Sediment delivery from the Fly River tidally dominated delta to the nearshore marine environment and the impact of El Niño

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[1] In tidally dominated deltas, sediment discharged from the river sources must transit through an estuarine region located within the distributary channels, where particle pathways can undergo significant transformations. Water column profiles and time series data were collected in the distributaries and shallow nearshore region offshore of the Fly River, Papua New Guinea, during monsoon conditions of 2003 and 2004. Within the distributaries of this tidally dominated delta, near-bed fluid mud concentrations were observed at the estuarine turbidity maximum. Discharge and morphology of the distributaries determine the location of estuarine processes and control the sediment flux to the marine environment. Sediment delivery from the delta to the broad shallow nearshore region northeast of the river mouth and to the steep Umuda shelf valley is influenced by the shorter pathway of the northernmost distributary. The shallowest portion of the nearshore region (<10 m water depth) is a zone of temporary storage where unconsolidated sediment is easily resuspended. Umuda shelf valley constrains and enhances tidal currents and provides a steep conduit to seaward transport of fluid muds. Comparison of 2003 and 2004 data shows evidence of reduction in freshwater supply to the Gulf of Papua during the moderate 2003 El Niño conditions. El Niño creates a large negative perturbation (i.e., low flow) to the relatively constant sediment discharge. This reduction of discharge limits transport of sediment from the distributaries to the nearshore zone of temporary storage. As the sediment stored near shore feeds the prograding clinoform found seaward, the perturbation propagates throughout the dispersal system.


1. Introduction

[2] Sediment discharged from rivers to the marine environment must transit through an estuarine region where fresh and saline water mix, and dramatic changes in transport mechanisms and particle packaging occur. These processes influence the movement of particles into, through, and out of the inner shelf, and ultimately impact construction of the continental shelf. In deltas, the estuarine processes may be located within the distributary channels during some periods of the tidal cycle. Also, in this case, the morphology of the deltaic distribution system and adjacent shallow water environment impacts the pathway sediment takes on its journey to deposition in shelf regions.

[3] Sediment that makes it through the deltaic gateway to the inner shelf is subject to intense water column and seabed reworking by processes that vary tidally, seasonally and over longer timescales. Rhythmic tidal processes within deltaic systems should lead to predictable changes in temporal deposition on the shelf. Yet seasonal and other subtidal processes can modify the impact of the tides, and coupling of these processes creates spatial and temporal variability in the transport and deposition of sediment. As an example, on the inner shelf off the Amazon River, estuarine processes created very large inventories of suspended sediment (i.e., fluid mud >10 g L^{-1}) that varied seasonally, as a result of changes in river discharge [Kineke et al., 1996] and intensity of reworking in shallow water [Kuehl et al., 1995]. Notably, the Amazon system has been studied from an interdisciplinary perspective [e.g., Aller et al., 1996; Nittrouer and DeMaster, 1996], and the sediment transport mechanisms in the estuarine and inner shelf regions were found to be critical for processing terrestrial and marine sedimentary components (e.g., organic carbon) and creating a major sedimentary feature on the shelf (i.e., clinoform deposit).

[4] In contrast to the river-dominated inner shelf of the Amazon system, where seawater does not penetrate into the
river channel, the Fly River discharges to the Gulf of Papua (Figure 1) through a tidally dominated delta. The convergent processes and flocculation typical of estuarine systems occur within the confined delta distributaries rather than on the open expanse of the continental shelf. Because tidally dominated deltas occur in many regions of the world (e.g., Ganges-Brahmaputra, Bangladesh; Ord, Australia), examination of sediment transfer through them is critical to understanding how a common type of sedimentary interface controls dispersal from fluvial source to marine sink.

This study examines sediment transport processes in the tidally dominated Fly River delta and adjacent inner shelf, which is the landward segment of an actively prograding clinoform structure [Harris et al., 1993; Walsh et al., 2004; Slingerland et al., 2008a]. This is a large-scale feature that builds upward and seaward on the continental shelf, and its presence confirms that sediment is escaping the Fly delta. The mechanisms delivering sediment from the delta to the shallow region of the Gulf of Papua have received little attention. We provide observations over tidal, seasonal, and interannual (El Niño and non-El Niño) timescales in a system that has had little anthropogenic impact. These observations help define the pathways of sediment dispersal, and aid the interpretation of shelf sediment accumulation patterns. This study contributes to the ultimate goal of the NSF MARGINS Initiative in the Gulf of Papua: the development of a quantitative understanding of processes that control the discharge, grain size, and delivery rate of sediment from the fluvial sources to marine sinks.

The specific objectives of this study are (1) to document water properties and suspended-sediment concentrations in the Fly River delta, and on the shallow topset region of the adjacent modern clinoform structure; (2) to investigate the influence of physical forcing mechanisms that control the sediment resuspension and transport through the delta and into the shallow nearshore; and (3) to compare the variability of sedimentary processes associated with the monsoon seasons during El Niño and non-El Niño conditions.

2. Background

2.1. Tidal Deltas and Estuarine Processes

Tidally dominated deltas are characterized by low-sinuosity distributary channels where water is stored and released with the tidal cycle. In deltas of this type, tidal flows are stronger than river flows. Where the ratio of tidal
amplitude to channel depth is high, the tide behaves as a finite amplitude wave with stronger flow than ebb velocities [Wright, 1977]. The distributaries are separated by elongate delta islands, and along-channel bed forms within the channels often guide the flood and ebb flows [e.g., Harris et al., 2004]. The Fly River delta has been used as an end-member example in delta classification schemes for tidally dominated deltas [e.g., Wright, 1985].

[8] Processes controlling the dispersal of sediment through the delta complex include those associated with the estuary. The zone of estuarine mixing in tidally dominated deltas is typically confined within the distributary channels for at least some parts of the tidal cycle under normal river flow conditions [Elliott, 1986]. Fresh water is delivered from a confined channel environment with relatively high velocities, to a zone where the channel cross-sectional area expands, causing velocities to decrease. The fresh water converges with estuarine water creating a typical circulation pattern of a landward flowing near-bed layer and a seaward flowing surface layer. The relatively high sediment load and introduction of salinity are prime conditions for particle flocculation, inducing rapid settling and an estuarine turbidity maximum [e.g., Allen et al., 1980].

2.2. Morphology of the Fly River Delta and Adjacent Clinoform

[9] The Fly River separates into multiple distributaries of the tidally dominated delta at an apex located ∼100 km from the distributary mouths (Figure 1). There are three major channels separated by elongate island complexes that lead to the Gulf of Papua, named the South, North and Far North distributaries. Seaward of the distributaries, the combined discharge of the Fly River and other rivers has formed a distinct clinoform deposit that is prograding over the drowned erosional landscape of the continental shelf [Slingerland et al., 2008a]. This sigmoidal-shaped deposit [Walsh et al., 2004] is characterized by low accumulation rates on the shallow, low-gradient topset due to intense reworking by waves and currents (water depths < ∼25 m); enhanced accumulation on the deeper (∼25–60 m) and steeper foreset due to decreased bed shear stresses; and slow accumulation on the bottomset due to limited sediment supply (> ∼60 m). Clinoform deposits in the Gulf of Papua extend from the southern flank of the Fly River delta northeastward around the gulf for > ∼350 km [Harris et al., 1996; Walsh et al., 2004]. However, the rates of accumulation are spatially heterogeneous along the clinoform, especially in partially filled shelf valleys [Crockett et al., 2008b]. As an example, Umuda Valley found seaward of the northernmost, or Far North, distributary (Figure 1) has a broad, alluvially influenced morphology, which accentuates modern marine sedimentary processes [Martin et al., 2008].

