Continental shelf sediment transport and depositional processes on an energetic, active margin: The Waiapu River Shelf, New Zealand

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Continental Shelf Sediment Transport and Depositional Processes on an Energetic, Active Margin: the Waiapu River Shelf, New Zealand

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The Faculty of the School of Marine Science
The College of William and Mary in Virginia

In Partial Fulfillment
Of the Requirements for the Degree of
Doctor of Philosophy

by
Yanxia Ma 2009
This dissertation is submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy

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Abstract (477 words)

The Waiapu River drains a small mountainous basin, characterized by steep terrain, heavy rainfall, and unconsolidated soft Tertiary mudstone and siltstone. These factors, combined with heavy deforestation over the past 100 years have created one of the world’s highest sediment yields. Water discharge of the Waiapu River is very episodic over both inter- and intra-annual timescales, and almost all of the discharge is associated with floods brought by cyclonic storms. The Waiapu River drains an active margin that has a narrow shelf and steep slope. Marine conditions on the Waiapu continental shelf are very energetic, with strong waves as well as shelf currents. This special river-ocean system makes the Waiapu area an ideal site to study gravity-driven flows.

Instrumented tripods deployed at water depths of 40 m and 60 m on the Waiapu shelf directly offshore of the river mouth recorded data on waves, currents, and sediment fluxes from May through August, 2004. The tripod data showed direct field evidence of current-supported gravity flows on the Waiapu shelf. Data analysis indicated that the Waiapu River floods were characterized by two distinct phases: a flood phase and a resuspension phase. The flood phase was characterized by large sediment input, coupled with moderate to strong waves but weak currents. Strong near-bed sediment signals, however, were not recorded by the tripods until the post-flood resuspension phases, during which seaward near-bed currents started to intensify. A one-dimensional boundary layer model provided the inference that those strong seaward near-bottom turbid flows during the resuspension phase were dynamically similar to wave-supported gravity flows observed on Eel and Po Shelves, except that both waves and currents were important for sediment resuspension. In contrast to thin and dense wave-supported gravity flows, current-supported gravity flows on the Waiapu shelf were significantly thicker and more dilute.

Another two-dimensional model for wave- and current-supported sediment gravity flows was used to estimate sediment deposition on the Waiapu shelf from September 2003 to August 2004. The time period for the model calculations was divided into two segments: a low-energy (September to May) and high-energy portion (May to August). Model results showed that sediment delivered by the Waiapu River were trapped between the 20- and 80-m isobaths during the low-energy period, but then redistributed obliquely across the shelf between the 60- and 120-m isobaths during the high-energy period. Depositional locations estimated for the low- and high-energy portions, respectively, matched well with short- and long-term observed accumulation patterns based on $^{7}$Be and $^{210}$Pb activity as reported by Kniskern (2007; Kniskern et al. 2008). Sensitivity analysis indicated that the gravity-driven flows on the Waiapu shelf were mainly wave-supported landward of the 40-m isobath, but became increasingly current-supported as wave orbital decayed in deeper water.

This dissertation provided the first documentation of current-supported gravity flows, and hence contributed greatly to the study of sediment transport on continental shelves.
Continental Shelf Sediment Transport and Depositional Processes on an Energetic, Active Margin: the Waiapu River Shelf, New Zealand
Chapter 1: Fine-grained Sediment Dispersal off shore of High-load Rivers

1. Introduction

Approximately 15-20 billion tonnes of sediment is annually transported from global rivers to the oceans (Milliman and Syvitski, 1992). This fluvial sediment can be trapped in the river's delta, deposited on the adjacent coastlines, distributed on the continental shelf, or delivered to the deep sea over the shelf break or via canyons (Meade, 1996). Understanding the transport and fate of fluvial sediment is crucial to interpreting the geological history, seabed morphodynamic evolution and biogeochemical processes (involving e.g., carbon, pollutants, and nutrient cycles) in the ocean. In addition, fluvial sediment plays significant roles in ecological processes and supplies nutrients that impact primary productivity, fisheries, and coastal hypoxia.

Large rivers, with catchment areas greater than $10^5$ km$^2$, have long been considered to be major conduits for sediment transfer from the land to ocean. For example, the world's four largest rivers in terms of sediment load (Amazon, Ganges-Brahmaputra, and the Yangtze and Yellow before damming) together debouched about 4 billion tonnes of sediment into the oceans, accounting for 20% of total global discharge (Milliman and Syvitski, 1992). Those rivers have been well studied in terms of river processes, sediment characteristics, river plume dynamics, sedimentary properties, and shelf sediment transport. Over the past decade, however, it has become increasingly evident that small rivers (basin areas < 10,000 km$^2$), especially small mountainous rivers (e.g., Waiapu, Waipaoa in northern New Zealand; Eel in western coast of the US; Choshui in Taiwan) play comparable roles in delivering sediment to the oceans. Due to
high rainfall, steep gradients, active tectonic activities and an abundance of easily erodible material, small mountainous rivers tend to have very large sediment yields (Milliman and Syvitski, 1992). A unique feature of a small mountainous river is that, due to its small catchment size, water and sediment supplied by oceanic floods (Wheatcroft, 2000) is strongly coupled to oceanographic conditions of the coastal ocean. This implies that sediment delivery by small mountainous rivers tends to occur when oceanographic conditions are energetic and best able to disperse sediment, a characteristic that contrasts with the seasonal floods associated with large rivers.

Several reviews have classified or compared dispersal systems of well-studied rivers (Mckee et al., 2004; Walsh and Nittrouer, 2005; Wright and Nittrouer, 1995; Wright and Friedrichs, 2006). Based on the long-term accumulation patterns of fine-grained sediment and major factors controlling sediment burial on continental shelves, including tidal range, wave height, sediment supply and shelf width, major conclusions of the fates of riverine sediment can be summarized as:

1. Sediment accumulates in the vicinity of the river mouth to form a massive progradational delta. This happens if high sediment load is discharged into a coastal area with low marine energy and a wide gently-sloping shelf. The river plume gradually loses its ability to carry sediment as it merges with coastal water due to dispersion, bed friction, and rapid settling of flocculated particles. Typical examples include the Mississippi River (Coleman et al., 1998) and Yellow River deltas (Bornhold et al., 1986).

2. If sediment supply is abundant, and the coastal margin is wide and moderately energetic, sediment can bypass the inner shelf to be deposited on the less-energetic middle shelf. A subaqueous clinoform offshore of the river mouth is usually found on
these types of river margins, including those offshore of the Amazon (Allison et al., 1995; Nittouer et al., 1986), Yangtze (Liu et al., 2007), and Fly (Walsh et al., 2004). Such deposits can be elongated along the shelf as sediment is transported by prevailing coastal currents, and extend far from the river mouth. It was shown recently that the across-shelf equilibrium profiles of such clinoforms may be maintained by gravity-driven flows (Friedrichs and Wright, 2004).

III. Precluding the type I and II deposition, sediment can be dispersed widely in the ocean, accumulating on the mid to outer shelf or escaping from the proximal shelf entirely. This sort of river margin usually has energetic waves and/or currents that transport sediment directly or episodically. Transport mechanisms on such shelves include transport within the freshwater plume, dilute suspension of material after it settles from the plume or is resuspended in the bottom boundary layer, and gravity-driven cross-shelf transport of turbid layers (Harris et al. 2005). Gravity-driven flow plays a major role in moving the sediment across-shelf from the inner shelf site of plume delivery to the mid- or outer-shelf where they may form a preservable deposit (Traykovski et al., 2000; Harris et al. 2005). For gravity-driven transport, the site of the shelf deposition is highly correlated with the wave height and controlled by the shelf bathymetry (Friedrichs and Wright, 2004). Prominent examples of this type of dispersal pattern include the Eel (Traykovski et al., 2000), Waiapu (Kniskern et al., 2008), and Waiapoa River shelves (Kuehl et al., 2006). An exceptional example is the Mississippi River, whose south and southwest outlets extend southward nearly 60 km into deep water close to the shelf break (Walker et al., 2005).
IV. Significant sediment can also escape from the shelf through submarine canyons. An extreme example is the Sepik, into whose river mouth a steeply-sloping submarine canyon intrudes (Walsh and Nittrouer, 2003). The Eel (Mullenbach and Nittrouer, 2000), Congo (Droz et al., 1996), and Ganges-Brahmaputra (Kuehl et al., 1989) provide additional examples where submarine canyons play important roles in trapping or exporting river-delivered sediment.

Wright and Nittrouer (1995) argued that at least four stages are involved in the fate of sediment seaward of river mouths: (1) supply via river plumes; (2) initial deposition; (3) resuspension and transport by marine forces; and (4) long-term net accumulation. In spite of the distinct stages, sediment dispersal from a fluvial source to a depositional sink involves river plume dynamics, bottom boundary layer processes, and sediment transport processes.

2. River plume dynamics

Terrigenous sediment is delivered into the oceans through either positively or negatively buoyant plumes, referred to as hypopycnal or hyperpycnal effluents, respectively. With the exception of the Yellow River (e.g., Wang et al., 2006), large rivers are often associated with hypopycnal conditions, because they carry low concentrations of suspended sediment and large volumes of low-density fresh water, so that the density of the receiving ocean water usually exceeds that of the freshwater plume. Hyperpycnal conditions, where the density of the freshwater plume exceeds oceanographic density, are often seen in small mountainous rivers, especially during oceanic flood periods when sediment concentrations are elevated. Dispersal of fluvial
sediment is influenced by buoyancy and the interaction between the freshwater plume and the coastal ocean. The buoyancy, \( b \), and reduced density, \( g' \), associated with the river plume are estimated as

\[
b = -g' = -g \frac{\rho - \rho_0}{\rho_0},
\]

where \( g \) is gravitational acceleration, and \( \rho \) and \( \rho_0 \) are densities of river plume and ambient seawater, respectively.

When the freshwater from rivers meets salt water from the oceans, bottom convergence at the land limit of the salt intrusion traps river-delivered sediment, producing estuarine-like turbidity maximums. The efficiency of trapping depends on the position and strength of convergence as well as the sediment settling rate (Geyer et al., 2004). Deep intrusions and intense convergences increase trapping efficiency, as does the settling velocity of the sediment. The balance between the river effluents and salt water intrusion is indexed by the densimetric Froude number \( F \) (Wright, 1977; 1985),

\[
F = \frac{U}{\sqrt{g' h'}},
\]

where \( U \) is the effluent outflow velocity, \( g' \) is the reduced gravity, and \( h' \) is the thickness of the plume around the neighborhood of the river mouth. The bigger the \( F \), the more the salt water intrudes into the river.

After leaving the river mouth, the river effluents mix with ambient seawater. A salinity front is commonly seen at some distance from the river mouth. The frontal zone acts as another sediment trap due to the convergence of the near-bed flow and separation of the outflow from resuspension-caused bottom turbulence (Geyer et al., 2004). The position and extension of the frontal zone depends on the outflow velocity, outlet
geometry, and the strength of tidal or wind-driven currents in the receiving water. Outflow with high velocity can push the frontal zone far offshore (e.g., Amazon, Geyer and Kineke, 1995). Shallow receiving water with strong wind or tide typically exhibits a well-mixed region in the vicinity of the river mouth, resulting in a broad frontal zone that extends far from the river mouth (e.g., Amazon; Yangtze) (Geyer and Kineke, 1995; Beardsley et al., 1985). A sharp frontal zone near the river mouth is typical of conditions where the receiving water is deep and subject to weak winds and tides that favor stratification of the plume (e.g., Sepik, Kineke et al., 2000). The efficiency with which a frontal zone traps sediment is indexed by a length scale $L$,

$$L = \frac{U}{w_s} h'.$$

(3)

in which $w_s$ is sediment settling velocity. If the width of the front zone (seaward extension) is bigger than $L$, significant amount of sediment will be trapped in the zone (e.g., Amazon, Geyer and Kineke, 1995), otherwise only coarse sediment can be trapped and fine material will bypass the zone (e.g., Sepik, Kineke et al., 2000).

