel-levee complexes (Figure 7B). The reflectors tend to be thickest near the channel and thin further from the channel. McHargue (1986) describes the continuous high amplitude reflectors within the outer levees to be the result of interbedded layers of fine-grained turbidites deposited during overbank flow from the channel.

Figure 7. A) An uninterpreted seismic cross section. B) An interpreted seismic cross section of 7A showing continuous high and low amplitude reflectors, discontinuous low amplitude reflectors and a reflection free zone.

The majority of the leveed channels interpreted in this study show an incision into the immediately underlying substrate at the base of the channel-levee complex. This erosional base is characterized by a discontinuous high amplitude reflector (D-HAR) (Figures 3B, 4B). Deptuck et al. (2003) describes this erosional base as an ‘erosional fairway’.
The cause for this high amplitude reflector is the result of the deposition of coarse clastic material along the channel bottom (McHargue, 1986; Mutti and Normark 1991; Amir and Kenyon 1993; Posamentier and Kolla, 2003). The shape of these erosional bases tended to be more U shaped.

In this study channel-fill high amplitude reflectors (CF-HAR) refer to similar type of deposits as the high amplitude discontinuous deposits (H-D) in (McHargue and Webb, 1986), ‘channel-fill’ elements described in Mutti and Normark (1991), and discontinuous and chaotic reflections of Deptuck et al., (2003) (Figure 3B). McHargue and Webb (1986) and Deptuck et al., (2003) suggest that these high amplitude, discontinuous and sometimes chaotic amplitude reflectors correspond to the thalweg (axial) deposits of channels. For this study, the channel-fill high amplitude reflectors are not to be confused with the discontinuous high amplitude reflectors. Channel-fill high amplitude reflectors are found above or adjacent to the edge of the basal erosion surface. Channel-fill high amplitude reflectors are important in showing the lateral and/or vertical migration of a channel.

Continuous low amplitude reflectors (C-LARs), discontinuous low amplitude reflectors (D-LARs) and reflection free zones are three seismic facies not characterized by high amplitude reflectors. Continuous low amplitude reflectors and reflection free zones are typically found in the outer levee component of the channel-levee complex (Figure 7B). Kolla and Coumes (1987) refer to the reflection free areas of the outer levees as homogeneous sediments. This supports McHargue (1991) that high amplitude reflectors in general result from a high contrast in the velocities relative to adjacent layers. Continuous low amplitude reflectors typically represent deposition of alternating layers of fine (silty) and very fine-grained (clayey sediments). Like the continuous high amplitude reflectors identified in the outer levees, continuous low amplitude reflectors also tend to thin out further from the channel. Discontinuous low amplitude reflectors were most often identified in the inner levees and stratigraphically just above channel-fill high amplitude reflectors (Figure 7B). McHargue (1991) and Deptuck et al. (2003) interpret them to form from deposition of overbank deposits or sculpting and slumping caused by erosion. The discontinuous low amplitude reflectors identified to be just above channel-fill high amplitude reflectors are potentially due to the deposition of waning turbidity flows. The homogeneous mud to very fine sand deposits would help explain the low amplitude reflectors due to a lack in velocity contrast between layers.

Table 1. The abbreviations and descriptions for the different type of seismic reflectors observed in this study.

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
</tr>
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<tr>
<td>CF-HAR</td>
<td>Channel-fill high amplitude reflectors</td>
</tr>
<tr>
<td>C-HAR</td>
<td>Continuous high amplitude reflectors</td>
</tr>
<tr>
<td>D-HAR</td>
<td>Discontinuous high amplitude reflectors</td>
</tr>
<tr>
<td>C-LAR</td>
<td>Continuous low amplitude reflectors</td>
</tr>
<tr>
<td>D-LAR</td>
<td>Discontinuous high amplitude reflectors</td>
</tr>
</tbody>
</table>
METHODS

CHANNEL-LEVEE SEISMIC INTERPRETATION

24 individual channel levee systems have been interpreted throughout the 2D seismic data set used for this study using seismic stratigraphic techniques (Kolla and Coumes, 1987; McHargue et al., 1986; Posamentier and Kolla, 2003) based on seismic facies reflection characteristics. Using the continuous high amplitude reflectors identified in Figure 7 a base and top for each of the twenty four channel-levee deposits was interpreted. Interpretations started at the base of the profile and gradually moved upward to shallower depths. A single channel high amplitude continuous reflector was typically able to be followed that onlapped the previous channel-levee system. These channel-levee systems were mapped throughout the entire survey in order to identify changes in channel-levee morphology, size and frequency. Changes in these factors may reflect variations in sedimentation rates over time.