2.3. Fly River Discharge and El Niño

[10] The Fly River transfers sediment from a terrestrial source that stretches from 4000-m elevations in the Western Highlands of New Guinea (see Figure 1 inset) to the marine sink in the Gulf of Papua, with a hydrograph of river flow that varies little seasonally due to the large catchment area and persistent rainfall. The water discharge averages ∼6500 m³ s⁻¹. The sediment load of the river was ∼85 × 10⁶ t a⁻¹ prior to mining activity along the Ok Tedi distributary and has increased to the present level of ∼115 × 10⁶ t a⁻¹ with the mining activity [Harris et al., 1993]. The mean monthly discharge for the period 1990–1999 at Obo, the lowermost gauging station, ∼300 km from the river mouth (Figure 2a) reveals a seasonal variation of approximately a factor of two, unlike temperate rivers where seasonal variability can be orders of magnitude [Warrick et al., 2004].

[11] The strongest fluctuations in the Fly River hydrograph result from El Niño climatic conditions in the western Pacific, which are associated with decreased rainfall and a dramatic reduction in river discharge (Figure 2b) [Dietrich et al., 1999]. The relative magnitude of El Niño conditions can be evaluated using the Southern Oscillation Index (SOI), which is calculated from the air pressure difference between Tahiti and Darwin. Sustained negative SOI values are usually accompanied by warming of the central and eastern tropical Pacific Ocean, a decrease in the strength of the Pacific Trade Winds and a reduction in rainfall over the west Pacific including northern Australia and New Guinea. A strong El Niño occurred in 1997–1998 with monthly SOI indices averaging −18.8 (Australian Bureau of Meteorology, http://www.bom.gov.au). It is hypothesized that the strongest variations in sediment discharge to reach the Gulf of Papua are not produced by seasonal increases to high sediment supply but by decreases in discharge due to reduction in outflows associated with El Niño conditions [Dietrich et al., 1999].

2.4. Circulation and Sediment Transport in the Gulf of Papua

[12] Circulation in the Gulf of Papua is influenced by the Coral Sea Current, river discharge, winds, and tides [Wolanski et al., 1995a; Hemer et al., 2004; Slingerland et al., 2008b]. The coastal current causes a general clockwise rotation and may produce a small counterclockwise eddy in the northern portion of the gulf, although observed low-frequency currents are weak [Wolanski et al., 1995a; Slingerland et al., 2008b]. Freshwater discharge causes salinity stratification throughout much of the Gulf of Papua with the plume extending 30–70 km from the shore, depending on the wind conditions. The flux of buoyancy from the rivers contributes to seaward surface flows and landward bottom flows in the coastal regions to ∼30 km from shore [Wolanski et al., 1995a; Slingerland et al., 2008b]. On the inner shelf, tidal currents have a mixed semidiurnal signal with a strong spring/neap variation, and are generally oriented across isobaths [Wolanski et al., 1995a]. Fortnightly tidal ranges off the Fly River mouth reach 4 m during spring tides and can be < ∼0.5 m during neap [Baker et al., 1995]. On the seasonal scale, trade wind/monsoon variations likely have the greatest impact on the transfer of sediment within and beyond the shelf. Over the gulf, monsoon winds blow from the northwest during December–March, and trade winds blow from the southeast during May–October. Throughout the monsoon period, the New Guinea landmass ensures a short fetch and calm wave conditions (significant wave heights, Hₜ, of 0.3 m [Thom and Wright, 1983]). During the trade wind period, onshore winds blow from the east or southeast and generate substantially larger surface waves (Hₜ = 1.3 m [Thom and Wright, 1983]).
A pilot study in the gulf during monsoon conditions \cite{Walsh:2004} documented tidal amplitudes on the foreset region (45 m water depth) of the clinoform that ranged from 1 to 4 m over an 11-d observation period. Tidal currents reached a maximum of 27 cm s$^{-1}$ and wave heights were minimal. During these conditions, water column profiles showed that much of the low-salinity plume was contained nearshore (<20 m water depth) and near-bed suspended-sediment concentration reached up to 1 g L$^{-1}$ nearshore (<20 m), with little suspended sediment at depths >40 m. The data suggest that the relatively strong tidal currents resuspended recently deposited, unconsolidated sediments. Indeed, time series data from the foreset region during calm wave conditions showed that sediment was resuspended by spring tidal currents (>15 cm s$^{-1}$), although concentrations were relatively low, reaching a maximum of 0.1 L$^{-1}$ at 44 cm above bed \cite{Walsh:2004}.

On the basis of these data and earlier work by Harris \cite{Harris:1993} and Wolanski and Alongi \cite{Wolanski:1995}, Walsh \emph{et al.} \cite{Walsh:2004} developed a conceptual framework of clinoform processes. During the monsoon season, the surface plume is confined nearshore and the frontal zone between the plume and adjacent shelf water is predicted to interact with the seabed in shallow water near the Fly River mouth. Walsh \emph{et al.} \cite{Walsh:2004} hypothesized that sediment settles on the inner topset region of the clinoform (<20 m depth), shoreward of this frontal zone. Tidal currents periodically resuspend sediments and keep them from consolidating, but tides alone cannot account for much transport seaward. When trade winds increase the wave energy in the gulf, significant amounts of unconsolidated sediment are resuspended. These conditions are conducive to the formation of fluid muds dense enough to flow down slope to reach the foreset region of rapid sediment accumulation. Processes in the
3. Methods

3.1. Data Set

[15] Data for this study come from two cruises in the Gulf of Papua. The first cruise in January 2003 was designed to investigate shallow water processes and seabed signatures, and sampling was performed off a small research vessel, R/V Western Venturer. The shallow draft of the Western Venturer allowed detailed sampling of the inner shelf in water depths between 5 m and 20 m. The bathymetry of the inner shelf was measured using a fathometer. Water column and seafloor sampling was performed within 35 km of distributary channel mouths and upriver through the main navigational channel to the delta apex at the junction of the distributaries. In addition, sampling occurred in a grid consisting of six across-shore transects northeast of the Fly River mouth (Figure 1) on the clinoform topset, herein denoted the NE region. An anchor station located middistance along the Far North distributary was occupied for 8 h capturing the falling to low tide. Overall, a total of >100 water column profiles were collected along with 58 seabed cores. A few additional profiles were obtained in water depths <5 m using handheld instrumentation from a small inflatable boat. Geochemical results from the study are discussed by Goñi et al. [2006, 2008].

[16] The second cruise, in January 2004, was designed to examine the topset to foreset transfer of sediment in the Gulf of Papua. Sampling was performed from a large research vessel, the R/V Melville, and by necessity all work was performed in water depths >15 m. Approximately 50 stations within a grid of 14 transects were occupied. All of the transects (T1 to T13 and GH) included a shallow station in water depths of 15–17 m (Figure 1). Three small-boat transects were conducted in the NE region using a small handheld water column sampling system.