Beyond the frontal zone, sediment remaining in the river effluent is transported along with the plume by coastal currents typically in the along-shelf direction. The width of the river plume scales with the Rossby deformation radius, $R'$,

$$R' = \frac{\sqrt{g' h'}}{f},$$

(4)

in which $f$ is the Coriolis parameter. At scales greater than $R'$, the river plume tend to be rotated along-shelf to the right (Northern Hemisphere) or left (Southern Hemisphere) by the Coriolis force. If the ambient current flows in the same direction, sediment in the river plume will be advected far from the river mouth creating a coastal mud wedge (e.g.,
Amazon, Geyer et al., 1996; Yangtze in winter, Milliman et al., 1985), otherwise the plume will be broadened and slow down (Geyer et al., 2004).

3. Bottom boundary layer process

The bottom boundary layer serves as a primary link between the seabed and the overlying water column. It is through this layer that the momentum associated with water motion is transferred to sea floor sediment. Friction between the seabed and overlying moving water causes velocity gradients. This generates bottom shear stress that, when strong enough, initiates sediment movement on the seafloor. A boundary layer can be divided into three parts: viscous sublayer, log layer and outer layer. At the seabed, the 'no-slip' condition causes water velocity to be approximately zero. Just above the bed, there exists a viscous sublayer no more than a few millimeters thick, in which the molecular viscosity and eddy viscosity are roughly equivalent, turbulence is not important, and the velocity profile is linear. Above the viscous sublayer, within fully turbulent boundary layers, is the logarithmic layer or constant stress layer, which for fully developed (steady) boundary layers may be about 1 to 2 meters thick. Here the shear stress ($\tau$) can be assumed to be approximately constant, and is defined to be the bed shear stress ($\tau_b$) which represents the intensity of turbulence felt at the seafloor. In this log-layer, the velocity profile can be expressed by the Von Karman-Prandtl equation (i.e. the 'law of the wall') where velocity varies linearly with the logarithm of height above bed (see Dyer, 1986). Above the log-layer is the outer or Ekman layer, within which the flow still feels frictional resistance of the bed, but turbulent stresses decrease and the Coriolis force is also important.
These regions of a boundary layer (viscous sub-layer, log-layer, and outer layer) form under steady- and non-steady conditions, but take some time to become fully developed. For conditions typical of continental shelves, both wave orbital motions and quasi-steady wind-, density-, or tidally-driven currents impart friction and turbulence within the bottom boundary layer. The thickness of boundary layers formed under surface gravity waves, whose periods of oscillation are typically ~10 s, are limited in thickness by the oscillations of the flow field. The typical thickness of the wave boundary layer is only on the order of 10 cm. Within the wave boundary layer, turbulent stresses can be very high. With the same magnitude of wave orbital velocity and mean current speed, the velocity shear, hence the bed shear stress within the wave boundary layer, is much stronger than in the current boundary layer because the wave boundary layer is so much thinner. Wave-induced bed stress often dominates sediment entrainment and bedform formation.

Boundary layers for the quasi-steady components of the flow (wind- or tidally-driven currents, for example) can fully develop and become much thicker than wave boundary layers, being on the order of 10-m thick. Above the wave boundary layer, turbulence within the log-layer for the quasi-steady currents is increased by the presence of waves (Grant and Madsen, 1979). Factors that influence boundary layer processes include acceleration/deceleration of the flow, sediment induced stratification and physical roughness (see Wright, 1995; Dyer, 1986).

Since the 1980s, continental shelf boundary layer studies have been extended by combining the effects of wave-current interaction (Smith, 1977; Grant and Madsen, 1979), suspended sediment induced stratification (McLean, 1992), and bed roughness (Grant and
Madsen, 1982; Wiberg and Harris, 1994). Turbulence in these boundary layers is often represented using an eddy viscosity closure (Wiberg and Smith, 1983; Grant and Madsen 1986). Numerical models such as those developed by Wiberg et al. (1994) and Style and Glenn (2002) have been widely used to address boundary layer and associated sediment suspension processes. In recent decades, advances in field observation such as instrumented benthic tripods have enabled accurate measurements of velocity and suspended sediment concentration within the bottom meter of the seabed (Cacchione et al., 2006). The instruments include Acoustic Doppler Profiler (ADP) or Acoustic Doppler Current Profiler (ADCP) for monitoring mean current profiles throughout the water column, Acoustic Doppler Velocimeter (ADV) for recording three dimensional turbulent velocities and acoustic backscatter intensity at a point, Pulse Coherent-Acoustic Doppler Profiler (PC-ADP) for observing the velocity profile within the bottom boundary layer, and Acoustic Backscatter Sensor (ABS) for measuring the near-bed suspended sediment concentration profile. With these field instruments, new robust methods have been developed to estimate bed shear stress ($\tau_b$) (Kim et al., 2000; Sherwood et al., 2006), including the inertial dissipation (Green, 1992; Stapleton and Huntley 1995) and covariance methods. Concomitantly, the near-bed wave orbital velocity can be obtained reliably from high-frequency velocity measurements using spectral analysis or by calculating the variance of horizontal velocities (Wiberg and Sherwood, 2008).
4. Sediment transport process

The crucial stage involved in dispersal of river-borne sediment is resuspension and transport of temporarily deposited sediment from river plumes (Wright and Nittrouer, 1995). Sediment can be suspended when bed shear stress ($\tau_b$) caused by strong waves or currents exceeds a critical value ($\tau_{cr}$). In steady uniform flow, the mean suspended sediment concentration $\bar{C}_s(z)$ at some height, $z$, can be estimated by the diffusion equation,

$$\bar{C}_s(z)w_s = -\xi \frac{d\bar{C}_s}{dz},$$

(5)

where $w_s$ is the particle settling velocity, and $\xi$ is the eddy diffusivity which is directly related to the eddy viscosity, $A_z$ ($A_z = k\nu_z$), by a constant, $\beta$. Equation (5) implies a balance between downward sediment settling and upward sediment diffusion by turbulence. Integrating Eq. 5 yields the well-known Rouse equation,

$$C_s(z) = C_a(z/z_a)^{-w_s/\beta\nu_z},$$

(6)

where $C_a$ is the reference concentration at height $z_a$.

The suspended sediment can be divided into different size classes ($j$) and each class treated as a distinct phase. Therefore the total suspended sediment concentration is the sum of the concentration of each class, $C_{sj}$,

$$C_{sj} = C_{aj}(z/z_a)^{-w_s/\beta\nu_z}.$$

(7)

To specify the concentration profile precisely, the reference concentration should be taken at a reference level as close to the seabed as possible, because the concentration increases rapidly toward the seabed. Often, $z_a$ is selected as the height of the bed
roughness length or some multiple of the representative grain size. One commonly used formulation of reference concentration is (Smith and Mclean, 1977),

\[ C_{oj} = C_{b_{oj}} \frac{\gamma_0 \theta_j}{1 + \gamma_0 \theta_j} , \]  

(8)

in which \( C_{b_{oj}} \) is the concentration of size class \( j \) in the seabed, \( \gamma_0 \) is a resuspension coefficient, \( \theta_j = (\tau_b - \tau_{cr,j}) / \tau_{cr,j} \) is the normalized excess shear stress, where \( \tau_{cr,j} \) is the critical shear stress required to suspend the size class \( j \).

Above the constant stress layer, shear stress varies and the eddy viscosity profile begins to diverge from a linear relationship with height above bed. There, the balance between sediment settling and upward turbulent diffusion (Eq. 5) often continues to dominate the suspended sediment profile. Because the eddy viscosity takes a more complicated form, however, the Rouse profile (Eq. 6) is not valid as height above bed increases. There, the vertical balance of sediment fluxes (velocity times concentration profiles) can be solved either numerically (e.g., Wiberg and Smith, 1983) or analytically (e.g., Grant and Madsen, 1979) depending on the form assumed for the eddy viscosity.

Suspended sediment tends to produce a vertical gradient of density, suppressing the turbulence caused by velocity shear. The Richardson number is an important parameter to evaluate the importance of the stabilizing effects caused by stratification relative to the destabilizing effects induced by velocity shear,

\[ R_i = gs \frac{\partial c' / \partial z}{(\partial u / \partial z)^2} , \]  

(9)

where \( s \) is the dimensionless submerged weight of sediment particles relative to seawater \((s=1.65)\), and \( c' \) is volumetric sediment concentration. A negative feedback exists
between density stratification and velocity shear, which keeps the $Ri$ in the neighborhood of a critical value ($Ri_{cr} \sim 1/4$) (Wright et al., 1999).

5. A recent advance in sediment transport process: gravity-driven flows

Numerous recent field observations and model results have shown that gravity-driven transport of fluid mud can be a dominant mechanism for transporting fine sediment across shelves (Traykovski et al., 2000; Scully et al., 2003; Friedrichs and Wright, 2004; Harris et al., 2005). These flows are considered “hyperpycnal” in the sense that the turbid suspensate is heavier than the surrounding seawater, which typically has a density anomaly in the neighborhood of 25 kg/m$^3$. In the case of wave or current supported gravity-driven flow, the hyperpycnal condition is achieved because resuspended sediment is added to salt water that already possesses a density of about 1025 kg/m$^3$. This situation is distinct conceptually from traditional hyperpycnal river discharge in which higher sediment concentrations (~40 kg/m$^3$; Mulder and Syvitski, 1995) are required to make the fresh water heavier than seawater. In addition, gravity-driven flows are dynamically different from classical turbidity currents. The gravity-driven flows utilize the velocity shear generated by ambient waves and/or currents to maintain sediment in suspension (Wright et al., 2001; 2002), whereas the turbidity currents ascribe autosuspension to particle-particle interaction.