Figure 8. A) Seismic line 209 showing an interpreted base and top horizon for individual channel-levee systems in a NE-SW direction. B) Seismic line 128 showing an interpreted base and top horizon for individual channel-levee systems in a NW-SE direction. IHS Kingdom version 8.8 was used to perform interpretations.
DEPTH CONVERSION

Before performing a two-dimensional decompaction of the data, the interpreted seismic profiles in two-way traveltime were first converted to depth. This model allowed for the depth conversion for all four seismic lines. Once depth converted a sediment budget was then able to be calculated. Calculation of a sediment budget was done by determining the thicknesses of individual layers and constrained by the amount of time it took for each layer to be deposited. The velocity vs. depth model shown below in Figure 9 was created using the interval velocities that were calculated from the original stacking velocities (Figure 10). These plots were created every 4 km across the length of the profile. By lining up the profiles side by side a contour was able to be drawn from one profile to another at an interval of every 0.5 km/s. The space between any two contours represents the average velocity at a given depth. The time-depth conversion is prone to errors because uncertainties in the interval seismic velocities may vary by as much as 20% across the length of the profile.

Figure 9. Velocity vs Depth profile for line 33.
DECOMPACATION

The unloading and decompaction of two-dimensional seismic data allows one to more accurately determine the volume of sediment deposited during designated time intervals. This is done by the removal of the overlying layers for each interpreted stratigraphic package in order to correct for the effects of burial which returns each stratigraphic package to its original thickness at the end of deposition. Over time as more and more sediment is deposited and then buried the stratigraphically lower sediment becomes more compacted chiefly due to dewatering (Clift et al., 2002).

In this study FLEX-DECOMP™ software package was used in the decompaction and unloading of interpreted seismic sections. Sclater and Christie (1980) model for sediment dewatering was applied. The parameters used to account for compaction are porosity, decompaction constant and density of sediment. Porosity values were determined by using the porosity vs. depth curve from Sclater and Christie (1980). That curve is calibrated and matches with regional porosity vs. density and porosity vs. depth information (Bachman and Hamilton 1976; Velde; 1996; Clift et al., 2002). The decompaction constant controls the exponential decrease in the rate of sediment compaction with

![Figure 10. Two examples of the velocity vs. depth plots that were used in creating figure 9.](image-url)
increasing burial. This constant can change with variation in lithologies so a value of 0.39 (1/km) for shaley sand was used. A lithology of shaley sand was used due to a lack of specific lithologic data but the assumption was made because shaley sand was the most common lithology drilled at DSDP Site 222 (Figure 1). Lastly, a density of 2.68 g/cm³ was used based on the previous assumption.

Age dates were used from the nearby Pak G2 exploration well (Figure 2), and interpretations from Carmichael et al. (2009) (Table 2). Figures 11, 12 and 13 show how the interpretations for specific age dates tie together from the borehole out to the two intersecting lines. Four different seismic profiles were depth converted and decompacted and two examples are shown in figures 14 and 15. The reason for choosing lines 33 and 37 was due to the availability of stacking velocities for these lines. Using the stacking velocities rather than a model would conceivably help to eliminate some error in the depth conversion process. Seismic lines 210 and 2065 were chosen for their higher confidence interpretations of the age picks and that they both interest lines 33 and 37. Seismic lines 33 and 37 which are approximately 80 km in length run along strike to the continental shelf. Seismic lines 210 and 2065 which are approximately 100 km in length running perpendicular to the continental shelf while intersecting both seismic lines 33 and 37.

Table 2. Age picks determined from the Pak G-2 well (Figure 11).

<table>
<thead>
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<th>TWT (sec)</th>
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<td>2862</td>
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</tr>
<tr>
<td>C</td>
<td>3</td>
<td>3120</td>
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</tr>
<tr>
<td>D</td>
<td>5</td>
<td>3372</td>
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<td>I</td>
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<td>5874</td>
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</table>

Figure 11. Interpreted seismic profile for line 31. The vertical red lines represent the intersections of lines 2065 and 210. The well path of the Pak G-2 well is represented by the black vertical line.
Figure 12. Interpreted seismic profile for line 2065. The vertical red lines represent the intersections of line 31.