3.2. Water Column Profiling

[17] Suspended-sediment concentration and water properties were measured and sampled using a small profiler designed to induce minimal flow interference and observe these properties from the sea surface to the seabed. The Boundary Layer In situ Sediment Profiler (BLISP) contained sensors to measure salinity, temperature, depth and suspended-sediment concentration, and was fitted with a pump system designed to collect water/suspended-sediment samples near the seabed. The first sample was triggered immediately upon touching the seabed, which placed the nozzle at z = 10 cm above the bed (cmab). A second sample was collected at an elevation of interest relative to water column properties. The profiling instrumentation is similar to that used to document fluid muds on the Amazon shelf [Sternberg et al., 1991]. An optical backscatter sensor (OBS) on the profiler measured suspended-sediment concentrations throughout the water column. At rest on the seabed, the OBS sensor is located at 10 cmab [Martin et al., 2008]. OBS calibration procedures were accomplished using the methods outlined by Kineke and Sternberg [1992]. Separate calibrations were evaluated for the two areas of the clinoform topset (seaward of the distributaries and NE region) and the distributary channel. The OBS response varied by ±10% over the different regions. The minimum limit of concentration observable with the OBS sensor on the BLISP profiler was 5–7 mg L⁻¹.

[18] A hand-held conductivity-temperature-depth (CTD)–OBS system was used from small boats to obtain data in areas not accessible with the research vessels. The OBS was the same as that used on the BLISP profiler, and calibrations could be extrapolated from the other profiling work. The system was lowered through the water column by hand, and no water samples were obtained below the surface. There were no associated current measurements with the small-boat profiling.

3.3. Rusty Tripod

[19] A bottom boundary layer tripod was deployed at the same location in Umuda Valley during both cruises in 2003 and 2004 (Figure 1). The tripod, nicknamed “Rusty,” was equipped in 2003 with an Aanderaa current meter and vane (115 cmab), an autonomous (self-logging) OBS sensor (45 cmab), a CTD (44 cmab), and pressure sensor (44 cmab). In 2004, the current meter was replaced with an InterOceans electromagnetic current meter (80 cmab). The tripod was deployed for 7 d in January 2003 and for 30 d in January 2004. The OBS sensor was calibrated in the laboratory using seafloor sediment from the deployment site, and calibrations compared favorably with field concentrations measured with the BLISP profiler at nearby stations. Additional bottom boundary layer tripods were deployed on the clinoform topset/foreset during the January 2004 cruise [see Martin et al., 2008; Slingerland et al., 2008b; Crockett et al., 2008a], but data from these instruments are not discussed here.

3.4. Waves, Currents, and Shear Stress Analysis

[20] The vertical structure of the current at each station during the 2003 cruise was measured using a ship-mounted downward looking acoustic Doppler current profiler (ADCP, 600 kHz Workhorse, RD Instruments) (see Table 1), equipped with bottom-tracking capability. The three components of flow were recorded continuously in 25-cm vertical bins while the vessel was on station and were subsequently time averaged over 9.7 min, the approximate period of sampling operations. The horizontal current velocities were transformed into the geographical orthogonal coordinate system (i.e., north and east current components). Wave heights were evaluated from the pressure measurements recorded at 4 Hz over 10 min every hour at the Rusty tripod. The data were converted to significant wave height using spectral methods and linear wave theory to correct for pressure attenuation with depth. Bottom shear stresses were estimated from the near-bed current speed obtained at ~100 cmab at each station and wave heights evaluated at the Rusty tripod, and assuming that the wave heights were relatively consistent over the sampling region. A wave-current interaction model [Grant and Madsen, 1979] was used to estimate the shear velocity (u*), where the shear stress, τb = ρ u*², and ρ is the density of the fluid. An estimated bed roughness of 0.5 cm was used as input into the model. The value reflects a muddy seabed with slightly uneven surface due to bed forms or minor biological
influence as visually noted in seabed cores [Martin et al., 2008; Crockett et al., 2008b].

4. Results

[21] Throughout the shallow areas studied in the Gulf of Papua, the water temperature generally was limited to 29–30°C, with minimal difference between the fresh river and the saline Gulf waters. There was little spatial and vertical variation of temperature, and salinity dominantly controlled the fluid density.

4.1. River Discharge and El Niño Conditions in the Gulf of Papua

[22] The SOI during 2002–2003 was negative and sustained, but the magnitudes were not high (mean SOI index of −8.3) indicating moderate El Niño conditions. The sampling in January 2003 was in the midst of this year long period. In the latter part of 2003, the SOI returned to normal values and at the time of sampling in January 2004, the SOI indicated no anomaly. Thus we will discuss the results from the two sampling periods as representative of El Niño and normal conditions in 2003 and 2004, respectively, and assume that river discharge was relatively low in 2003 and normal in 2004. Measurements of river discharge during the time of sampling have not been published, but direct observations in January 2003 confirmed low river levels.

4.2. Delta Distributaries

[23] Under monsoon conditions in 2003, zero salinity fresh water extended into the delta distributaries, reaching about halfway between the apex and the mouth (Figure 3). Mixing with saline water started near station A8.5 and produced higher salinities of 13–18 at A10–A11 (Figure 3b) and 24 at the channel entrance (A12, Figure 3b). Note that these observations were performed over ~1.5 d and some of the contoured variability is due to tidal stage (Figure 3a). The vertical profiles of salinity under these low-discharge and low-wind conditions were strongly mixed, with stratification only apparent at the stations near the mouth and seaward in Umuda Valley (Figure 3b).

[24] At river stations upstream of the contact with saline water (A8.5), the suspended-sediment concentration was very high (>1 g L⁻¹ near the bed) and relatively well mixed throughout the water column (Figure 3c). Concentrations up to 700 mg L⁻¹ were observed near the water surface and were variable, both spatially and as a function of tidal stage. Downstream, at stations A9 and A10, salinities were 5–12 and lower concentrations of sediment throughout the water column were observed (generally <300 mg L⁻¹). Between this region and station A12, it appears that much of the suspended-sediment load had settled and was trapped in a near-bed layer leading to lower suspended-sediment concentrations throughout the water column (Figure 3c).

[25] An anchor station performed near station A8.5 captured the falling and low part of the tidal cycle and the beginning of the next flood cycle (Figures 3a and 4a). Profiles of salinity and suspended-sediment concentration were repeated on a half-hourly basis using the BLISP. The station was located nearshore to avoid shipping traffic, and thus as the small vessel rotated around anchor, the depth changed for various profiles causing the apparent irregular water surface shown in Figures 4b and 4c. Over the 8-h anchor station, the salinity fell from 10 (near high tide) to 0 (following low tide), and was generally well mixed in the vertical (Figure 4b). As the tide started to fall, a small amount of stratification was evident, but was quickly replaced by well-mixed, relatively fresh water. Large concentrations of suspended sediment were observed in the water column soon after high-slag and low-slag tides (Figure 4c). The suspended-sediment layer occurred in association with the upper limit of saline intrusion, was approximately 9 m thick, reached concentrations in excess of 3 g L⁻¹, and displayed a relatively sharp lutocline (i.e., vertical gradient in suspended-sediment concentration) in the upper portion of the water column.