Gravity flows were first documented on the Amazon delta front by Sternberg et al. (1996) as ‘downslope migration of fluid muds’ supported by ‘boundary shear stress generated by waves and currents’. Detailed description of such fluid mud migrations, however, was not reported until the execution of the STRATAFORM program on the Eel
continental shelf in Northern California. Facilitated by advanced instrumentation of the bottom boundary-layer, Traykovski et al. (2000) quantified the important role of wave-induced density-driven flows in the cross-shelf sediment transport on the Eel shelf. Wright et al. (2001, 2002) revised the traditional Chezy equation and developed an analytic theory for the gravity-driven flows. The momentum balance in these flows was assumed to be between an upslope frictional drag and a down-slope pressure gradient force induced by suspended sediment,

$$ C_D u_{max} u_g = B \sin \theta, $$

where $C_D$ is the frictional drag coefficient ($=0.003$–$0.005$), $u_g$ is the velocity of the gravity-driven flow, $\theta$ is the bottom slope, whose sine is usually indicated as $\alpha = \sin \theta$, $u_{max}$ is the magnitude of the velocity at the top of the layer that includes the wave, current, and gravitational components of flow as defined below. $B$ is the depth-integrated buoyancy anomaly over the gravity flow thickness, $\delta$,

$$ B = g s \int_{z=0}^{\delta} c'dz. $$

On many shelves, $\theta$ is too gentle to support sustained gravity-driven transport via a condition of autosuspension, and particle settling will extinguish the flow unless additional turbulence from waves or currents maintains suspension. The effects of waves and/or currents on the gravity-induced flow velocity is appreciably augmented by wave-induced orbital velocity, $u_w$, and/or by the near-bed velocity, $v_c$, of wind and tide-driven along-shelf currents,

$$ u_{max} = \sqrt{u_w^2 + v_c^2 + u_g^2}. $$
Within gravity-flow layers, the velocity shear created by strong waves and/or currents enhances turbulence, but also resuspends sediment causing stratification. Stratification, however, suppresses turbulence and induces sediment settling, which in turn reduces the effect of stratification. When sufficient sediment is available, this negative feedback seems to keep the Richardson number, \( Ri \), in the neighborhood of a critical value \( (Ri_{cr} \sim 1/4) \). Wright et al. (2001) simplified the expression of \( Ri \) using scaling process as

\[
R_i = g s \frac{\partial \varepsilon'/\partial z}{(\partial u/\partial z)^2} = \frac{\partial^2 B/\partial z^2}{(\partial u/\partial z)^2} = \frac{B/H^2}{(U_{max}/H)^2} = \frac{B}{U_{max}^2} \approx Ri_{cr}. \tag{13}
\]

The above equation gives the theoretical maximum sediment load that a gravity flow can hold:

\[
\int_0^\delta \varepsilon' \, dz = B/g s = Ri_{cr} U_{max}^2 / (g s). \tag{14}
\]

Therefore, the maximum across-shelf flux can be estimated as

\[
Q_{g_{\max}} = u_g \varepsilon' = \rho \alpha R_i_{cr}^2 U_{max}^3 / g s C_D. \tag{15}
\]

Substituting (13) into (10) for the case of no ambient waves or currents yields the minimum shelf slope at which turbulent sediment gravity flows can be self-maintaining, i.e. auto-suspending (Wright et al., 2001):

\[
sin \theta_{auto} = C_D / Ri_{cr}. \tag{16}
\]

For \( C_D \approx 0.003 \) and \( Ri_{cr} \approx 1/4 \), \( \theta \approx 0.012 \). For shelves with a more gentle slope, ambient waves or currents are required to sustain sediment gravity flows.

Scully et al. (2002, 2003) extended gravity flow theory by considering wave-supported cases, and elucidated the significance of bathymetry to the formation of
flood deposition on the Eel shelf (Scully et al., 2002). If currents are negligible on the shelf (i.e., $v_c \approx 0$), by combining Eqs. (10), (12) and (13), $u_{\text{max}}$ can be simplified as

$$u_{\text{max}} = u_w / \sqrt{1 - \beta^2},$$

in which $\beta = \alpha R_{icr} / C_D$. Net deposition or erosion associated with gravity flows was thus determined by across-shelf gradients in sediment flux:

$$\text{erosion} = \frac{\partial Q_{\text{max}}}{\partial x} = \frac{\rho_s \alpha R_{icr}^2}{g\sigma C_D} \frac{\partial u_{\text{max}}}{\partial x} = -Q_r \left[ \frac{3}{u_w} \frac{\partial u_w}{\partial x} - \frac{1}{\alpha} \frac{\partial \alpha}{\partial x} \right].$$  

Equation 17 showed that across-shelf gradients in bed slope and wave energy control deposition and erosion. Sediment deposition is favored if the shelf is concave upward, while erosion is favored if the shelf is concave downward.

Motivated by results of Scully et al. (2003), Friedrichs and Wright (2004) developed an analytical solution for equilibrium bathymetric profiles offshore of river mouths. Assuming that waves dominate bed shear stresses relative to currents and that waves are linear and monochromatic ($u_w = \omega H / 2 \tanh(kh)$), the local slope $\theta$ is then related to water depth following the relation

$$\sin \theta \left[1 - (R_{icr} C_D)^{-1} \sin \theta \right]^{3/2} = 8 (\omega H)^{-3} (\sinh kh)^3 Q_r g \sigma C_D R_{icr}^{-2} \rho_s^{-1},$$

where $Q_r$ (kg/m/s) was the rate of riverine sediment discharge per unit distance along-shelf, and $\omega$, $k$ and $H$ were wave frequency, wave number and wave height. Towards shore, Eq. (18) converges on the asymptote

$$\sin \theta \propto Q_r H^{-3} h^{3/2},$$

which yielded the convex-upward profile typically observed on prograding river-nourished shelves and also on the landward portion of subaqueous deltae and
clinoforms. The equilibrium shelf slope tends to increase with river discharge and water
depth, and decrease with increasing wave height.

Besides direct field observations and these analytical approaches, numerical
models have been developed or updated to address the importance of gravity flows for
sediment transport. These include a one-dimensional model by Traykovski et al. (2007),
two-dimensional models by Scully et al. (2003, Eel shelf), Friedrichs and Scully (2007,
Po shelf), and Hsu et al., (2007), as well as a three-dimensional model developed by
Harris et al., (2004, 2005, Eel shelf), and also used by Kniskern (2007, Waiapu shelf).

6. Examples of sediment dispersal offshore of large and small rivers

Three large (watershed area >10^5 km^2), two intermediate (10^4~10^5 km^2) and three
small (<10^4 km^2) mountainous rivers are presented as examples in this section to
illustrate the relative roles of different dispersal modes of river delivered sediment. The
examples tend to cover distinct situations in river size and location, oceanographic energy,
fate of sediment, and degree of coupling between the ocean and river.

6.1 Large rivers

The Amazon is the prototypical example of a system with high sediment and
water discharge and strong oceanographic energy on its shelf. The river-born sediment
deposited on the inner shelf has formed a modern subaqueous delta that is prograding
seaward (Allison et al., 1995). The clinoform-shaped subaqueous delta extends northward
along the shelf for several hundred kilometers (Nittrouer et al., 1986). Suspension by
currents and transport within gravity flows both play important roles in the formation and
configuration of the delta. As the largest river in the world, the Amazon discharges about 6300 km$^3$/yr of fresh water and $10^9$ T/yr of sediment into the ocean with little temporal variation (Meade et al., 1985). The buoyant plume can be entrained to the northwest by the North Brazil Current for hundreds of kilometers. Additionally, a strong plume pushes the estuarine-like salinity front zone offshore to the mid-shelf region where much sediment is trapped to form fluid mud (Kineke and Sternerg, 1995). Sternberg et al. (1996) found that the important source of sediment to the prodelta region was gravity-driven down slope migration of fluid mud that was formed in shallow water by flux convergence. Strong currents are the major source of turbulence on the Amazon shelf and control the resuspension of bottom sediment there (Geyer et al., 1996). Rapid accumulation of sediment occurs on the mid-shelf in response to a modest reduction in tidal amplitude (Kuehl et al., 1986).

The Yangtze is the largest river in the southern Asia with mean water discharge and sediment load of about 900 km$^3$/yr and 480 MT/yr prior to upstream damming, respectively (Milliman and Syvitski, 1992; Xu et al., 2006). It drains into the East China Sea (ECS) and is characterized by the fact that water and sediment discharge tends to peak at a different time of the year than sediment transport (Milliman et al., 1985). The Yangtze is under a monsoon regime and periodically experiences dry winter storms with very strong winds from the north (winter monsoon). The peak discharges of water and sediment, however, occur primarily in the wet summer (summer monsoon) when the ocean is relatively calm. During summer the freshwater plume extends southeastward immediately upon leaving the river mouth, following the direction of the river channel, then turns to the northeast due to local wind or potential vorticity conservation (Hu et al.,
Retarded by mixing with ambient seawater and bed friction, the plume loses its ability to carry sediment. Thus, some sediment (~40%) is trapped near the river mouth and the remainder escapes seaward to nourish the subaqueous delta (DeMaster et al., 1985). During winter, storms cause intense waves and fairly strong southwestward coastal currents. The unconsolidated sediment is resuspended by waves and/or currents (DeMaster et al., 1985), and then transported southward by the coastal current reaching as far as the Taiwan Strait (~1000 km) (Liu et al., 2007; Xu, 2006). Few near-bed field surveys have been conducted on the Yangtze Shelf, and there is little direct evidence of gravity-driven flow on the shelf of East China Sea. Intense turbid layers, however, were found over the shelf during winter by Yang et al. (1992) and, more recently, confirmed by Oguri et al. (2003), suggesting the possibility of density flows.

The Mississippi River provides the prototypical example of a large river emptying into an area having low oceanographic energy. It debouches a relatively large amount of freshwater (530 km³/yr) and sediment (210 MT/yr; 400 MT/yr prior to damming) into the Gulf of Mexico through multi-order distributaries (Milliman and Meade, 1983). Historically, the long-term and large scale alongshore dispersal of Mississippi sediment has depended more on lobe switching than on oceanographic forces (Wright and Nittrouer, 1995). Short-term sediment dispersal is largely controlled by river plume processes and local wind dynamics (Walker et al., 2005) because waves and tidal currents are typically not strong enough to resuspend previously deposited sediment and along-shelf flow that would transport resuspended sediment is weak, due to the impediment of bird-food delta (Adams et al., 1982). Bed friction and buoyancy-dominated plumes are well represented at the mouth of the Mississippi (Wright,
201977). The Pass a Loutre and other distributaries to the east of the South Pass enter shallow water, thus turbulent bed friction impacts river effluents there. Middle-ground bars and bifurcating channels were found around the mouths of these passes (Wright and Coleman, 1974). On the other hand, the South and Southwest Passes protrude into deep water near the edge of the continental shelf. Well-developed stratification caused by salt wedge intrusion effectively isolates the river effluent from the bed friction, hence buoyancy-dominated deposition is favored (Wright, 1985) and some sediment can escape from the continental shelf entering into the deep sea (Walsh et al. 2006). Additionally, the river plume structure and consequent sediment deposition are highly modified by local winds. During periods of east winds, prevalent in winter, autumn and spring, the river plumes are driven westward around the delta, and usually turn toward the coast at 50 km west of the delta feeding a clockwise gyre. This helps trap the river water and sediment on the shelf (Walker et al., 2005). In summer, the west winds reverse the directions of river plume and current, and disperse river plume along with associated sediment offshore (Walker et al., 2005).

The Yellow River is famous for its large sediment load (1100 MT/yr prior to damming), and extremely high sediment concentration with 25 kg/m³ on average (Ren and Shi, 1986), ranging from 11 to 222 kg/m³ (Qin and Li, 1983). Its drainage basin is influenced by the same monsoon regime as the Yangtze, hence its highest sediment discharges occur primarily in the summer. However, the Yellow River enters the relatively shallow Gulf of Bohai (average depth ~ 30 m) (Wright et al., 1990), which has a much smaller fetch than the open ECS. Thus the waves and wind-driven currents in the Bohai are weaker than those on the shelf of the ECS. The along-shelf flow offshore of the
river mouth is dominated by tidal currents with velocities on the order of 1 m/s (Wiseman et al., 1986). Although hyperpycnal effluents are frequently seen in the Yellow River, the outflows are rapidly extinguished close to the river mouth, and most of the sediment is trapped within the near shore areas, 30 km off the river mouth (Bornhold et al., 1986; Wright et al., 1990) due to rapid deposition and sharp lateral shear fronts (Li et al., 2001). The unconsolidated fine sediment can be resuspended by tidal currents as well as storm waves. This sediment may be transported by along-shelf tidal currents around the Shandong Peninsula, southward to the South Yellow Sea and mix there with sediment from other sources (Ren and Shi, 1986; Liu et al., 2004; Yang and Liu, 2007). Wright et al., (2001) applied gravity-driven flow theory to field measurements from the 1980s. They found that turbidity flows maintained by tidal currents had the potential to move sediment downslope during slack tide when eddy viscosity was temporally relaxed. Martin et al. (1993) also found nepheloid layers with high sediment concentration between the Yellow River mouth and Shandong Peninsula, and concluded that these layers played an important role in sediment transport from the Bohai to the Yellow Sea.