Figure 13. Interpreted seismic profile for line 210. The vertical red lines represent the intersections of line 31.
Figure 14. Table showing ages for each layer. Location map of the four seismic lines that were depth converted and used for calculating sediment budget. Figure A, seismic profile 33 with interpretations for each of the ten layers. Figure B, depth converted profile of seismic line 33 from FLEX-DECOMP™.
SEDIMENT BUDGET CALCULATION

The four seismic lines create an enclosed box with an area of 700 km$^2$, highlighted in yellow on figures 14 and 15. First step was to determine the thickness of each of the 10 layers, for lines 33, 37, 210 and 2065, once the layers had been decompacted by removing the above layers. An average thickness for each layer was calculated by recording the thickness of the layer at an interval of every 5 km across the profile. This process was done for each of the 10 layers on all four profiles. Now that an average thickness for each layer had been determined for each profile, an average thickness needed to be calculated that represented the volume of the enclosed box.
A volume was calculated by averaging the four thickness values from each profile for a single layer and multiplying it by the area which results in a volume. That step was repeated for all 10 layers. The volume of the three dimensional ‘box’ that is enclosed by the four seismic lines was determined. The calculated volume of sediment within the boundaries of the four seismic lines equaled 5830 km$^3$. That volume was extrapolated out to be representative of the entire 4.5$^6$ km$^3$ Indus Fan. The Indus Fan has a total volume of about 4.5$^6$ km$^3$ so the volume calculated from the study area only represents 0.12 % of the entire Indus Fan volume. This introduced some error by having such a relatively small volume be representative of something so large, for example not being able to account for thickening and thinning out beyond the seismic lines. However this method does provide a more accurate sediment budget than using single core data sets. The sedimentation rate for each layer was determined by dividing the volume of the layer by the duration of sedimentation.
RESULTS

CHANNEL-LEVEE COMPLEX 1

Channel-levee complex 1 seen in figure 16 is used as an example because it shows a quality example of channel-levee system abandonment, migration and activation in a new location. It shows a well-defined sequence of how multiple systems can build and stack on top of older deposits. The location of channel-levee complex (CL) 1 is on seismic line 119 (Figure 17) and extends over the length of the survey. Take note that it is located stratigraphically deeper than any of the other three channel-levee complexes. Using figure 17 at the depth of channel-levee complex 1 very few complexes are beneath it supporting the idea that sedimentation may have been relatively low around that time. This line is the furthest northeastern seismic line for this 2D survey, about 45 km from the continental shelf. Each system is labeled CL 1-3. CL 1 being the oldest and deposited first while CL-3 is the youngest. Determination of which system was deposited first was done by interpreting the first high amplitude reflector that was continuous across the top of CL-1. This reflector was interpreted to be below the base of the channel-levee system of CL-2. In order to confirm that CL-2 was deposited after CL-1 I followed the first high amplitude reflector that was continuous across the top of CL-2 which onlapped on to CL-1. With the assumption that only a single channel-levee system can be active at a time CL-1 was deposited and abandoned before the deposition of CL-2 occurred. All three of the channel-levee systems are depositional according to the classification method used by Normark (1970) and Mutti and Normark (1991). The absence of an erosional base below the outer levees was the first indication of a depositional system. CL-1’s channel edge has vertically aggregated through time with no horizontal migration. After the avulsion and abandonment of CL-1, CL-2 became active about 15 km to the southeast. CL-2’s channel history is slightly different than CL-1 in that the channel slightly migrates to the northwest while also vertically aggrading.

Evidence of this horizontal and vertical movement of the channel can be seen by the interpretation of the previous channel edges that are slightly sub parallel to one another. Homogenous, fine grained sediment is deposited as active channel-fill as the channel migrates vertically and to the northwest. This is indicated by the discontinuous low amplitude reflectors just inside the right inner levee. Like CL-1 the channel-fill high amplitude reflectors bounded by the most recent channel edge are indicative of coarse-grained sediment. CL-3 has a slightly different channel shape than the previous two channel-levee systems. The channel shape of CL-3 is much more v shaped with a deeper thalweg. This v shaped channel may have been caused by turbidity currents with higher velocities or increased sediment loads. Consistent with the previous two channel-levee systems, channel-fill high amplitude reflectors were deposited along the channel-axis.
The location of channel-levee complex 2 is northwest of the previous channel-levee complex on seismic line 119 (Figures 17, 18). It too is mapped over the length of the survey.

**CHANNEL-LEVEE COMPLEX 2**

The location of channel-levee complex 2 is northwest of the previous channel-levee complex on seismic line 119 (Figures 17, 18). It too is mapped over the length of the survey.
Channel-levee complex 2 sits stratigraphically higher and in the midst of what appears to be a very active period of time for sedimentation. This theory is supported by the numerous other channel-levee complexes below, above and on either side of channel-levee complex 2. Reason for this example is to show how a channel axis will aggrade supported by evidence of shingled channels edges and vertically stacked channel-fill high amplitude reflectors. Channel-levee complex 2 is combination of both erosional and depositional (Normark, 1970; Mutti and Normark, 1991). Initially the channel complex was erosional indicated by the discontinuous high amplitude reflectors and erosional base interpreted just below the outer levees (Figure 18). The chaotic and discontinuous channel-fill high amplitude reflectors represent a channel migration and aggradation to the southeast. Further support of this channel migration is the identification of the sub parallel channel edges that also move in the same general direction. Discontinuous low amplitude reflectors are visible on the lower section of the right inner levee. Causes for those types of reflectors may be from erosion and slump failure of the inner levee as the channel migrated further to the southeast.