[26] The shear stress (characterized by $u_*$) was calculated for the duration of the anchor station, and $u_*$ ranged from 0.4 to 4.1 cm s⁻¹ (Figure 4d). Stresses due to currents only were calculated, as surface gravity waves did not propagate into the delta channels at this time. An inverse relationship between suspended-sediment concentration and $u_*$ indicates that although $u_*$ values are high enough to resuspend sediment from the bed of the distributary channel, local resuspension is not the process that controls the suspended-sediment concentration in the estuary. Other processes (e.g., flocculation, rapid settling and trapping) in the low-salinity region (salinity of 0–6) appear to be controlling the sediment concentrations. During low-flow conditions encountered in January 2003, these estuarine processes occurred well within the Far North distributary channel.

4.3. Distributary Mouths

[27] Sampling in shallow water seaward of the distributary channels was undertaken to evaluate the influence of
the different distributary channels on seaward transport processes. A fathometer survey in these uncharted areas was performed to locate depressions (maximum morphological relief of \( \leq 5 \) m) extending seaward of the distributary channels. These areas were subsequently sampled by a series of at least five stations (circled areas in Figure 1), including stations in the deeper axis and stations on the shallower flanks. Sampling was performed during various stages of tide so that the comparison of specific stations is difficult, but the ranges of depth, salinity and suspended-sediment concentration observed in each station group can be compared and are summarized in Table 2.

At the southwest corner of the study area, the salinities and suspended-sediment concentrations observed in the G station groups show little river influence (Table 2). The bottom salinities varied between 32 and 34. Toward the northeast, bottom salinities ranged from 27 to 33 off the South distributary (E station group), were closer to oceanic salinities off the North and Far North distributaries (32–35 at the D and C station groups), and were relatively low to the north of the Far North distributary (17–30 at AA line). In addition, stations that were located at the outer edge of the delta islands (A and B station groups) showed significantly reduced bottom salinity (14–21). Overall, concentrations of suspended sediment were small in the SW (19–44 mg L\(^{-1}\) at G station group) and became progressively greater toward the northeast, with the greatest observed concentrations off the Far North distributary at the AA line. Suspended-sediment concentrations on the AA line were never less than 170 mg L\(^{-1}\) and at times reached concentrations >10 g L\(^{-1}\) (i.e., fluid mud concentrations). The more landward stations (as measured by distance from the apex, Table 2) off the North/Far North distributary (A and B station groups) exhibited intermediate suspended-sediment concentrations between those to the southwest and those observed on the AA line.

Figure 3. Transect through the Far North distributary. The 8-h anchor station was performed near station A8.5. (a) The tidal stage varied at each profile as the transect was performed over a day and a half time period. Profiles of (b) salinity and (c) suspended-sediment concentration vary as a function of distance from the apex.
4.4. Umuda Valley

[29] A transect extending from near the Rusty tripod site (∼16 m water depth) to 35 m water depth was performed in both 2003 and 2004 (Figure 5). The salinity structure of the water column in 2003 had maximum values of 34–36 near the seabed (Figure 5a). There was a near-surface layer with low salinities that ranged between 26 and 28, and the 32 isohaline was located at ∼5 m water depth. In contrast, during 2004, a much larger portion of the water column exhibited lower salinities (<32). The 32 isohaline intersected the seabed at >15 m depth at the landward extent of the transect, and was ∼10 m deep in the most seaward region (Figure 5c).

![Figure 4. Anchor station in Far North distributary occupied during a relatively strong spring tide.](image)

(a) The tidal stage ranged between falling and rising elevations. The profiles of (b) salinity and (c) suspended-sediment concentration vary as a function of the water surface elevation (Ht, height), as well as (d) current-induced shear velocity.

Table 2. Distribution of Sampling, Near-Bed Salinity, and Suspended-Sediment Concentrations at and Near the Distributary Mouths During the January 2003 Sampling Period

<table>
<thead>
<tr>
<th>Station Group (Location Description)</th>
<th>Distance From Apex, km</th>
<th>Depth Range, m</th>
<th>Salinity Range</th>
<th>Concentration Range, mg L⁻¹</th>
</tr>
</thead>
<tbody>
<tr>
<td>G (southernmost boundary)</td>
<td>130</td>
<td>10–15</td>
<td>31.9–34.2</td>
<td>19–44</td>
</tr>
<tr>
<td>E (off South distributary)</td>
<td>122</td>
<td>5.5–12</td>
<td>27.2–32.8</td>
<td>25–92</td>
</tr>
<tr>
<td>D (off north distributary)</td>
<td>123</td>
<td>5.3–16</td>
<td>31.8–34.0</td>
<td>29–198</td>
</tr>
<tr>
<td>C (valley off Far North distributary)</td>
<td>150</td>
<td>12–17</td>
<td>34.4–35.0</td>
<td>18–130</td>
</tr>
<tr>
<td>AA line (clinoform topset off Far North distributary)</td>
<td>139</td>
<td>8.1–10</td>
<td>17.4–30.2</td>
<td>170–1000+</td>
</tr>
<tr>
<td>B (mouth of North distributary)</td>
<td>114</td>
<td>4.0–13</td>
<td>16.3–20.7</td>
<td>48–270</td>
</tr>
<tr>
<td>A (mouth of Far North distributary)</td>
<td>107</td>
<td>16–21</td>
<td>14.0–18.5</td>
<td>68–560</td>
</tr>
</tbody>
</table>
The suspended-sediment concentrations observed between the two years are also quite different. In 2003, the peak observed concentrations were on the order of 50 mg L\(^{-1}\) and were only seen near bed at the shallowest stations (Figure 5b). At 35 m depth, there was negligible suspended-sediment concentration in the bottom boundary layer. In 2004, the peak concentrations were >100 mg L\(^{-1}\) at the Rusty tripod station. At the 35 m water depth site, suspended sediment in the bottom boundary layer was \(\sim 10-40 \text{ mg L}^{-1}\) (Figure 5d). In both years, suspended-sediment concentrations in the surface plume revealed little turbidity, rarely reaching values greater than the instrument resolution level of 5–7 mg L\(^{-1}\).

### 4.5. NE Region (Inner Topset)

In the NE region during the 2003 monsoon season, the low-salinity water was confined to shallow depths (Figure 6a). It is worth noting that even the shallowest stations are >10 km from the shoreline. Near the Far North entrance (AA transect), the salinity profile is better mixed than farther alongshore to the northeast where a stratified water column exists (e.g., compare AA line with EE line in Figure 6a). The surface lens of fresher water was about 3–4 m thick. In the offshore direction (lines L5 and L7, respectively, Figure 6a), the surface lens of low-salinity water becomes increasingly saline and near-bed salinity increases.

Suspended-sediment concentrations in the NE region were also highly variable in 2003 (Figure 6b), with bottom concentrations ranging from approximately 40 mg L\(^{-1}\) to >10 g L\(^{-1}\) [see also Goñi et al., 2006] and minimal sediment in the surface plume. The highest suspended-sediment concentrations (>10 g L\(^{-1}\)) were found at some sampling times on the AA transect (8–10 m water depths) and in the shallow regions of the DD and FF transects (~6 m water depth). In general, the greatest observed suspended-sediment concentrations (>100 mg L\(^{-1}\)) were found in the shallowest water depths, and a thick near-bed layer (~2 m) of suspended sediment persisted in shallow water depths (<8 m water depth, L3 and L5 in Figure 6b). This layer was no longer distinct where depths were >10 m (e.g., L7 in Figure 6b).