6.2 Intermediate rivers

Favored by heavy rainfall (~ 6 m/yr) and extreme elevation of the drainage basin (head water elevation ~ 4000 m), the Fly River (Papua New Guinea) delivers a large amount of sediment (~85 MT/yr prior to mining, ~125 MT/yr today) into the Gulf of Papua (GOP) (Harris et al., 1993). The river plume seldom extends beyond the inner shelf (20-m isobath). Hence a significant fraction of Fly sediment is trapped in the estuary and deposited adjacent to the mouths of the river creating a large subaqueous clinoform delta.
(Walsh et al., 2004). Based on $^{210}$Pb activity, however, the long-term maximum accumulation rate (4 cm/yr) occurs at the foreset of the clinoform delta, which is located far away from freshwater plume deposition (Walsh et al., 2004). The oceanographic energy of the GOP varies highly with the seasons. During the southern hemisphere summer, the mild north-westerly monsoon dominates the wind field and the Gulf is relatively quiescent (Walsh and Nittrouer, 2003). Sediment exported from the estuary is deposited temporarily on the inner shelf. During the winter, the GOP is energized by large waves produced by south-easterly trade winds and long-period swells propagating from the Coral Sea (Hemer et al., 2004). The stored sediment along with newly discharged material is resuspended and concentrated as fluid muds by convergent estuarine circulation. Those fluid muds are then transported by gravity-driven flows to the less-energetic clinoform face (Walsh et al., 2004; Harris et al., 2004).

The Po River flows into a low energy system, a micro-tidal area with small waves. It annually delivers 15 million tonnes of sediment to the Adriatic Sea mostly during fair-weather conditions (Nelson, 1970). The Po mouth is a salt wedge type, and the salt wedge intrusion causes significant flocculation just within the river channels (Fox et al., 2004). As soon as the freshwater enters the sea, it spreads and diffuses kinetic energy quickly. A significant fraction of the sediment settles rapidly and forms a mud deposit in very shallow water depths of 4-15 m (Fox et al., 2004). Field and modeling results applied to the Po subaqueous delta showed that wave-driven gravity flows during high wave events move sediment seaward, contributing to progradation of the subaqueous Po delta (Traykovski et al., 2007; Friedrichs and Scully, 2007). Although waves offshore of the Po delta are weak, the wave-driven current and southeastward flowing western Adriatic
coastal current (WACC) still impact the along-shore dispersal of Po sediment. Though, during Boras, much Po sediment may be trapped in the northern Adriatic, a portion is carried southward where it can mix with sediment from numerous smaller Apennine rivers (Harris et al., 2008). Materials from the Po and those smaller rivers act to create a very long dispersal system (>500 km) in a shape of lateral cliniform (Nittouer et al., 2004; Palinkas and Nittouer, 2007). During the southward transport, bottom Ekman flows veer the suspended sediment offshore to facilitate the seaward progradation of the Apennine cliniform (Nittouer et al., 2004).

6.3. Small rivers

The Eel is a prominent example of a small mountainous, or an “oceanic” (Wheatcroft, 2000) river. It has a high sediment yield (1720 T/km²/yr), and extremely episodic, seasonal water and sediment discharge. Eel River floods usually coincide with large winter storms and thereby high waves and strong wind-driven currents (Wheatcroft, 2000). Eel plumes are strongly influenced by local winds and Coriolis force. Upon leaving the river mouth, the plumes typically turn northward in response to southerly wind forcing and Coriolis rotation, and are typically trapped against the coast in shallow waters by downwelling favorable winds (Geyer et al., 2000). River plume sediment settles out of the plumes and is concentrated within the wave bottom boundary layers as fluid muds (Traykovski et al., 2000). Episodically, large wave energies carry sufficient suspension drive about some of those sediment to move across shelf as gravity flows, carrying much of the 20% of the total Eel load that is preserved on the mid-shelf flood deposit (Harris et al., 2005; Sommerfield and Nittouer, 1999). The remainder of the Eel
load is either buried in near-shore sands (Crockett and Nittrouer, 2004), transported along-shelf as dilute suspension (Harris et al., 2005), or carried offshore within the Eel Canyon (Puig et al., 2003).

The Waipaoa River, located on the east coast of northern New Zealand, discharges 15 MT sediment into the Pacific Ocean each year. Although the drainage basin is small (~2205 km²), sediment yield of the Waipaoa is extremely high (~6800 T/km²/yr) (Page et al., 2001). The Waipaoa continental shelf is an active margin with a steep slope (200-m isobath at 25 km offshore). Marine conditions within Poverty Bay and on the open shelf are fairly energetic, with local storm waves and ocean swell in the bay exceeding 4 m and 6 m, respectively, and persistent long period (8-12 s) swell on the shelf reaching 1-2 m (Orpin, 2004). The thickness of Holocene sediment on the mid-shelf offshore of the Waipaoa is about 30 m (Foster and Carter, 1997). Recent analysis of ²¹⁰Pb (Kuehl et al., 2006) showed the long-term sediment accumulation rate to be low just offshore of the bay mouth, but to increase both to the north and south on the mid- and outer shelf. A bypassing zone is hypothesized to exist just offshore of Poverty Bay, but transport mechanisms there remain unknown (Kuehl et al., 2006). Considering its high sediment load, intense marine forces and steep slope, it is very likely that gravity-driven flow dominates the across-shelf sediment dispersal there.

Seldom rivers in the world have experienced more frequent hyperpycnal event, with real sediment concentration measurement, than the Taiwanese Rivers. This is the combined results of Taiwan’s nature setting, which is characterized by active tectonic activity, erodible rocks, high runoff, steep gradient and frequent impacts of typhoon events, and high human activities including agriculture expansion, urbanization, and road
construction (Kao and Milliman, 2008; Milliman et al., 2007). Of all the primary Taiwanese rivers, the Choshui River, located on the central-western Taiwan, has received comprehensive field measurements near the river mouth (Milliman et al., 2007) as well as offshore on the shelf (Liu et al., 2008; Xu, 2006). The Choshui River drains a mountainous area of 3300 km$^2$, with mean basin gradient of 1.8 % and heavy rainfall of 2.2 m/yr (from Taiwan Water Resource Agency report, TWRA). Annually, it discharges 64 MT of sediment into the Taiwan Strait (TWRA), and its sediment yield is about 20,000 T/km$^2$/yr, one of the highest in the world (Milliman and Syvitski, 1992). Most of the sediment (>75%) is delivered during typhoon-induced floods (four typhoons on average during summer), which are often related to hyperpycnal events (Kao and Milliman, 2008; Milliman and Kao, 2005). Oceanographic condition offshore of the Choushui River is fairly energetic, with strong tidal currents (mean speed of about 0.5 m/s) and weak waves (wave height of 0.3 m with period of 8 s) (Wang et al., 2003). The sediment deposition offshore of the Choushui River is characterized by massive sand waves in near-shore area, and large silt-sand-dominated deltaic clinoform far offshore (Liu et al., 2008). This clinoform appears at about 30-m isobath, reaches its maximum thickness (up to 50-m thick) at about 70-m isobath, and progrades radiantly to the west (Liu et al., 2008). The deposition feature of the Choushui sediment and strong currents in the Taiwan Strait suggest that current-supported gravity-driven flows could dominate the across-strait sediment dispersal there.
7. Dissertation objectives and Hypotheses

Most recent studies on gravity flows focused on wave-supported cases. Current-supported gravity flows have seldom been documented due to the lack of high resolution near-bed observations of currents and suspended sediment profiles. The Waiapu River, with high sediment yield, small size, energetic coastal ocean (strong waves and currents) and an adjacent steep continental shelf, provided an ideal natural laboratory to further examine the role of gravity flow mechanisms in the offshore transport of riverine sediment. A recent boundary layer field experiment there using state of the art instrumentation provided evidence that near-bed gravity flows may not be confined to the wave boundary layer and can in fact be triggered by sediment concentrations that are much smaller than previously thought. To explore these issues, my dissertation addressed the following hypotheses:

(1) Based on an existing model of wave-induced gravity-driven flows and examination of bathymetric and wave data on the Waiapu shelf, it was hypothesized that shoreward of 50 m water depth sediment is bypassed to deeper water. It was expected that maximum deposition should occur seaward of 50 m water depth where the steepness of the slope begins to decrease and wave agitation diminishes.

(2) Gravity-driven flow is likely the dominant mechanism of rapid seaward sediment transport and of subaqueous clinoform progradation. Since currents are very strong on the Waiapu shelf, gravity flows there are expected to be current supported.

(3) The Holocene mud wedge preserved on the Waiapu shelf was predicted to be controlled by current-supported gravity-driven transport.
The objective of this dissertation was to understand sediment dispersal patterns offshore of the Waiapu River. Field tripod data were analyzed and a one dimensional boundary layer model was used to examine mechanisms that cause across-shelf sediment transport. Since current-supported gravity-driven flow was shown to occur in this environment, an existing two-dimensional gravity flow model was modified and applied to estimate depositional patterns on the Waiapu continental shelf. This dissertation has helped improve understanding of the physical processes that control the movement of sediment across continental shelves. Currents, either wind-driven or tidal, are a dominant marine force on many continental shelves reviewed in this chapter including the Amazon, Yangtze, Yellow, Fly and Choushui. With high sediment load and sloping seabed, current-supported gravity flows have great potential for moving sediment offshore of initial deposits.

8. Outline of dissertation

In this dissertation, Chapter Two reported the initial field observations of the relationships among high river discharge events, wave- and current-induced bed stresses, suspended sediment concentration, shelf slope, and the across-shelf transport of near-bed turbid layers. An emphasis was placed on a specific two-day flood event to compare those observed relationships to an analytic theory for sediment gravity flow. The chapter provided direct field evidence of current-supported gravity flows on the Waiapu shelf.

Chapter Three further analyzed the field observation data by focusing on three flood events. Each of the floods was characterized by two distinct phases: a flood phase.

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1 Chapter Two has been published as Ma, Y., L.D. Wright, C. Friedrichs, 2008. Continental Shelf Research 28, 516-532.
with initial riverine sediment input and a resuspension phase few days later with sediment remobilization. A boundary layer model was used to complement the field observation, evaluate the significance of current-supported gravity flows relative to wave-supported transport, and identify the mechanisms dominating cross-shelf sediment fluxes. During the three events, estimated gravity flow velocities of current-supported were much larger than those of wave-supported, and matched well with observed near-bed across-shelf velocities. The thickness and concentration of those gravity flows were significantly thicker (1-2 m) and more dilute (2-4 kg/m$^3$) comparing with those of wave-supported gravity flows observed on the Eel and Po shelves.