Figure 18. Channel-levee system 2. 2A) uninterpreted seismic cross section. 2B) interpreted seismic cross section shown in 2A indicating the seismic facies, inner levees, outer levees, channel edges, layered overbank levee deposits and an erosional base.
At about 4.85 seconds channel migration switches direction towards the northwest indicated by the channel edge interpretations and the better stratified inner levee on the right hand side. The large reflection free zones on the outer levees, lack of numerous continuous high amplitude reflectors and the abundance of discontinuous low amplitude reflectors in the inner levee supports my interpretation that very little coarse-grained sediment was deposited outside the banks of this channel-complex.

CHANNEL-LEVEE COMPLEX 3

The location of channel-levee-complex 3 extends over many profiles and intersects seismic line 128 approximately 50 km southwest of the channel-complexes 1 and 2 and about 100 km from the continental shelf (Figures 19, 20). This example is used to show a channel-levee complex further from the shelf that initiated as highly erosional and cut deeply into the underlying strata. Channel-complex 3 is also a combination of both and erosional and depositional system (Normark, 1970; Mutti and Normark, 1991). High amounts of erosion occur at the initiation of this channel-levee complex. The erosional base interpreted from the discontinuous high amplitude reflectors located well below the base of the outer levees is about 2 km in width at its widest points. Erosion occurred about two tenths of second down into the levee and channel fill of an older channel-levee complex. Although the outer levee on the left hand side of CL-3 seems to conform to the channel-levee complex below (CL-1), it is indeed two separate complexes because of the observed onlap of CL-3 on top of CL-2 on the right hand side of the cross section. The onlap of the second oldest channel levee complex (CL-2) at about 4.6 seconds onto CL-1 supports that there was a previously active channel-levee complex prior to CL-3. As the channel aggrades and migrates in a northwestern direction notice the lack of inner levee on the left side. The inner levee has been entirely eroded away. Two separate rectangular shaped ‘terraces’ or inner levees are visible on the right hand side. Sharp erosional contacts help to separate the two (shown in yellow). The inner levees appear to have slumped down as the channel migrated to the northwest. The upper most interpreted channel edge shows a combination of channel-fill high amplitude reflectors and discontinuous low amplitude reflectors. Notice that the more coarse-grained sediment indicated by channel-fill high amplitude reflectors are located closer to the base of the channel edge and the finer-grained sediment or discontinuous low amplitude reflectors lie just above the channel-fill high amplitude reflectors. The discontinuous low amplitude reflectors may have been sediment deposits of the waning stages of a turbidity flow. Above the discontinuous low amplitude reflectors deposition of passive channel fill occurred; which takes place after the abandonment of channel-levee complex 3.
Figure 19. Channel-levee system 3. 3A) uninterpreted seismic cross section. 3B) interpreted seismic cross section shown in 3A indicating the seismic facies, inner levees, outer levees, channel edges, layered overbank levee deposits and an erosional base.

Figure 20. Uninterpreted seismic profile for line 128. The box outlines the location of channel-levee complex (CLC) 3.
CHANNEL-LEVEE COMPLEX 4

The location of channel-levee complex 4 is on seismic line 119 northwest of channel-complexes 1 and 2 (Figures 17, 21). This complex is used to show deposition at time closer to present and less affected by burial and compaction. It also shows just how complex the morphology can get over the life span of the complex by showing interpreted periods of activation, abandonment and reactivation all inside the same outer levees. I interpreted channel-levee complex 4 as a combination of both erosional and depositional (Normark, 1970; Mutti and Normark, 1991). Channel-levee complex 4 is the largest of the interpreted complexes. A lack of overburden and high sedimentation rates may account for its larger levees and more complex channel history. Three separate depositional/erosional events took place over the life span of this complex. They are numbered 1-3, 1 being the oldest and 3 being the youngest. A large wide erosional u shaped erosional base, with a width of almost 2.5 km, marks the initiation of the channel-levee complex with discontinuous high amplitude reflectors barely below the bottom of the outer levees. Like the previous two complexes the channel transitions from erosional to depositional and aggrades both vertically and horizontally in the northwest direction. Notice the large inner levee on the right side compared to the inner levee on the left despite the lateral migration towards the northwest. At about 3.8-3.9 seconds I have interpreted a relatively continuous high amplitude reflector (in green) that only partially extends across from one outer levee to the other. I interpreted this to be deposition of overbank deposits or hemipelagic drape while channel-levee complex 4 avulsed or temporarily abandoned. The interpreted high amplitude reflector shown in green may correspond to the unusually thick package at about 4.1-4.2 seconds located on the left outer levee (shown in purple). After an unknown amount of time reactivation of channel-levee complex 4 takes place where channel migration and vertical aggradation continues toward the northwest. Channel edges were able to be interpreted to help better show channel migration both laterally and vertically. Notice how above the green high amplitude reflector the inner levee on the right side is almost completely eroded while the inner levee on the right side is very well stratified and extend further out away from the outer levee. Before this channel-levee complex was abandoned for good, one last transition to an erosional setting took place indicated by the large u shaped erosional base (3). Again the presence of passive channel fill overlies the abandoned channels.
SEDIMENT BUDGET