Wave conditions throughout the observation period were relatively low (\(H_s < 0.22 \text{ m}\)) and currents in the nearshore region were strong, reaching 42 cm s\(^{-1}\) at 100 cmab, and were largely driven by the tides. Calculated shear stresses in the NE region were highly variable, as stations were sampled at different tidal stages. Over the region, the wave-current shear velocities ranged from 0.9 to 2.7 cm s\(^{-1}\) (current-only shear stresses were calculated to be in the range of 0.2–2.0 cm s\(^{-1}\)). The estimated wave-current interaction shear velocity was dominated by the currents.

In the January 2004 monsoon season, limited sampling from water depths of 8–10 m in the NE region reveals differences in delivery of fresh water compared to January 2003. Surface salinities were generally lower by 2–6, and a distinct bulge of low salinity on transect CC was evident, with surface salinities of 18 during high tide at the same stations that had salinities of 28–30 in January 2003 [Goñi et al., 2006]. Suspended-sediment concentrations were similar to those observed in January 2003. Water column suspended-sediment concentrations were 0–20 mg L\(^{-1}\), and no sediment-laden surface plume was evident. Near-bed concentrations reached 80 mg L\(^{-1}\). These values are within the range of those seen in the January 2003 sampling at similar water depths.

### 4.6. Outer Topset (15–18 m Water Depth)

During the monsoon season of 2003, a transect along the 15-m isobath in front of the delta and NE region (Figure 7) showed only a relatively thin (2–3 m), low-salinity signal on the outer part of the topset of the clinoform (stations G15 to EE9)(Figure 7a). This lens, with a pycnocline centered on the shelf, is an important feature in the sediment budget of the Fly River delta.
Salinity, generally thickened and became fresher moving from southwest to northeast along the outer topset. Offshore of the NE region, from Umuda Valley to the northeast, there was a distinctly fresher body of water that extended to 15 m water depth. In contrast, during the monsoon season of the following year (2004), sampling on a similar transect was performed at water depths between 15 m and 18 m (Figure 8). In 2004, significantly lower salinity was seen in this shallow region (Figure 8a). The vertical profiles appeared somewhat more mixed with no distinct pycnocline through most of the region. Similar to observations in 2003, the data indicate an increase in the amount of fresh water from the southwest to the northeast. In this sampling, the lower surface salinity at Umuda Valley was a persistent feature, as reported for anchor station results of Martin et al. [2008]. They show that the salinity over the tidal cycle varied at levels below those seen in the surrounding transects. Thin surface lenses up to 2–5 m thick of low-salinity (<29) water extended.

Figure 6. Surveys of the shallow NE region showing (a) the spatial salinity structure and (b) the range of suspended-sediment concentrations. A lens of low-salinity water (<20) is retained very close to shore in water depths <5 m. The conditions in these shallow depths allow a significant amount of suspended sediment to be maintained in the water column. Note that the episodic observations of fluid mud are not included in these transects, but stars mark the locations where they were observed on transects at other times.
Figure 7. Transect along the outer topset in January 2003 (El Niño conditions) at ~15 m water depth showing (a) salinity and (b) suspended-sediment concentration. Note that this transect was performed over varying tidal stages and that although general trends can be detected, tidal variations at individual stations may also be present.

Figure 8. Transect along the outer topset in January 2004 (non-El Niño conditions) at ~15–20 m water depth of (a) salinity and (b) suspended-sediment concentration. Note that this transect was performed over varying tidal stages and that although general trends can be detected, tidal variations at individual stations may also be present.
over the NE region and distinct shallow plumes were detected offshore of the South and Far North distributaries. During the monsoon season of 2004, the magnitudes of suspended-sediment concentration were greater than in 2003, but their spatial pattern was similar. During the 2003 monsoon period, suspended-sediment concentrations were relatively low (generally <80 mg L$^{-1}$; Figure 7b) throughout the outer topset transect, but there are two regions where enhanced concentrations were observed in the current boundary layer seaward of the South and Far North distributaries. These regions are associated with the two major shelf valleys, Kiwai Valley and Umuda Valley, which impact the morphology of the clinoform structure [Crockett et al., 2008]. The greatest suspended-sediment concentrations were recorded at Umuda Valley and seaward of the NE region in a relatively thick (4 m) bottom boundary layer. In 2004, peak concentrations seaward of the South distributary were ~120 mg L$^{-1}$ and seaward of the Far North distributary were >500 mg L$^{-1}$ (Figure 8b).

4.7. Time Series Observations

The Rusty tripod deployment in January 2003 was for a 7-d period and extended over both spring and neap tidal conditions. In contrast, the tripod deployment in January 2004 lasted 30 d. This difference in deployment periods limits comparison of the time series observations over a full fortnightly cycle between the two years. Hence a subset of the January 2004 data set was extracted that displays similar tidal variation to those observed during January 2003 (Figure 9). Over the two time periods, the tidal variations ranged from a minimum of 0.57 m during neap tides to a maximum of 2.9 m during spring tides in 2003, and 0.66 m and 3.3 m in 2004, respectively.

There was little variation in salinity over the 2003 time series period. Salinity near the seabed ranged from 33.9 to 34.2 with little apparent variation between spring and neap tides. The currents in the bottom boundary layer varied with the tidal cycle, and maximum speeds were 20–25 cm s$^{-1}$ in the neap tide portion of the record and peaked at ~50 cm s$^{-1}$ in the spring tide portion. The suspended-sediment concentrations also varied with the tidal cycle, ranging from 15–20 mg L$^{-1}$ during neap tide to >150 mg L$^{-1}$ during spring tide. Note that these concentrations were measured at 45 cm ab, an elevation within the ~5-m-thick sediment-laden bottom boundary layer (Figure 5a), but may not have been close enough to the seabed to observe thin fluid mud layers.

In 2004, there was a stronger influence from the river influx during spring tide conditions than in 2003. The near-bed salinity throughout neap tides ranged from 33.9 to 35.0 and during spring tides ranged from 30.0 to 32.9 (Figure 9). The bottom boundary layer currents were stronger than in 2003, and more asymmetric over a daily tidal cycle (Figure 9), leading to a residual flow (landward near the seabed) that was most evident at spring tides. Peak currents during neap tide were 40–50 cm s$^{-1}$ and spring tidal currents reached 79 cm s$^{-1}$. The suspended-sediment concentrations at 45 cm ab peaked at 5–15 and 690 mg L$^{-1}$ during neap and spring tide, respectively. This is consistent with data from the BLISP transect shown in Figure 5b.