Chapter Four extended the small-scale analyses facilitated by tripod data to the lateral direction by applying a two-dimensional gravity flow model to estimate sediment deposition on the Waiapu shelf. The model was used to represent September 2003 to August 2004, encompassing a low-energy period (LEP, Sep. to May 2003) and a high-energy portion (HEP, May to Aug. 2004) of the field deployment. The estimated depositions for LEP and HEP match remarkably with observed short-term (5 months) and long-term (100 years) deposition previously observed by Kniskern, (2007), respectively. The main conclusion reached in this chapter was that during low-energy periods the sediment delivered by the Waiapu River was trapped in water depths shallower than 60-80 m, whereas during the high-energy period, previously deposited sediment was moved further offshore along with any newly delivered material. The mechanisms dominating this seaward sediment transport transition from being wave-supported gravity flows to current-support gravity flows as the material moves from the inner shelf to deeper water.
This dissertation was the first in the literature to document field observation of current-supported gravity flows, and provided the first attempt to identify the thickness and dynamics of these flows. It reproduced the Holocene mud deposition on the Waiapu shelf via a gravity flow model and elucidated the mechanisms of sediment fluxes on the Waiapu shelf. The dissertation contributed by extending our knowledge of key processes that operate on continental margins with high sediment yields.
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Chapter 2: Observations of Sediment Transport on the Continental Shelf off the Mouth of the Waiapu River, New Zealand: Evidence for Current-Supported Gravity Flows*

By Yanxia Ma, L. Donelson Wright, Carl T. Friedrichs

Abstract

Instrumented tripods deployed at depths of 40 m and 60 m on the shelf off the mouth of the Waiapu River on the east coast of New Zealand’s North Island recorded data on waves, currents, and sediment fluxes from May 22 through August 10, 2004. Three major flood events and several wave events occurred during the deployment. Data from acoustic Doppler velocimeters and profilers revealed that downslope sediment fluxes accompanied a flood event of late June, during which near-bed downslope current speeds approached $0.5 \text{ m s}^{-1}$. The most pronounced downslope transport within the benthic layer occurred the day after peak flood but coincided with strong isobath-parallel currents. Suspended sediment concentrations about one meter above the bed were on the order of 2-4 g l$^{-1}$ at times of maximum seaward flow. Suspension of freshly-discharged sediment within the hyperpycnal layer over much of the profile was maintained by the high bed stresses associated with the strong benthic currents, in contrast to the wave-supported sediment gravity flows recently reported on other river-nourished shelf systems. Nonetheless, observed concentrations and velocities were largely consistent with the dynamics of critically-stratified sediment gravity flows based on equations previously applied elsewhere to wave-supported cases. Despite lower concentrations in current-supported gravity currents, the greater thickness results in similar total loads. In further contrast to earlier results, our data suggest that as sediment off the Waiapu flowed into deeper water across the seaward steepening bottom profile, autosuspension may have aided in thickening and accelerating the gravity current.
Keywords: Gravity flow; Sediment transport; Hypopycnal plume; Autosuspension; Waiapu River shelf; New Zealand

1. Introduction

In recent years, convergent field and modeling studies have highlighted the significance of sediment gravity flows to the transport of fine sediment across continental shelves (Ogston et al., 2000; Traykovski, et al., 2000, 2007; Wright, et al., 2001, 2002; Scully et al., 2002, 2003; Friedrichs and Scully, 2007; Hill et al., 2007). As noted in a recent review article by Wright and Friedrichs (2006), wave- and/or current-supported sediment gravity flows have now been identified on shelves off more than a dozen high load rivers. Unlike classical turbidity currents, which provide enough turbulence via their gravity-induced motion to keep sediment in suspension, wave- and current-supported gravity flows utilize the velocity shear associated with wave orbitals and/or the ambient along-shelf current to help maintain sediment in suspension.

Several recent studies of gravity flows near river mouths have specifically focused on the dynamics and consequences of wave-induced fluid muds, most notably on the Eel shelf off northern California and off the Po River in the Adriatic Sea (see references cited above). On these shelves, turbulence confined within the thin (≤ 10 cm) wave boundary layer has been observed to maintain very high concentration suspensions (~ 10 to 100 g l⁻¹), forming gravity flows trapped below an intense lutocline. The wave-generated turbulence then allows the gravity currents to move downslope across much gentler slopes than flows associated with auto-suspending turbidity currents or with direct hyperpycnal discharge from rivers.
Unanswered questions highlighted by Wright and Friedrichs (2006) include the details of how wave-supported gravity flows evolve as they move offshore into depths with less energetic waves and/or steeper bed slopes. As wave-supported gravity flows move into deeper water, the difference in velocity shear between the wave boundary layer and mean currents in the overlying water column drops, and upward spreading of sediment through a weakened lutocline may result. The loss of sediment to upward spreading may help extinguish the gravity flow. Where bed slope increases offshore, the contribution of the gravity flow itself to total velocity may eventually exceed the magnitude of ambient waves and currents. This would further favor upward growth of the boundary layer, entrainment of overlying fluid, and could also mix away the gravity flow. Alternatively, if the slope becomes sufficiently steep, the gravity flow could accelerate offshore and perhaps transition to autosuspension.

The shelf off the Waiapu River on the east coast of New Zealand's North Island provides an ideal laboratory for further investigating the dynamics of sediment gravity flows, including the role of ambient shelf currents and the offshore evolution of gravity flows with increased depth and bed slope. Few of the world's rivers have sediment yields as high as that of the Waiapu (Hicks et al., 2000), and seasonally energetic swell from the Southern Ocean favors frequent wave suspension near shore. Along-shelf currents associated with persistent regional circulation are also strong (Chiswell, 2000), suggesting that mean currents are also likely to play a role. The shelf directly off the river mouth is bathymetrically relatively smooth and steep (with slopes at 60-m depth more than two times greater than off the Eel). Finally, the bathymetric profile across the inner-to mid-shelf is convex upward, potentially favoring a transition to autosuspension with
depth. In this paper, we report on initial field observations of the relationships among high river discharge events, wave- and current-induced bed stresses, suspended sediment concentration, shelf slope, and the across-shelf transport of near-bed turbid layers. We further compare these observed relationships to analytic theory for wave- and current-supported sediment gravity flows. Our goals are to elucidate better the mechanisms of hyperpycnal sediment fluxes on continental shelves and to extend our knowledge of key processes that operate on margins with high sediment yields.

2. Theoretical background

The evolving theory of gravity-driven transport on shelves is detailed in a recent series of papers (Wright et al., 2001, 2002; Scully et al., 2002, 2003; Friedrichs and Wright, 2004; Wright and Friedrichs, 2006). The negative buoyancy of turbid near-bed layers is expressed by the depth integrated buoyancy anomaly, $B$ ($m^2 s^{-2}$), which depends on the suspended sediment volume concentration, $c'$, and turbid layer of thickness, $h$, via

$$B = gs \int_0^h c' dz,$$

(1)

where $g$ is the acceleration of gravity, and $s$ is the submerged weight of sediment particles relative to seawater. For hyperpycnal layers on a sloping bed with slope $\theta$, the quantity $B \sin \theta$ yields a downslope pressure gradient force.

On many shelves, $\theta$ is too gentle to support sustained gravity-driven transport via a condition of autosuspension, and particle settling will extinguish the flow. However, where wave or current induced flows add their effects to that of the gravity-induced flow velocity, $u_g$, the quadratic friction term is appreciably augmented by wave-induced orbital
velocity, $u_w$, and/or by the near-bed velocity, $v_c$, of wind and tide-driven along-shelf currents. At lowest order the force balance can be represented by

$$B \sin \theta = C_D U_{\max} u_g,$$  \hspace{1cm} (2)$$

where

$$U_{\max} \approx \sqrt{u_g^2 + u_w^2 + v_c^2},$$  \hspace{1cm} (3)$$

and $C_D$ is a dimensionless bottom drag coefficient on the order of 0.003 to 0.004.

Another closely related conclusion of Wright et al. (2001) was that, on gentle slopes, feedback between turbulent flows and sediment-induced density stratification within a gravity flow maintains the gradient Richardson number, $R_i$, close to the critical value of 1/4. The Richardson number, $R_i$, which indexes the relative influence of buoyancy in suppressing turbulence versus the ability of shear to generate turbulence within the gravity flow is:

$$R_i = \frac{g \frac{\partial \rho}{\partial z}}{(\frac{\partial u}{\partial z})^2} \equiv \frac{g_s (\partial c' / \partial z)}{(\partial u / \partial z)^2} \equiv \frac{B}{U_{\max}^2}.$$  \hspace{1cm} (4)$$

When $R_i < 1/4$, intense turbulence suspends additional sediment, increasing $B$ and $R_i$, while for $R_i > 1/4$, decreased turbulence causes sediment to settle, decreasing $B$ and $R_i$. Through the resulting feedback (c.f., Trowbridge and Kineke, 1994; Kineke et al., 1996), $R_i$ tends to remain near $R_{icr} = 1/4$.

Substituting (4) into (2) for the case of no ambient waves or currents yields the minimum shelf slope at which turbulent sediment gravity flows can be self-maintaining, i.e., auto-suspending (Wright et al., 2001):
\[ \sin \theta_{auto} = \frac{C_D}{Ri_{cr}} . \]  

If a shelf slopes more gently than this, ambient waves or currents are required to sustain sediment gravity flows. For \( C_D \approx 0.003 \) and \( Ri_{cr} \approx 1/4 \), \( \sin \theta_{auto} \approx 0.012 \). Solving (2) – (4) for the case of ambient waves but no current (other than the gravity flow) yields

\[ u_g = u_w (Ri_{cr} C_D^{-1} \sin \theta)^{\left\{1 - (Ri_{cr} C_D^{-1} \sin \theta)^2\right\}^{-1/2}} . \]  

Motivated by results of Scully et al. (2003) that suggested the majority of gravity-driven sediment transport bypasses the convex upward portion of the Eel shelf, Friedrichs and Wright (2004) used (2) – (4) to develop an analytical solution for equilibrium bathymetric profiles off river mouths. Friedrichs and Wright (2004) defined an equilibrium shelf profile as that for which the gravity-driven sediment flux directed offshore equals the inshore sediment supply from the river during large discharge events. Assuming that waves dominate along-shelf currents and that waves are linear and monochromatic, the local slope \( \theta \) is then related to water depth following the relation

\[ \sin \theta \left\{1 - (Ri_{cr} C_D^{-1} \sin \theta)^2\right\}^{-3/2} = 8(\omega H)^{-3} (\sinh kh)^3 Q_r g s C_D Ri_{cr}^{-2} \rho_s^{-1} , \]  

where \( Q_r \) (kg m\(^{-1}\) s\(^{-1}\)) is the rate of riverine sediment discharge per unit distance along-shelf, \( \rho_s \) is the density of the sediment, and \( \omega, k \) and \( H \) are wave frequency, wave number and wave height.

Towards shore, Eq. (7) asymptotes to

\[ \sin \theta \propto Q_r H^{-3} h^{3/2} , \]  

which yields the convex-upward profile typically observed on prograding river-nourished shelves and also on the landward portion of subaqueous deltas and clinoforms.
Furthermore, (7) and (8) both highlight the tendency of equilibrium shelf slope to increase with river discharge and decrease with wave height. Equilibrium slope increases with $h$ to compensate for the effect of decreasing orbital velocity in order to maintain a spatially uniform sediment flux toward the shelf break without across-shelf convergence. Equilibrium slope increases with sediment supply because a greater slope is required for the gravity-induced offshore sediment to match a larger $Q_r$. Equilibrium slope decreases with increased wave height to compensate for larger $u_w$.