A sediment budget was able to be determined for ten different layers that span from 65 Ma to recent (Figure 22). According to Royer et al. (1992) and Miles et al. (1998) continental break up began around 65 Ma allowing for the commencement of deposition. A relatively low rate extends from the mid Paleocene to the early Oligocene. An overall increase in sedimentation rates occurs up until the late Miocene to Early Pliocene where the maximum rate is reached. This maximum rate is more than double the amount during the late Oligocene to early Miocene. Beginning near the early Pliocene a steady decrease in rates continues into the Pleistocene and to recent. Mass uncertainties of up to 20% were attributed to uncertainties in the velocity-depth conversion.

Figure 21. Channel-levee system 4. 4A) uninterpreted seismic cross section. 4B) interpreted seismic cross section shown in 4A indicating the seismic facies, inner levees, outer levees, channel edges, layered overbank levee deposits and an erosional base.
Figure 22. Sediment budget for the study area portion of the Indus Fan from 65 Ma to present. A table showing the results of previous studies for comparison.

<table>
<thead>
<tr>
<th>Author</th>
<th>Peak Sedimentation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rea (1992)</td>
<td>2-4 Ma, 6-9 Ma</td>
</tr>
<tr>
<td>Burbank et al. (1993)</td>
<td>Before 8 Ma</td>
</tr>
<tr>
<td>Métivier et al. (1999)</td>
<td>2-3 Ma</td>
</tr>
<tr>
<td>Clift and Gaedickel (2002)</td>
<td>16-11 Ma</td>
</tr>
<tr>
<td>Clift et al. (2006)</td>
<td>2 Ma</td>
</tr>
</tbody>
</table>
DISCUSSION

CONTROLS ON EROSION

The three main factors affecting erosion in the western Himalayan and Karakoram Mountains are tectonics, glaciation and monsoon strengthening (Burbank et al., 1993; Prins and Postma 2000; Galy and France-Lanord 2001; Clift et al., 2002). Each of these three has played a role since mountain building initiation began near 50 Ma. This section of the thesis will look at how each of these factors helped shape the relatively high sedimentation rates of the Indus Fan during the mid to late Miocene.

The relationship between tectonics and climate change has been discussed in many papers (Molnar and England 1990, Rea 1992, Zheng et al., 2001). In the Himalaya as chemical weathering increased the draw-down of atmospheric CO$_2$ gets deposited in limestone which causes a reduction in this greenhouse gas and results in global cooling (Raymo et al. 1988). This in turn could lead to the onset of glaciation which can increase erosion. The widespread glaciation that began around 3 Ma may be another contributor for the higher rates because glacial erosion can be higher than fluvial erosion (Hallet et al, 1996). Zhang et al. (2001) suggests that variation in climates such as the transition from a wet and warm climate to one that is cold and dry, prevents fluvial and glacial systems from establishing equilibrium states thus leading to higher erosion rates. On the other hand in the case of the Tibetan Plateau summer heating of the air above it can cause the strengthening of the monsoon, which leads to higher winds and increased rainfall (Manabe and Terpstra, 1974; Prell and Kutzbach 1991 and Molnar et al., 1993). An uplifted Tibetan Plateau has both physical and chemical effects on the nearby region. During the summer a source of heating in the lower atmosphere occurs due to the uplifted plateau that creates a vast, low-pressure system over central Asia by drawing in humid and warm air from the Indian Ocean towards the Plateau (Webster, 1987).