5. Discussion

5.1. Discharge of Fresh Water and Sediment to the Gulf

Prior investigations have concluded that bathymetry is a key factor in determining the deltaic tidal dynamics [Dalrymple et al., 2003; Harris et al., 2004]. During calm weather (e.g., monsoon conditions) in 1989–1992, it was predicted that about 60% of the freshwater discharge of the Fly River was directed through the South distributary, with the North and Far North distributaries carrying 30 and 10%, respectively [Wolanski et al., 1997]. However, small morphological changes at the point where distributaries join have subsequently been observed to be responsible for diverting flow from one distributary to another [Dalrymple et al., 2003]. Hence the present distribution of discharge through the distributaries is unknown. Sediment discharge estimates are even more problematic because of flow and sediment discharge complexities observed within channel cross sections [Harris et al., 2004] and the lack of observations from closely spaced instrumentation.
Comparison of the salinity and suspended-sediment concentration on the outer topset provides insight to recent water and sediment discharges from the individual distributaries. The stations at the southwest edge of the study area (G station group) are characterized by relatively high salinity and low suspended-sediment concentrations (Table 2) indicating that little fresh water and sediment were deflected toward the southwest and into the Torres Strait at the time of sampling. Although a low-salinity zone exists seaward of the South distributary, little sediment was associated with the fresh water signal, suggesting that sediment is trapped by estuarine processes within the South distributary. The station groups located off the Far North distributary (i.e., AA, A, B, and C) were generally less saline and showed the greatest suspended-sediment concentrations along the outer topset (Table 2). The groups with the lowest salinities were collected closer to the apex of the delta in the Far North distributary (i.e., A and B station groups, Table 2). The stations on line AA were just as far from the delta apex as those in station groups offshore of the other distributaries, yet exhibited the greatest suspended-sediment concentrations. These observations indicate that significant quantities of sediment escaping the delta pass through the Far North distributary during the monsoon season.

Along-shelf gradients are also seen during 2003 on the outer topset in deeper water (Figures 7a and 7b) with a general northeastward decrease in salinity and increase in near-bed suspended sediment. The data collected in 2004 (Figures 8a and 8b) exhibit lower salinities and higher suspended-sediment concentrations, presumably because of greater discharge during non-El Niño conditions. Overall, however, the spatial patterns during 2004 were similar to those observed during 2003. These results suggest that the northeastward gradient of decreasing salinity and increasing suspended-sediment concentration occur independently of El Niño conditions. The fresh water lens thickened and freshened toward the northeast (Figures 7a and 8a), suggesting discharge was additive as the clockwise general circulation advected the buoyant flow to the northeast (see circulation summary by Slingerland et al. [2008b]). The suspended-sediment gradient cannot be characterized as a consistently increasing trend but has independent zones of increasing concentration within the near-bed regions of the major shelf valleys (Figures 7b and 8b).

This observation is conceptually consistent with seabed studies [Crockett et al., 2008a] that show lower sediment accumulation rates in the southwestern region (transects T1-T7) than in the northeast region (transects T7-T13) suggesting that more sediment has accumulated in the northern part of the study area. It also is consistent with the seabed textural map developed by Harris et al. [1993], which shows a large band of fine-grained sediment trending seaward from the Far North distributary and toward the NE region, but with unresolved boundaries. An isopach map of this deposit shows greater thickness in the Umuda Valley and to the northeast than those of the sedimentary deposits in the southern part of the delta [Harris et al., 1993].

Studies by Wolanski et al. [1998] suggest that there is a net import of sediment to the Fly delta through the distributary channels during the trade wind season. However, on the basis of sediment accumulation data at the delta, they conclude that export to the marine environment must occur during the monsoon season. If this is true, and patterns observed in this study are maintained, the bulk of the sediment leaving the distributary channels appears to be doing so via the Far North distributary with delivery to Umuda Valley and northward into the NE region.

5.2. Transfer of Sediment From the River to the Outer Topset

5.2.1. Sediment Transport Processes Within the Delta Distributaries

Prior observations and modeling confirm that the extent of the salt intrusion is limited to the Fly River distributary channels [Wolanski et al., 1995b, 1997; Dalrymple et al., 2003; Harris et al., 2004]. Upstream of the salt intrusion, the suspended sediments are not highly flocculated [Wolanski and Gibbs, 1995] and, as seen in the Far North distributary, their concentrations are vertically uniformly throughout the water column (Figure 3c). Under the low-flow El Niño condition of 2003, where the river water first encountered the slightly saline water (salinity of 0–6) within the channel, the water column cleared. This is likely due to flocculation processes, consistent with observations of Wolanski and Gibbs [1995] in the South distributary. Much of the fluvial sediment load directly contributed to the estuarine turbidity maximum and was subsequently deposited in this region or maintained in suspension closer to the seabed boundary. The brackish water that reaches the mouth of the Far North distributary has greatly reduced suspended-sediment concentration (<100 mg L⁻¹ at the time of transect in Figure 3) with minimal sediment in the upper water column. Although surface plumes seaward of the distributary mouths are visible in satellite photographs of this region [e.g., Swarzenski et al., 2004], concentrations of sediment in this surface plume are low (Figures 7a and 8a) [Goni et al., 2006], usually <10 mg L⁻¹.

In addition to suspended load distributed throughout the water column, near-bed fluid muds have been documented in the Far North distributary and on the inner topset northeast of the Far North distributary [this study; Dalrymple et al., 2003; Goni et al., 2006] and within the North and South distributaries [Wolanski and Eaggie, 1991; Wolanski et al., 1997]. The seabed slope within these distributary channels is very low, with virtually no gradient between the mouths and the apex of the delta (i.e., estimated to be ~1 × 10⁻⁴ by Harris et al. [2004]). Calculations by Harris et al. [2004] estimated that a 1-m-thick layer of fluid mud formed as part of the estuarine turbidity maximum would flow seaward at <2 cm s⁻¹, and therefore sediment is likely not escaping via this pathway for much of the tidal cycle. However, if fluid muds were formed farther down channel in the Far North distributary during spring low tides, they would experience steeper seabed slopes of ~1 × 10⁻³ at the seaward extent of the estuarine turbidity maximum. These slopes do not exist at the mouths of the other distributaries. Therefore fluid mud suspensions in the Far North distributary would be nearer to open water regimes where they could escape seaward. Fluid muds were observed as part of the estuarine turbidity maximum at station A8.5 within the Far North distributary. The down-channel excursion of the turbidity maximum was estimated from measured velocities at the anchor station to be approximately 10 km, or a
location equivalent to that between stations A9 and A10 (Figure 1). This distance would be greater during peak spring tides and when the river flow was greater. In the following trade wind season when greater river discharge occurred, fluid muds were seen during spring tidal currents at an anchor station (T8-18) in Umuda Valley [Martin et al., 2008]. This observation suggests that fluid mud associated with the turbidity maximum could reach the zone of higher seabed slopes and flow down slope in Umuda Valley. Fluid mud is likely to form in other fine-grained tidally dominated deltas as sediment supply is sufficient and estuarine conditions are constrained within energetic tidal distributaries. Whether or not that fluid mud escapes the delta distributaries depends on the seabed slopes in the regions of fluid mud formation. [48] The Far North distributary, being open on its north side to the northeast inner shelf, allows water and sediment to flow into the NE region. The mouth of the Far North distributary is slightly closer (by ~10 km) to the apex of the delta than the other distributaries (Table 2), and thus estuarine processes will allow sediment to be released seaward at ebb and low tide more frequently (Figure 10). The other distributaries are more constrained because of the surrounding morphology (i.e., the landmass to the south and the elongate islands). The other distributaries also receive sediment discharged from upstream, but estuarine processes would be contained within them under normal river flow conditions [Wolanski et al., 1995b; Harris et al., 2004]. It is also possible that as the delta islands form and evolve over long periods of time, the channel of dominance changes much like delta lobe switching. The switching would change the relative locations of the estuarine turbidity maximum in each distributary causing differing distributaries to dominate in the discharge of sediment. Thus the morphological characteristics of tidally dominated deltas (e.g., apex bathymetry, topographic constraints of the distributaries) influence the rate and location of sediment discharge to the marine environment. 5.2.2. Sediment Transport Processes on the Inner Topset (NE Region) [49] As sediment reaches the NE region under normal river flow conditions, rapid deposition should occur due to flocculation and the decrease in near-bed stresses as the flow moves from the confined channel system to the open topset region. Seabed sampling by Crockett [2006] shows significant sediment accumulation at sites near the Far North distributary (AA line and BB5) over 100-a timescales. The sediment accumulation rates are maximum in the shallow regions near the Far North distributary (>3 g cm$^{-2}$ a$^{-1}$ at AA5 and 2.6 g cm$^{-2}$ a$^{-1}$ at BB5) and decrease alongshore toward the NE and seaward. Here, the seabed shows a surface mixed layer (>10 cm thick) that is regularly reworked and homogenized by physical processes, suggesting rapid delivery and/or regular resuspension.
by much higher concentrations relative to northeasts of AA and BB). The other domain is characterized away from Umuda Valley (across-shelf transects to the shoreline (generally depths of 7 m and greater) and seabed due to tidal currents. These stations are located away from the site and/or trapped in frontal zones due to tidal currents and waves. The other domain (above top dashed line) shows increased concentrations for a given value of shear velocity and reflects trapping and advection from other regions.