As shelf slope steepens with depth offshore, the bracketed term on the left hand side of Eq. (7) asymptotes to zero such that

$$\sin \theta(h \to \infty) = \sin \theta_{auto} = \frac{C_D}{Ri_{cr}},$$

regardless of wave height or riverine sediment supply. Note that both the shoreward (8) and seaward (9) asymptotes are each independent of wave period, and thus also hold (within the assumptions of (1) – (4)) for much lower frequency current oscillations for which $U \approx 0.5 H/h (gh)^{1/2}$, such as tides and Kelvin waves. Thus these relationships may hold at least asymptotically for current-dominated shelves as well. A main limitation on the validity of (9), however, may be associated with the original assumption inherent in (1) – (4) that high sediment concentrations remain trapped with a thin bottom boundary layer.

As $h$ increases and $\sin \theta \to \sin \theta_{auto}$, Eq. (6) predicts that the mean velocity of the sediment gravity flow itself grows rapidly until it exceeds that of the wave orbital velocity which supported the suspension in the first place. With $C_D \approx 0.003$ and $Ri_{cr} \approx 1/4$, Eq. (6) specifically predicts $u_g > u_w$ for $\sin \theta \geq 0.0085$. At this point there will no longer be a strong gradient in shear between the wave boundary layer and the mean current. The
suspension is then likely to spread rapidly upward, reducing the near-bed sediment concentration and possibly reducing the strength of the gravity current. Alternatively, the greater slope could facilitate a self-maintaining, auto-suspending sediment gravity flow.

Wave-supported gravity currents similar to those predicted by (2) - (4) have been previously documented in detail by tripod deployments on the Eel Shelf (Ogston et al., 2000; Traykovski et al., 2000) and on the Po prodelta (Traykovski et al., 2007). However, in both these cases the bed slope at the tripod deployment sites was significantly less than \( \sin \theta_{auto} \), namely \( \sin \theta \approx 0.005 \) off the Eel and \( \sin \theta \approx 0.002 \) off the Po (Traykovski et al., 2007). In both of these cases, it was observed that \( u_g \) was less than \( u_w \), consistent with the predictions of (6) for these relatively gentle slopes. The much steeper mid-shelf slope at tripod deployment sites off the Waiapu (\( \sin \theta \approx 0.010 \) to 0.012) provides a unique opportunity to examine the effects on sediment gravity flows of \( \sin \theta \) approaching \( \sin \theta_{auto} \) and also the effects of mean currents exceeding the strength of wave orbital velocities.

3. The field site

The Waiapu River (Fig. 1) drains a mountainous, tectonically active landscape comprised largely of unconsolidated Tertiary mudstone. The river has an extreme specific sediment yield of 20,520 tonnes km\(^{-2}\) yr\(^{-1}\) of suspended sediment, which is among the highest yields on Earth, and it supplies about 35 x 10\(^6\) tonnes of suspended sediment per year (Hicks and Griffiths, 1992; Hicks et al., 2000). This high yield, relative to an average discharge rate of 80 m\(^3\) s\(^{-1}\), causes suspended sediment concentrations to be exceptionally high: the average of suspended sediment concentration is greater than 10 kg
m$^3$, and scattered point measurements since 1979 (NIWA, unpublished data) show maxima of $\sim 75$ kg m$^{-3}$ accompanying water discharge rates of $\sim 200$ m$^3$ s$^{-1}$. River flows with suspended sediment concentrations $>40$ kg m$^{-3}$ may be considered “hyperpycnal”, meaning that the bulk density of such flows exceeds the density of seawater. The Waiapu typically produces several hyperpycnal events during any given winter flood season (June through September). The pair of photographs in Fig. 2 shows a segment of the highly erodable catchment (Fig. 2a) and the sediment-charged river effluent (Fig. 2b) of the Waiapu.

The continental shelf to which the Waiapu River drains is composed of a Holocene sediment wedge that is over 100 m thick directly off the river mouth in the region of the 100-m isobath where the highest accumulations appear to have occurred (Orpin et al., 2002). This wedge thins to 20-30 m beneath the inner shelf and pinches out near shore. This significant Holocene deposition has created a convex-upward shelf profile as is commonly seen over the landward portion of progradational clinoform shelves. Between 10 m and 40 m, in particular, the profile shape is consistent with the simple equilibrium model of Friedrichs and Wright (2004) that explains this shape in terms of a balance between the supply of river sediment near the “top” of the convex portion of the profile and gravity-driven bypassing of an equal amount of sediment at its base. The seaward steepening of the bed relatively near shore is considered to allow increases in downslope gravity to offset the diminution with depth of wave agitation.

In the case of the Waiapu shelf, bed slopes at depths shallower than 30 m are gentler than the critical slope, $\sin \theta_{cr}$, of 0.01 required to sustain autosuspending turbidity flows (Wright et al., 2001), whereas slopes seaward of $h = 30$ m are adequate. The
steepest portion of the shelf occurs around the 35-m isobath where \( \sin \theta = 0.013 \) (Fig. 3). Slopes in the region between the 40-m and 60-m isobaths, where our instruments were deployed, varied locally from \( \sin \theta = 0.012 \) at the 40-m isobath to 0.010 at a depth of 60 m. The average slope over the 20-m depth interval separating the two tripods was 0.0104, which is near the critical value for autosuspension. Not all sediment transferring this region of the Waiapu shelf is necessarily bypassed, however. Based on coring and seismic profiling, Wadman and McNinch (2007) have recently concluded that some of the sediment being delivered by density flows is being deposited in distinct layers of silts and clays within the more prevalent muddy sands.

The wave climate of the Waiapu shelf is mixed storm waves and swell. Gorman et al. (2003) hindcasted a 20-year (1979-1998) record of coastal waves around New Zealand using the WAM wave generation model of Hasselmann et al. (1988). The results showed that the waves affecting Tatapouri (the station nearest to the field site) arrive primarily from the south and south-south east. The mean significant wave height at Tatapouri is 1.6 m and the maximum annual wave height typically exceeds 8 m. Corresponding mean and maximum wave periods are 7.4 s and 14.0 s (Gorman et al., 2003). Strong coast-parallel currents systems affect the bottom boundary layer of the Waiapu shelf. The inner shelf is subject to the northward flowing Wairarapa Coastal Current (WCC) while the outer shelf and upper slope are swept by the southward flowing East Cape Current. The WCC is an extension of the merged Southland and D'Urville currents with an estimated volume transport of 1.6 Sv and mean surface current of 0.2 m s\(^{-1}\) (Chiswell, 2000). The ECC is an extension of the south-east flowing East Auckland Current with a volume transport of 24 Sv and geostrophic speeds of up to 0.5 m s\(^{-1}\) (Stanton et al., 1998).
4. Methods

Two instrumented bottom boundary layer tripods were deployed on the Waiapu shelf at depths of 40 and 60 m to obtain data on mechanisms and products of hypothesized gravity-driven transport and to verify and refine analytical and numerical models. The two tripods recorded time series allowing characterization of waves, bed stress, buoyancy anomalies, near-bed current and sediment flux profiles, and information on currents in the overlying water column. Figure 4 shows the configuration of a tripod, and Fig. 1 indicates the deployment locations. The tripods were deployed over most of the southern hemisphere winter (late May through August) of 2004, capturing the season of most significant floods and swell-dominated high-energy events which mobilize bed sediment.

Each tripod held an acoustic Doppler velocimeter (ADV) for measuring three-dimensional turbulence and suspended sediment concentration; a conductivity sensor; pressure sensors; an upward-facing profiling Doppler current meter for measuring currents between the sea surface and the region a short distance above the tripod; and sediment settling traps. The tripods were also equipped with downward-aimed pulse coherent-acoustic Doppler profilers (PC-ADP) that provide high-resolution near-bed current profiles and bed elevation. These instruments are capable of reporting velocities and suspended sediment concentrations in vertical bins spaced 1-2 cm apart, down to approximately 1.6 cm above the bed. Acoustic backscatter sensors (ABS) were also on the tripods, but the ABS instruments were severely damaged and the data were not recovered. A YSI multi-purpose sensor on the 60-m tripod provided additional data on
the concentration of total suspended solids in addition to temperature and salinity. Unfortunately the YSI on the 40-m tripod failed. Table 1 summarizes the tripod instruments and their deployment elevations above the bed.

Suspended sediment concentrations were determined after post-deployment calibration. Acoustic backscatter from the ADVs and turbidity readings from the YSI transmissometer were related to suspended sediment concentration through laboratory calibration, executed in a 2-m high stirring chamber which generates strong turbulence to mix the fluid and sediment well. Native sediment collected by sediment traps mounted on the 40- and 60-m tripods during the deployment were utilized. For each case, sediment samples were separated into mud and sand using a 63-μm sieve and up to 15 sediment concentrations (0.013 - 3.5 g l⁻¹) were calibrated using various mud and sand fractions. The instruments used for post-experiment calibration were the 60-m YSI and the 40-m lower ADV because the other ADVs were damaged during either deployment or retrieval. It was assumed that the calibrations for the other ADVs that returned data were similar since they were identical models purchased together.

Velocity data from the ADV and PC-ADP were recorded in their respective beam coordinate systems. PC-ADP data were processed to remove velocity ambiguities and rotated into the earth coordinate system (north-east-up) according to the heading, pitch and roll measurements from the compass (Lacy and Sherwood, 2004). The ADV instrument package did not include an integral compass, but the PC-ADP and ADV were visually aligned on the tripod, and velocity data from ADV were rotated in parallel with the PC-ADP, which did include a compass. The velocity profile data from the Doppler profilers were recorded directly in the earth coordinate system. Accounting for magnetic
declination (−23° E for east shore, north island, NZ), the current velocities measured by
the above instruments at both the 40- and 60-m sites were rotated to along- and
across-shelf directions relative to the orientation of the local bathymetric contours on the
Waiapu shelf.

To complement the observations, a 1-D numerical model was used to estimate
combined wave-current bed stresses. The basic theory of this model was formulated and
developed by Smith (1977) and modified by Wiberg (e.g. Wiberg and Smith, 1983;
Wiberg et al., 1994). It assumes current flows are steady and uniform and uses an
exponentially spaced grid in the vertical direction to ensure very fine resolution within
the bottom boundary layer. The model uses a continuous eddy-viscosity profile to
estimate vertical velocity profiles and combined wave-current shear stresses, and it also
accounts for the effects of sediment-induced stratification and bed roughness. The model
requires as inputs wave and current velocities as well as characteristics of sea-bed
sediment.

5. Results

The tripods recorded data on waves and currents from May 22 through August 14,
2004. During this time, three major flood events and several high-energy wave and
current events occurred. The period of most reliable data from both tripods extended from
about 1 June through 21 July, 2004, so we focus on the events that took place during that
time. Within this period, we attach major emphasis to a particular event that occurred on
30 June 2004. During this event, the Waiapu River reached a flood stage and the
associated discharge, measured at a gauging station 20 km upstream from the river mouth,
exceeded 2000 m$^3$ s$^{-1}$; total suspended solids within the river, calculated from the rating curve of Hicks et al. (2004), exceeded 50 kg m$^{-3}$. About a day after the peak flood, the same storm that brought the rains to the river catchment brought strong wind-generated currents and high swell to the shelf. A significant positively-buoyant surface plume spread seaward and turned northward during the time of maximum river discharge. A day later, seaward-setting flows also prevailed near the bed in association with high current- and wave-induced bed stresses. In what follows, we detail the forcings and across-shelf sediment transport consequences of this and two comparable events.