SEDIMENT BUDGET COMPARISON

Rea (1992) study shows relative flux maxima at 2-4 Ma and from 6-9 Ma. He relates those high rates during those two periods on two different stages of rapid uplift of the Himalayan Mountains. Burbank et al. (1993) shows decreasing rates after 8 Ma in the nearby Bengal Fan despite monsoon intensification. He attributes the inverse relationship between monsoon intensification and sedimentation rates to three hypotheses. Those hypotheses include a decline in uplift rates, strengthening monsoon intensities associated with decreased glaciation which lead to decreased mechanical weathering and the stabilization of slopes due to increased vegetation growth. Métivier et al. (1999) budget showed low rates during the Cenozoic with a large increase in the Pleistocene. The difference between Métivier et al (1999) and this study could be attributed to the location differences. While this study was located off the shelf on the upper fan, Métivier was located on the shelf where the subsidence rate is to slow to accommodate all the sediment transported by the Indus River.
Clift and Gaedicke (2002) results are from using multi-Channel seismic reflection data showed a peak in sedimentation during the middle Miocene (16-11 Ma). The more recent Clift et al. (2006) study used a 2D seismic survey instead of drill sites which altered the previous budget by showing a second increase in sedimentation rates during the Pleistocene (2 Ma).

Table 3. Table showing each study, time period they concluded to be peak sedimentation and the reason behind their results.

<table>
<thead>
<tr>
<th>Author</th>
<th>Peak Sedimentation</th>
<th>Cause for Results</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rea (1992)</td>
<td>2-4 Ma, 6-9 Ma</td>
<td>Two periods of rapid Himalayan uplift</td>
</tr>
<tr>
<td>Burbank et al. (1993)</td>
<td>Before 8 Ma</td>
<td>Decreased uplift rates, mechanical weathering, and increased stabilization of slopes</td>
</tr>
<tr>
<td>Métivier et al. (1999)</td>
<td>2-3 Ma</td>
<td>Extrusion and Crustal shortening</td>
</tr>
<tr>
<td>Clift and Gaedicke (2002)</td>
<td>16-11 Ma</td>
<td>Uplift in source region, monsoon strengthening and glaciation</td>
</tr>
<tr>
<td>Clift et al. (2006)</td>
<td>2 Ma</td>
<td>Rock uplift and strengthening of Asian Monsoons</td>
</tr>
</tbody>
</table>

**SEDIMENT BUDGET INTERPRETATION**

After review of the different factors effecting erosion and the previous work one can conclude that various factors have to be considered when trying to determine the cause for changes in sedimentation rate over time. Biases can occur in a study due to type of data set whether it's from a drill site or seismic data or location (foreland basin, shelf or submarine fan).

The peak sedimentation rates observed in Figure 20 during the late Miocene and early Pliocene are representative of a combination of tectonic uplift, increased strengthening of the Asian monsoons, global cooling and relatively lower sea levels. It is unanimously believed that the altitude of the southern Tibetan Plateau has not significantly changed since about 15 Ma and plays a significant role in the intensity of the Asian monsoons (Kutzbach et al., 1989; An et al., 2001; Harris 2006). Previous studies suggest an active monsoon around 11-16 Ma and a renewed strengthening around 2.6-3.6 Ma (An et al, 2001; Clift et al., 2002). The more recent strengthening of the East Asian monsoon has been associated with uplift of the northern plateau by Li and Fang (1999) and Zheng (2000). On a regional scale the elevated terrain provides potential energy to important factors of erosion, rivers and glaciers. Globally Zheng et al. (2001) presented evidence supporting increased sedimentation rates from 2-4 Ma being attributed to global cooling and the onset of glaciation that was proposed by Hallet et al, (1996). With evidence supporting the possibility of relatively high erosion rates during the late Miocene and early Pliocene the only factor lacking is the ability for the sediment to bypass deposition on the shelf and be deposited on the abyssal plain.

Global sea level changes have a large impact on the sediment input to the oceans. It is understood that times of rising and high sea level generally relate to periods of reduced sediment transport off the continental shelf to the deep sea. Opposed to times of falling, rapidly changing or low sea level relate to periods of increased input of sediment to the deep sea. Relative sea level records
determined by Haq et al. (1987) and Greenlee and Moore (1988) can be seen in Figure 23. Higher sea levels in the middle Miocene with a distinct lowstand at 10 Ma can be observed. Following the lowstand at 10 Ma a broad highstand between 9 and 6 Ma is followed by rapidly changing sea levels between 6.5 and 3.5 Ma. When comparing the Greenlee and Moore (1988) record due to its higher detail, to the calculated sedimentation rate a correlation between low or fluctuating sea levels and increasing or max sedimentation rates can be assumed.

SEDIMENT SUPPLY AND CHANNEL LEVEE MORPHOLOGY

Growth of submarine fans by channel-levee systems generally occur during sea-level lowstands. Sediment is able to bypass the shelf due to river entrenchment and canyon formation in the continental shelf, accompanied with the basinward shift of deltas to the shelf edge (Posamentier and Vail, 1988; Mutti and Normark, 1991). As sedimentation rates change so does the sediment flux, velocity and frequency of turbidity flows that help transport sediment from the shelf down the slope onto the abyssal plain. The following section will show how individual channels and channel-levee complexes differ depending on sedimentation rates. Figure 24 A and B are two isochron maps of two different channel-levee systems. They correspond to the seismic cross sections in Figures 24 and 25. Both cross sections were taken from the same seismic line.