Suspended-sediment profiles in the NE region during 2003 (Table 2) suggest a sediment source from the Far North distributary, and a storage zone in the extensive area that is persistently reworked by tidal processes between the shallowest stations (line L3, Figure 1) and the shoreline. The relationship between shear velocity and suspended-sediment concentration (Figure 11) in the NE region is separated into two domains. One domain shows a consistent relationship between \( u^* \) and suspended-sediment concentration, and appears to be related to local resuspension of the seabed due to tidal currents. These stations are located away from the shallowest stations (generally depths of 7 m and greater) and away from Umuda Valley (across-shelf transects to the northeast of AA and BB). The other domain is characterized by much higher concentrations relative to \( u^* \) values. These stations fall significantly above the domain of tidal resuspension, indicating that this sediment must have been advected to the site and/or trapped in frontal zones in shallow water. All sites on transect AA showed higher suspended-sediment concentrations than those nearby, likely resulting from the close proximity to fine sediment that is delivered from the North and Far North distributaries. On all the transects, observed concentrations were high at the shallowest stations, even when shear stresses were unlikely to have caused enhanced suspended-sediment concentrations (Figure 11). This sediment must reside in the water column as high-concentration suspensions, or as a temporary deposit on the seabed without time to consolidate due to the strong tidal currents. The greatest shallow water concentrations were irregularly distributed along shelf, i.e., on some transects (AA, BB, and DD) but not others (CC and EE). This is likely due to processes that vary along the shoreline (e.g., tidal current focusing between mangrove regions and within the smaller river estuaries).

Clinoform structures have broad topset regions that provide the potential for temporary storage during relatively calm oceanic conditions. This storage capacity might be key for providing a bypass mechanism that delivers much sediment to the foreset, which clinoform deposits need to prograde. The conceptual model of clinoform progradation is that near-bed shear stresses on the topset keep sediment in suspension, and the mechanisms causing across-isobath transfer are generally modeled as a diffusive term, but not identified [e.g., Pirmez et al., 1998]. With net landward flows due to tidal asymmetry and near-bed landward flows due to buoyant influx of fresh water, other mechanisms are needed to provide the seaward transport of sediment. One likely possibility is fluid mud formation and gravity-driven flow during seasonal or episodic increases in wave or current energy. In the Gulf of Papua, this occurs seasonally with the onset of trade wind conditions. The temporary storage of sediment on the inner topset increases the exposure time of sediment to the water column, which has dramatic impacts not only on sedimentary processes but also on diagenetic chemical processes within the seabed [e.g., Aller et al., 2004; Goñi et al., 2006]. Frequent deposition/resuspension cycles expose the sediment and associated carbon to intense oxidation. Consequently, the temporary storage and reworking on the inner topset can cause extremely efficient processing of terrestrial and marine carbon.

5.2.3. Sediment Transport Processes on the Outer Topset

The wave environment did not change greatly between the monsoon sampling in 2003 and 2004, and thus the capacity to locally resuspend sediment at 15 m water depth on the outer topset should have been similar. However, in Umuda Valley, differences in the amount of sediment in suspension between the years were observed. In both years, the tripod results generally show that the suspended-sediment concentration increases as a function of shear velocity (Figure 9). There appears to be little interannual difference in the critical shear velocity needed for seabed erosion of \(0.8–1.3 \text{ cm s}^{-1}\) (Figure 12), which corresponds to tidal current speeds of \(25–30 \text{ cm s}^{-1}\).

Figure 11. Relationship between the wave-current shear velocity \( (u^*) \) and the suspended-sediment concentration (z = 10 cm ab) in the NE region showing two different domains. One domain (bracketed by dashed lines) shows a consistent relationship between \( u^* \) and suspended-sediment concentration and appears to be related to local resuspension of the seabed due to tidal currents and waves. The other domain is characterized by much higher concentrations relative to \( u^* \) values. These stations fall significantly above the domain of tidal resuspension, indicating that this sediment must have been advected to the site and/or trapped in frontal zones in shallow water. All sites on transect AA showed higher suspended-sediment concentrations than those nearby, likely resulting from the close proximity to fine sediment that is delivered from the North and Far North distributaries. On all the transects, observed concentrations were high at the shallowest stations, even when shear stresses were unlikely to have caused enhanced suspended-sediment concentrations (Figure 11). This sediment must reside in the water column as high-concentration suspensions, or as a temporary deposit on the seabed without time to consolidate due to the strong tidal currents. The greatest shallow water concentrations were irregularly distributed along shelf, i.e., on some transects (AA, BB, and DD) but not others (CC and EE). This is likely due to processes that vary along the shoreline (e.g., tidal current focusing between mangrove regions and within the smaller river estuaries). [50] Suspended-sediment profiles in the NE region during 2003 (Table 2) suggest a sediment source from the Far North distributary, and a storage zone in the extensive area that is persistently reworked by tidal processes between the shallowest stations (line L3, Figure 1) and the shoreline. The relationship between shear velocity and suspended-sediment concentration (Figure 11) in the NE region is separated into two domains. One domain shows a consistent relationship between \( u^* \) and suspended-sediment concentration, and appears to be related to local resuspension of the seabed due to tidal currents. These stations are located away from the shoreline (generally depths of 7 m and greater) and away from Umuda Valley (across-shelf transects to the northeast of AA and BB). The other domain is characterized by much higher concentrations relative to \( u^* \) values. These stations fall significantly above the domain of tidal resuspension, indicating that this sediment must have been advected to the site and/or trapped in frontal zones in shallow water. All sites on transect AA showed higher suspended-sediment concentrations than those nearby, likely resulting from the close proximity to fine sediment that is delivered from the North and Far North distributaries. On all the transects, observed concentrations were high at the shallowest stations, even when shear stresses were unlikely to have caused enhanced suspended-sediment concentrations (Figure 11). This sediment must reside in the water column as high-concentration suspensions, or as a temporary deposit on the seabed without time to consolidate due to the strong tidal currents. The greatest shallow water concentrations were irregularly distributed along shelf, i.e., on some transects (AA, BB, and DD) but not others (CC and EE). This is likely due to processes that vary along the shoreline (e.g., tidal current focusing between mangrove regions and within the smaller river estuaries).