5.1 River discharge, winds, waves and currents

Numerous recent papers on the dispersal and deposition of river sediment in coastal seas (e.g. Ogsten et al., 2005; Wright and Friedrichs, 2006; Wright and Nittrouer, 1995, among others) rely implicitly on the concept that rivers deliver sediment to the inner shelf and oceanographic forces then move those sediment onward. The primary determinants of the processes we address are thus the river discharge, the wave- and current-producing winds and the sediment-disturbing waves and currents.

Time series of surface winds, Waiapu River discharge, total suspended sediment concentrations within the lower course of the river, and root mean squared (r.m.s.) wave height over the period 1 June- 21 July 2004 are shown in Fig. 5. The r.m.s. wave height was estimated from linear theory as $H_{rms} = 2u_w \sinh(kh)/\omega$, where the representative r.m.s. amplitude of wave orbital velocity $u_w$ and radian frequency $\omega$ were determined from the 40-m tripod velocity records via power spectral analysis of near-bed velocity following Madsen (1994). An attempt was made to solve for the full surface wave spectra,
but without a high frequency near-bed pressure record to help isolate wave oscillations at higher frequencies, turbulence from strong mean currents contaminated the high frequency tail of the velocity spectra. Unfortunately, the wind data were obtained from a station located about 100 km to the northwest of the field site and may not accurately represent local conditions.

Three prominent flood events with durations of 2-3 days are evident in the discharge data shown in Fig. 5c; they are centered on June 20, June 30/July 1 and July 20. During each event, water discharges approached or exceeded 2,000 m$^3$s$^{-1}$, more than 20 times higher than the annual mean value of 85 m$^3$s$^{-1}$; Based on the Hicks et al. (2004) rating curve, suspended sediment concentrations within the lower course of the river reached 50 kg m$^{-3}$, making the river water distinctly hyperpycnal with respect to seawater (at least within the lower river channel). In each of the three cases, the weather systems that brought the rains responsible for the floods began with strong northerly winds that ultimately veered to become westerly or southerly as the event progressed. Moderately energetic waves with r.m.s. wave heights of 2.5 to 3.5 m (significant waves heights of 3.5 to 5 m) coincided with, or followed shortly behind, the floods.

Time series of the r.m.s. amplitude of bottom orbital velocity, $u_w$, and absolute burst-averaged mean current speeds, $u_c$, as measured by acoustic Doppler velocimeters (ADVs) on the 40-m and 60-m tripods are shown in Fig. 6. The resulting absolute speed, $U_{\text{max}}$, (from Eq. 3; assuming that $u_g$ is embedded within the observed mean current speed, $u_c$) is also shown in the bottom panels. From these time series, we see that currents dominated $U_{\text{max}}$, and current velocities were almost twice as strong as the waves at both the 40-m and 60-m isobaths. Although moderate wave-induced bed agitation took place
during, or within a day after each of the three floods, strong mean currents at both sites were only involved with the June 30/July 1 event. During that event, maximum near-bed currents, primarily setting toward the north parallel to isobaths, reached speeds of 0.55 m s\(^{-1}\) and the corresponding \(U_{\text{max}}\) reached about 0.56 m s\(^{-1}\). These strong currents did not coincide exactly with maximum river discharge (5:00, 6/30/04), but lagged the flood peak by about one day (5:00, 7/1/04). From the perspective of our data interpretation, this lag is fortunate because it allows us to separate the effects of river discharge from those of sediment resuspension by waves and currents. If a hyperpycnal discharge from the river had moved directly downslope at as little as 0.2 m s\(^{-1}\), it would have taken only about six hours to reach the 40-m tripod. Thus it is clear that an uninterrupted hyperpycnal flow from the river was not the origin of the event documented at the tripods. In what follows, we place detailed emphasis on the three-day event that began on June 30, 2004.

5.2 Depth-varying currents during a high energy flood event: ADCP results

Figure 7 displays two pairs of ADCP and PC-ADP profiles from the 60-m tripod, with the first pair collected on June 30 as the flood was rising to its peak, and the second pair collected on July 1, about 20 hours after the flood peak. (Unfortunately, the upward-looking current profiler on the 40-m tripod failed.) During the rising stage of the flood on June 30, the ADCP records (upper parts of profiles) indicate relatively weak and landward near-surface flows (surface backscatter precluded using data from the upper 3 meters of the water column). Within the bottom layer to elevations of about 4 meters above the bed, flows were weak and shoreward; there was no indication of offshore flow within the bottom layer at this time.
On July 1, as the flood was waning, shelf currents began to intensify, and at times northward flows of 0.6-0.9 m s\(^{-1}\) prevailed throughout much of the water column. These strong isobath-parallel currents were accompanied by seaward across-shelf flows that reached speeds of 0.33 m s\(^{-1}\) within the lower 4 meters of the 60 m water column. This significant offshore flow lasted for about five hours at an elevation of 6 m, but endured longer within the lower meter as detailed in the following subsection.

5.3 Near-bed currents and current profiles: ADV and PC-ADP results

Strong seaward (downslope) flows within the bottom meter of the water column are clearly evident in the acoustic Doppler velocimeter (ADV) and PC-ADP records. Figure 8 shows time series of the across-shelf and along-shelf near-bed flows as measured by ADVs. The features highlighted here are the strong seaward flows that began late on June 30 and peaked on July 1. These downslope flows reached 0.33 and 0.46 m s\(^{-1}\) at the 40-m and 60-m isobaths, respectively. At both depths, these flows were accompanied by northward flows approaching 0.4 m s\(^{-1}\). By expanding the time series for the two-day period of June 30 to July 2 (Fig. 9), it is evident that the strong seaward transport event at both the 40-m and 60-m depths consisted of two phases of about six hours duration each, separated by a brief interval of weak flow. The first phase corresponded with the seaward flows observed at higher elevations in the ADCP profiles described above. The second phase, on the morning of July 1, coincided with the time of strongest isobath-parallel flows, but was not seen at higher elevations within the ADCP record. The brief hiatus separating the two phases may have been caused by opposing tidal currents, but this is uncertain.
A striking feature evident in Fig. 9 is the downslope acceleration between the 40-m and 60-m isobaths during the second phase of the across-shelf transport event. Notably during this phase, the downslope benthic flows approached 0.5 m s\(^{-1}\) and exceeded the "supporting" along-shelf current speeds by 0.15 m s\(^{-1}\). This suggests that the observed flows may have been autosuspending for the five hour period involved. This is consistent with the fact that, locally, the bed slope was close to the critical slope necessary for autosuspension to occur (Eq. 9). We discuss this possibility in more detail later.

Details of velocity profiles from within the bottom 40 cm of the water column, as measured by the PC-ADPs on the 40-m and 60-m tripods prior to and during the transport event on July 1, are shown in Fig. 10. The ADV data points superimposed on the PC-ADP velocity profiles show reasonably good agreement between the two sensors. As is expected for velocity profiles within the bottom boundary layer, the along-shelf velocity generally increased with elevation above the bed. However, the across-shelf profiles all exhibit a clear convexity seaward with velocity maxima occurring at about 20 cm and 35 cm above the bed at 40-m and 60-m depths, respectively. This shape is highly suggestive of gravity-driven turbid flows.

However, in contrast to the thin, ~10 cm-thick, wave-supported gravity flows described by Traykovski et al. (2000), Ogston et al. (2000), and Wright et al. (2001), among others, the flows observed in this study were significantly thicker, presumably because the sediment suspension occurred within much thicker current boundary layers. As discussed near the end of Section 2, gravity flow theory suggests that if sufficient sediment is available, the slope of the Waiapu at the tripod sites is sufficiently large (\(\sin \theta\)
that it must follow that $u_g > u_w$ during gravity flow events. The convex-seaward shape is expected because below the elevation of the velocity maximum friction dominates over the downslope pressure gradient force, whereas above that elevation the driving force of $B$ diminishes upward. The overall height of these current-supported gravity flows is not entirely clear, however. Based on the combined ADCP and PC-ADP profiles displayed in Fig. 7, we estimate the thickness of the gravitationally-dominated downslope flow at 60-m depth to be at least one meter.

Another striking feature of these offshore flows is the apparent growth in flow thickness between 40 and 60-m depth suggested by the PC-ADP profiles in Fig. 10. This growth in thickness with distance offshore is consistent with the entrainment of overlying fluid made possible by a mean current boundary layer as opposed to a wave boundary layer. Local depth also plays a role, since wave orbital decay with depth causes $u_w$ to be still weaker relative to the mean current at 60 m than at 40 m. Growth in thickness is also consistent with possible autosuspension: as additional sediment is suspended from below, both the strength and thickness of the downslope current will increase.

5.4 Wave-current bed stresses

The bed shear stresses from combined waves and currents were more than adequate to sustain suspension of muds freshly delivered by the Waiapu River during floods (Fig. 12). The bed stress was estimated using two methods: the inertial dissipation method (ID) and a one-dimensional model (evolved from that of Wiberg et al., 1994). The ID method assumes that there is an inertial sub-range in the turbulence spectrum where turbulent kinetic energy (TKE) is cascaded from large to small scale eddies, and
the energy dissipation rate remains constant with frequency (Kim et al., 2000; Wright, 1995). The power spectral density of TKE, $\Phi(k)$, is a function of wave number, $k$, and energy dissipation rate, $\varepsilon$, which is related to shear velocity through $\varepsilon = \frac{u_*^3}{\kappa z}$. In the field however, velocity fluctuations are recorded as time series. To infer wave number spectra from the frequency spectra, Taylor’s “frozen turbulence” hypothesis was applied following Trembanis et al. (2004). Power spectral density of the vertical turbulent fluctuations ($w'$) were used because $w'$ is least likely to be contaminated by waves.

The input data for the 1-D model included mean current velocities, significant wave orbital velocities and wave frequency observed by the ADVs. The specified properties of sea-bed sediment included density, grain size distribution, settling velocity and critical shear stress. Analyses of grain size distributions of native sediment collected by sediment traps were carried out using a laser particle size analyzer, and the results are shown in Fig. 11. Assuming sediment density to be 2650 kg m$^{-3}$, settling velocities and critical shear stresses for each grain size were estimated from Dietrich’s empirical equation (Dietrich, 1982) and the non-dimensional Shields curve (Smith, 1977), respectively. Silts and finer particles, however, dominated the suspended sediment at both isobaths, allowing for the likelihood that much of the finer material may have been incorporated into flocs in the natural setting. Thus settling velocities less than 1 mm s$^{-1}$ and critical shear stress less than 1 dy m$^{-2}$ were set to 1 mm s$^{-1}$ and 1 dy m$^{-2}$, respectively, to better account for flocculation. Preliminary results from SWATH mapping (J. McNinch, personal communication) suggest that the seabed in the vicinity of the tripods included ripples and hummocky features. A choice of 0.1 cm for the hydraulic roughness in the 1-D model produced results for bed stress that agreed reasonably well with
observations based on the ID method from the 60-m site (Fig. 12). The peak stress, on July 1, reached 3 Pa, which is adequate to suspend all sizes of particles on the bed at 60 m, assuming that the bed had become partially consolidated prior to resuspension.