What differs between CLS-5 (A) and CLS-6 (B) is the environment in which they were deposited. CLS-5 was deposited during the early Miocene.
Just after the onset of rapid turbidite deposition (Kolla and Coumes, 1987). Due to its older age compaction from younger sediments may have decreased the vertical thickness of CLS-5. CLS-6 is younger and deposited during the late Miocene and early Pliocene. As discussed before, sedimentation rates were not similar at the time of initiation, deposition and abandonment. Characteristics including relative size, shape and architectural elements of these channel-levee complexes are due to the different depositional environments while they were active. Figure 24A is representative of a predominantly aggradational and depositional channel. The yellows and reds indicate the thickest sections which correspond to the channel axis. The transition from dark green to light and darker blues is indicative of the thinning outer levees. The smaller CLS-5 appears to have much thinner and further extending levees than CLS-6, indicated by the wide extent of the light blue color. This may be caused by the overspill of sediment after the channel has aggraded and filled up all available space between the outer levees. Take note of the decrease in reds and yellows as the channel extends further south. This supports the idea that less and less sediment is making its way out to the furthest extend of the channel. The loss of velocity or sediment load may cause the turbidity current to deposit its load prior to reaching the end of the channel mouth.

Figure 24. A) Isochron map of CLS 5. B) Isochron of CLS 6. C) Location map referencing the location of the intersection between the channel and the seismic profile shown in figure 15.
At first glance the isochron of CLS-6 shows much more reds and yellows indicating greater channel thicknesses than CLS-5. Vertically taller or thicker channel may be indicative of possible erosion into underlying older strata and/or increased deposition and aggradation of outer levees. Both isochron maps show only a slight amount of meander or sinuosity. This was common throughout all of the channels that were mapped. In CLS-6 there is much less light blue extending from the edges of the channel compared to CLS-5. The shapes of the outer levees of CLS-6 are not only thicker but they also have a much sharper wedge shape. Like CLS-5, CLS-6 may abandon its current channel as soon as its outer levees cannot contain the passing flow. Instead of thin splay deposits out away from the channel during more frequent and higher energy turbidity flows channels will avulse and initiate a new channel. This can be seen much better in figures 25 and 26. By the evidence seen on the channel-levee isochron maps sedimentation rates can be interpreted by comparing channel axis thickness, channel width, and outer levee shape.

![Figure 25](image.png)

Figure 25. Channel-levee system 5. 5A) uninterpreted seismic cross section. 5B) interpreted seismic cross section shown in 5A indicating the seismic facies, inner levees, outer levees, channel edges, layered overbank levee deposits.
Figures 25 and 26 help to show the differences in channel-levee architecture such as channel width and depth, channel-fill type, and outer levee size and internal structure when sedimentation rates differ. Looking first at figure 24 channel-levee system 5, notice the little to almost no amount of erosion occurring below the base of the outer levees. According to the classification system of Normark (1970) and Mutti and Normark (1991) this channel-levee system could be classified as either depositional or slightly a mixture between erosional and depositional. Due to the low amount of sinuosity very little lateral channel migration can be seen. That is indicated by the left to right movement of the interpreted channel edge. Another characteristic of a low sedimentation rate is the lack of mid to high reflectors visible in the outer levees. The continuous low amplitude reflectors and reflection free zone supports the concepts of very fine to fine-grained sediment such as muds and clays. That would support that during times of lower sedimentation, due to low sediment transport by the Indus River, turbidity flows have less energy thus unable to continue erosion and carry more coarse-grained sediment down fan.

![Channel-levee system 6. 6A) uninterpreted seismic cross section. 6B) interpreted seismic cross section shown in 6A indicating the seismic facies, inner levees, outer levees, channel edges, layered overbank levee deposits and an erosional base.](image)

Figure 26. Channel-levee system 6. 6A) uninterpreted seismic cross section. 6B) interpreted seismic cross section shown in 6A indicating the seismic facies, inner levees, outer levees, channel edges, layered overbank levee deposits and an erosional base.
The following equation from White (1940) helps to show how different factors affect the necessary shear stress required to initiate and keep grains in motion.

\[ T_{t0} = ac(\rho_s - \rho)gD \tan \Phi \]

Where \( T_{t0} \) is the bed shear stress, \((\rho_s - \rho)\) is the density difference between the particle and the fluid. In this case a value of 2.65 g/cm\(^3\) for quartz sand could be used for \( \rho_s \). \( D \) is the grain diameter which can vary due to the proximity to the coast. More coarse grains will fall out of suspension quicker than finer sediment due to a loss of energy as the turbidite moves further offshore. The packing coefficient is expressed as \( c \) and \( \tan \Phi \) is the angle of repose. If grain diameter, density and packing coefficient increase along with a higher angle of repose (affected by grain shape) increase, a larger bed shear stress is required.