Figure 12. Relationship between the wave-current shear velocity \( (u^*) \) and suspended-sediment concentration (z = 45 cm ab) from the Rusty tripod in Umuda Valley for the 7-d periods shown in Figure 9 during both 2003 (gray circles) and 2004 (black crosses).
Similar to the results from spatial studies in the NE region during 2003 (Figure 11), the temporal results in Umuda Valley show two domains. The first relates a predictable increase of suspended-sediment concentration with shear velocity (dashed lines in Figure 12) and the second consists of greater concentration than expected from the shear velocity (i.e., above the dashed lines in Figure 12). There are two significant differences between the time series data in 2003 and 2004: (1) stronger tidal currents were observed in 2004 and (2) a larger number of observations in the second domain were made in 2004. Both of these are consistent with more active advection to Umuda Valley in 2004 than in 2003. These differences indicate that less fresh water input throughout El Niño conditions of 2003 reduced the residual currents associated with estuarine circulation, and although the salt intrusion is located within the distributary, the estuarine circulation pattern extends to the outer topset.

[53] Along the outer topset, little suspended sediment was observed, confirming the conceptual picture of reduced sediment delivery to or redistribution at depths >10 m during monsoon conditions (Figures 7b and 8b). The exceptions are Kiwai and Umuda valleys, where the fresh water input and focusing of tidal currents are more tightly linked to the estuarine processes (Figures 7 and 8). In Umuda Valley, enhanced concentrations extended to ~35 m water depth. The existence of these shelf valleys with their enhanced slope and focusing of currents appears crucial to the transfer of sediment beyond delta distributaries and to the outer topset.

5.3. Impacts of El Niño Conditions

[54] The transect from the apex through the Far North distributary channel was not repeated during non-El Niño conditions in 2004. However, we can speculate on the impact of increased river flow, which would push the estuarine processes farther seaward (as seen by the decreased salinities during 2004 in Umuda Valley; Figure 5). Under these conditions, more fresh water would be able to leave the delta channel system and flow into the NE region. At spring low tides, the salt intrusion would extend to the Far North distributary mouth, decreasing flocculation and hindering particle settling. Thus the river load of sediment could be carried farther out the distributary channels, facilitating the formation of fluid muds in the NE region and in the steeper seabed morphology of Umuda Valley (Figure 10a).

[55] In contrast, observed suspended-sediment concentrations within Umuda Valley at similar depths on the outer topset were significantly less during the El Niño period compared to the non-El Niño period. This indicates that during El Niño, even at spring tides, the estuarine turbidity maximum was likely trapped landward of the mouth of the Far North distributary hindering sediment delivery to the steeper slopes found in Umuda Valley. Under these conditions, little sediment would have been transported down slope by fluid muds, and less unconsolidated sediment would be available for resuspension in the valley.

[56] The sediment delivery to the shallow waters of the NE region likely impacts transport processes in the following trade wind season, as there was less sediment in temporary storage on the inner topset (Figure 10b). These supply deficits were manifested in the seabed as variable levels of excess $^{210}$Pb [Crockett, 2006], and were likely the cause of seabed erosion in the following year on the outer topset [Martin et al., 2008]. Therefore, unlike temperate shelf systems where storms and associated floods (large positive events) dominate sediment discharge and transport and are recorded in the stratigraphic record, the dominant signal in the Gulf of Papua is a negative one of reduced river flow and presumably reduced sediment discharge.

[57] In summary, El Niño climate conditions reduce the river discharge reaching the continental shelf beyond the Fly River delta. The reduced discharge has multiple influences on the transfer of sediment to the marine environment, and the impacts reach to the prograding clinoform foreset. Reduced river discharge not only means less sediment reaches the delta, but also that estuarine processes occur farther upstream within the delta distributaries. Sediment that is part of the estuarine turbidity maximum will not reach the steeper slopes seaward of the distributary, where gravity flows can carry them toward the marine environment. Nor do they reach the NE region, which is an inner topset zone of temporary storage. In the subsequent trade wind season there is less poorly consolidated sediment to be resuspended and delivered to the forest region. Thus a reduced signal in river discharge propagates throughout the dispersal system.

6. Conclusion

[58] In a tidally dominated delta, such as the Fly River delta, distributary morphology and river discharge control the location and magnitude of sediment flux to the marine environment because estuarine processes that impact sediment settling and convergence occur within the delta distributaries where sediment can be retained. Rapid water column clearing due to flocculation processes and enhanced near-bed suspended-sediment concentrations were observed at the estuarine turbidity maximum in the Far North distributary of the Fly River delta, and the location of these processes moved up and down the distributary with tidal oscillations. The morphology causes a preferential delivery of sediment to the NE region, due to the proximity of the estuarine turbidity maximum to the mouth of the Far North distributary at spring tides. Also, the estuarine turbidity maximum may reach to the upper portion of Umuda Valley where seabed slopes are greater than those at the mouths of the other distributaries, and can enable fluid muds to travel seaward.

[59] The shallowest region of the NE inner topset (0–10 m water depth) is a zone of temporary storage of unconsolidated sediment. Tidal currents in the monsoon season are capable of resuspending recently deposited sediment, and trapping and advective processes actively maintain this zone of storage. The increased energetic of the trade wind season disperses the sediment to the foreset region, through fluid mud flows that were observed episodically near the river mouth. In general, clinoform structures have broad topset regions that provide the potential for temporary storage during relatively calm oceanic conditions. This storage capacity is important for producing the bypass mechanism of fluid mud formation and gravity-driven flow during seasonal or episodic increases in wave or current energy. These flows to the foreset deliver much
sediment, which clinoforms need to prograde. On the outer topset, near-bed sediment suspension is minimal in the calm monsoon season, with the exception of Umuala Valley where greater suspended-sediment concentrations were observed due to morphology that enhances residual velocities and creates a more direct link to discharge from the Fly River.

[60] In this system, El Niño conditions create a large negative perturbation to the relatively constant sediment discharge into the marine system. They cause a reduction in river discharge, which moves the estuarine turbidity maximum landward, limiting the sediment discharge from the distributaries to the inner topset. During El Niño conditions, therefore, deposition in shallow water of the NE region is reduced. This is a site of temporary deposition that normally provides sediment that is transported farther seaward. Therefore the impacts of reduced river discharge due to El Niño conditions propagate through the sediment dispersal system.

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