Figure 26 showing a seismic cross section of the younger and larger CLS-6. Starting at the base of the channel-levee complex a deep erosional base scoured down into the underlying deposits. Notice the large width and depth of the erosional base allowing for higher amounts of higher energy flows to move down fan. In order for erosion to occur the turbidity flow must have enough energy to break the bed shear stress and cause the motion of grains along the channel bed. Erosional channels are much more indicative of times of increased sediment flux because an increase in sediment flux can be related to periods of increased velocity resulting in erosion (Nummendal and Prior, 1981). Interpreting the morphology of CLS-6 is a bit more complicated than CLS-5. After the initiation and erosion of the channel vertical aggradation takes place along with the buildup of the outer levees. The interpreted continuous high amplitude reflector in the outer levee may correspond to a period of time where large amounts of more coarse-grained (typically channel-fill) sediment breached the levee walls and were deposited. The cause for the high amplitude reflector is due to a contrast in velocities between the high amplitude reflector and the sediment that lies directly above and below. For example, in this case the high amplitude reflector being medium to coarse-grained sand deposited during a period of high energy and frequency of turbidity flows overlain by very fine to fine-grained sand deposited once turbidite flow energy subsided. Evidence of slight lateral channel migration is shown by the left shift in the interpreted channel edges. The large reflection free zone that appears to be in between two different channels are the remnants of an outer levee that was eroded into by a younger channel-levee system. Figure 26 helps to support that sediment flux through the channel was higher during the time the channel was active based on the multiple erosional periods, the larger outer levees, wider channels, and increased frequency of channel activation.
CONCLUSION

This study used multiple data sets to interpret the change in sediment budget of the Indus Fan over the course of the Cenozoic. Using the techniques of depth conversion and decompaction, volumes and sedimentation rates were calculated and extrapolated to see what the trends would be if these rates applied to the entire submarine fan. Although some error was brought in due to the necessary interpretation of the age picks from the exploration well Pak-G2 and from not being able to account for the thickening and thinning of the interpreted layers outside the four seismic profiles, results supported by previous work were compiled. The results suggested a peak in sedimentation rate during the late Miocene and early Pliocene. The increase in sedimentation rates are supported by increased tectonic activity of the northern Tibetan Plateau. This increase in uplift allowed for greater erosion not only from increased relief but from the strengthening of the Asian Monsoons. Further conditions that suggest conditions for higher erosion was the onset of global cooling that began around 3 Ma. The climate change did not allow for fluvial and glacial systems to reach equilibrium. This increase in sediment flux allowed for the transportation of sediment from the source to the continental shelf. Sediment was then able to reach out beyond the shelf due to relatively quickly fluctuating sea level changes from 3-5 Ma.

24 separate channel-levee complexes were able to be interpreted throughout the 2D seismic data set. The interpretation and breakdown of the multiple architectural elements that make up a channel-levee complex aided in the ability to reconstruct the morphology of multiple complexes. With isochron mapping and further cross section interpretations, characteristics such as increased channel axis and levee thicknesses able to be related to periods of either relatively high periods of sedimentation like during the Neogene or low periods of sedimentation that occurred during the early Paleogene.
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Figure A. Interpreted seismic profile for line 123. 24 separate channel-levee complexes are interpreted.
Figure A. Channel-levee system 1

Figure B. Channel-levee system 2
Figure C. Channel-levee system 3

Figure D. Channel-levee system 4
Figure E. Channel-levee system 5

Figure F. Channel-levee system 6
Figure G. Channel-levee system 7

Figure H. Channel-levee system 8
Figure I. Channel-levee system 9

Figure J. Channel-levee system 10
Figure K. Channel-levee system 11

Figure L. Channel-levee system 12
Figure M. Channel-levee system 13

Figure N. Channel-levee system 14
Figure O. Channel-levee system 15

Figure P. Channel-levee system 16
Figure Q. Channel-levee system 17

Figure R. Channel-levee system 18
Figure S. Channel-levee system 19

Figure T. Channel-levee system 20
Figure U. Channel-levee system 21

Figure V. Channel-levee system 22
Figure W. Channel-levee system 23

Figure X. Channel-levee system 24
Figure Y. Outline of all 24 channel-levee systems.
VITA

Taylor Berlin is from Tomball, Texas. He attended Hardin-Simmons University from 2007 to 2011 and received a Bachelor of Science in Geology and a minor in Environmental Science in 2011. He began work toward a Master of Science in Geology at Louisiana State University in the fall of 2011.