5.5 Near-bed sediment suspension

The YSI transmissometer from the 60-m tripod was calibrated using native sediment from the sediment traps. Similarly, the remaining ADV from the 40-m tripod was calibrated in the sediment chamber to relate backscatter to concentration. It was assumed that the response of the ADVs damaged during retrieval was similar. The laboratory calibration curves that appeared most consistent with the field data employed 70% mud for the YSI and 100% mud for the ADV. Most likely the true field suspensions were somewhere between these two extremes. The R-square values for the calibration curves were 0.9996 for the YSI (9 measurements) and 0.9819 for the ADV (12 measurements). Although field results from both sensor types were comparable at the right order, the ADVs were somewhat less responsive and returned lower estimates than the transmissometer. In this mixed grain size setting, we consider the transmissometer to be the more reliable instrument, since it is expressly designed to measure sediment concentration and the ADVs are not. The top panel of Fig. 13 shows the observed suspended concentration ($C_s$) time series from the transmissometer record at 91 cm above the bed (cmab) at the 60-m site.

Notably, during three high-energy events, $C_s$ exceeded 1 kg m$^{-3}$ at 1-m elevation and reached at least 3 kg m$^{-3}$ during the major event on July 1 when the maximum response range of the transmissometer was exceeded. We may assume that much higher
concentrations prevailed closer to the bed. The ADV backscatter time series at the 40-m site yielded comparable results, but did not show values as high as 3 kg m$^{-3}$ (Fig 13(b)). This may be because the highest concentration events were dominated by finer riverine sediment, and the acoustic response of the ADV is relatively less sensitive to finer sediment. Conversely, it is also possible that high sediment concentrations attenuated the ADV backscatter signal. Sediment concentration results from the 40-m isobath were less conclusive than those from 60 m because the transmissometer there malfunctioned, and we have less confidence in the ADVs as turbidity sensors.

Corresponding time series of the gradient Richardson number, $Ri$ (Eq. 4) are shown in Fig. 13b (log scale). These values were estimated using observed burst-averaged vertical gradients in maximum velocity $(u_c^2 + u_w^2)^{1/2}$ from ADVs and vertical gradients in $C_s$ based on ADV backscatter. There is appreciable scatter, reflecting, in part, the method of estimation. However, it is clear that $Ri$ was often high enough relative to the critical value of 0.25 to suggest that suspended sediment was playing a significant role in stably stratifying the bottom boundary layer.

Comparison of across-shelf velocity (at about 10 cm ab) and sediment concentration (at 91 cm ab; Fig. 13) reveals that the peak sediment concentrations were coupled with strong near-bed seaward transports. On July 1, the seaward flow of 0.35 m s$^{-1}$ coincided with the sediment concentration maximum of over 3 kg m$^{-3}$, which is further evidence suggesting that the seaward flows were related to gravity-driven hyperpycnal transport.
6. Discussion

The field evidence just described provides yet another case study in support of the notion that gravity-driven transport of turbid benthic layers is an important mechanism for transporting sediment across continental shelves, provided that externally forced oceanographic flows (currents and waves) are sufficient to support the sediment in suspension. This conclusion is not new; a growing body of recent literature makes it increasingly compelling (see review by Wright and Friedrichs, 2006). However, in earlier studies, such as the STRATAFORM study off the Eel River, the hyperpycnal layers were supported largely by waves (Traykovski et al., 2000; Wright et al., 2001). It is likely that waves were the dominant cause of sediment suspension over the inner portion (depth < 30 m) of the Waiapu shelf, although we do not have direct measurements from that region. From the foregoing, waves made only a secondary contribution to supporting gravity flows over the mid-shelf where our instruments were deployed.

6.1 The role of currents

On the basis of collective prior experience, we hypothesized in advance of our field study that the combination of depth attenuation of bottom wave orbital velocities and a slight seaward decrease in bed slope between the 40-m and 60-m isobaths on the Waiapu shelf (Fig. 3) would cause deposition (flux convergence) within that depth interval, and that downslope gravity flows would probably die out before reaching the 60-m isobath. That is not what we observed. Instead, during the three significant transport events captured by our tripod data, downslope flow velocities observed at 60 m were as strong as, and sometimes stronger than, at 40 m. This is because the hyperpycnal
conditions were sustained more by currents than by waves and those currents remained constant, or in some cases accelerated, with depth.

If our observations are representative of the long-term regime, we must infer that more sediment are bypassed to deeper water than would be the case for purely wave-supported gravity flows. This inference is compatible with the recent results of other researchers. Based on radioisotope observations, Kniskern (2007) found that most fine sediment delivered from the Waiapu River bypasses to at least to 50-m depth. Kniskern (2007) identified terrestrial pulse event layers on the mid- to outer-shelf, extending along-shelf in both directions, that are consistent with gravity flows traversing into deep water while being advected along-shelf by strong along-isobath currents. Further offshore of the Waiapu, Addington et al. (2007) found evidence of deposition by gravity flows in a trough-shaped region near the shelf edge based on Chirp sub-bottom profile analysis.

We must also infer that the equilibrium shape of the depositional shelf profile (Fig. 3) is molded by a more complex set of processes than was envisioned by the wave-dominated model of Friedrichs and Wright (2004). This shape must reflect not only the predictable depth-dependent variability of wave orbital velocities, but also the poorly understood across-shelf variations in benthic currents. Future work should include a more thorough study of the mechanisms controlling the ambient shelf currents in the region. In 3D numerical simulations of Waiapu shelf circulation, for example, Kniskern (2007) found that local wind forcing could not adequately simulate the strong observed along-shelf benthic currents. Regional flows, including the Wairarapa Coastal Current
and the East Cape Current (Stanton et al., 1998; Chiswell, 2000), which are driven in part by large-scale pressure gradients, likely play a significant role.

6.2 Ri, \( u_g \) and gravity-driven flux, \( Q_x \)

For the river-nourished muddy gravity flows discussed by Wright et al. (1999) and Scully et al. (2002), it was common for the gradient Richardson number to equal or exceed the critical value of \( \text{Ricr} = 1/4 \), implying that turbulence and associated drag were suppressed or limited by sediment induced stratification. Fig. 14 suggests that the same situation probably prevailed on the shelf off the mouth of the Waiapu River. If \( Ri \) estimated from observed concentration and velocity gradients is plotted against current velocity shear, it is seen that \( Ri \) tended to cluster around the critical value of \( Ri = 1/4 \) as velocity shear increased.

It is worthwhile to apply observed values to Eqs. (1) - (4) in order to confirm that the magnitude of the documented downslope velocities, \( u_g \), are indeed consistent with the hypothesized dynamics of current-supported gravity flows. Combining Eqs. (1) and (2) and solving for depth-averaged suspended sediment volume concentration yields:

\[
<c'> = C_D U_{\text{max}} u_g (g \cdot s \cdot h \cdot \sin \theta)^{-1}.
\]

Assuming depth-averaged \( U_{\text{max}} \) and \( u_g \) are on the order of 0.4 and 0.3 m s\(^{-1}\), respectively (based on Fig. 10), \( C_D = 0.003 \), \( g = 9.8 \) m s\(^{-2}\), \( s = 1.65 \), \( h = 1 \) m, and \( \sin \theta = 0.01 \), gives \( <c'> = 0.0022 \), equivalent to an average mass concentration of 5.8 g l\(^{-1}\) over the bottom meter of the water column. Given that 3 g l\(^{-1}\) was observed by a transmissometer 91 cm above the bed, and considering the "lowest order" status of the above equations, an average value of 5.8 g l\(^{-1}\) over the lowest meter seems quite reasonable.
Alternatively, one can solve Eq. (4) for depth-averaged concentration, assuming the bottom boundary layer is critically stratified:

\[ <c'> = Ri_{cr} U_{max}^2 (g \cdot s \cdot h)^{-1} \]  

(12)

With \( Ri_{cr} = 1/4 \), and other values as applied to (11) above, Eq. (12) then yields a depth-averaged concentration of 6.6 g l\(^{-1}\), which again is reasonably consistent with an observed value of 3 g l\(^{-1}\) near the top of the layer. Thus the above described observations of current-supported gravity flows off the Waiapu appear to support continued application of many aspects of the basic dynamics that were first applied to wave-supported gravity flows by Wright et al. (2001). Current-supported gravity flows are spread over a greater thickness compared to the wave boundary layer, causing the depth-averaged concentrations to be lower. Nonetheless, the total depth-integrated sediment loads appear to be comparable.

Following Wright et al. (2001) for those cases where an unlimited supply of fine sediment allows the maintenance of \( Ri \equiv Ri_{cr} \), the limiting gravity-induced velocity, \( u_{gcr} \) is

\[ u_{gcr} \approx \frac{\sin \theta \cdot Ri_{cr} U_{max}}{C_D} \] 

(13)

The implications of Eq. (13) are that, when \( Ri \approx Ri_{cr} \), \( u_{gcr} \) is proportional to, rather than retarded by, \( U_{max} \). This is because the "unlimited" availability of suspended sediment permits increases in total bed stress associated with \( U_{max} \) to be accompanied by increases in sediment-induced stratification, which suppresses turbulence and thereby reduces the frictional resistance of the plume.

Substituting \( U_{max} \) for wave orbital velocity in the expression suggested by Wright et al. (2001) for downslope sediment transport, one can then solve for the
depth-integrated across-shelf sediment load per unit length along-shelf associated with a critically-stratified wave-or current-supported gravity flow, \( Q_{cr} \):

\[
Q_{cr} = \rho_s (\sin \theta) Ri_c^{2} U_{\text{max}}^{3} (gsC_p)^{-1},
\]  

where \( \rho_s = 2650 \text{ kg m}^{-3} \) is sediment density. Application of Eq. (14) for the event of July 1, 2004 at 40 m (\( \sin \theta = 0.012 \)) and 60 m (\( \sin \theta = 0.010 \)) with values otherwise matching those applied to (11) and (12) gives \( Q_{cr} \sim 2.6 \text{ kg m}^{-1} \text{ s}^{-1} \) and \( 2.2 \text{ kg m}^{-1} \text{ s}^{-1} \) at the two depths, respectively. This suggests that, assuming no autosuspension, deposition would have been expected between the two isobaths even though significant across-shelf flux continued beyond 60 m. However, Eq. (14) is highly sensitive to \( U_{\text{max}} \), so it is also possible that higher peak values of \( U_{\text{max}} \) at 60 m may have prevented convergence of across-shelf sediment flux. Three-dimensional spreading of the plume may also have played a role. Alternatively, a transition to autosuspension could have helped prevent transport convergence (see below).

It is interesting to note that strong downslope velocities tended to coincide with strong up-coast along-shelf currents, which is consistent with bottom Ekman layer veering in the southern hemisphere. The ratio of the depth-integrated Coriolis acceleration induced by the along-shore flow relative to the downslope buoyancy-induced pressure gradient is given by

\[
\frac{\text{Coriolis}}{\text{Buoyancy}} = f v_c / (gs \langle c' \rangle \sin \theta).
\]  

With \( f = 8.9 \times 10^{-5} \text{ s}^{-1} \), \( v_c = 0.3 \text{ m s}^{-1} \) (averaged over the lowest meter), and the remaining values as in (11), Eq. (15) predicts the magnitude of Coriolis acceleration to be only about 8% of the downslope buoyancy force during the peak of the event observed at 60 m on July 1. Thus the downslope velocity was probably dominated by gravity-induced flow.
rather than rotation effects. Nonetheless, Coriolis may still play a role in helping to initiate downslope flows which then evolve into gravity flows during times of strong up-coast flow.

6.3 Transition to autosuspending turbidity flows

As pointed out in Section 5.3, the fact that speeds and bed shear stresses associated with downslope flows exceeded those of the combined waves and along-shelf currents at the 60-m isobath during the second phase of the July 1 transport "event" suggests that flows may have been autosuspending at that time. The apparent growth in the thickness of the downslope flow between 40 and 60 m is also consistent with autosuspension. Some degree of autosuspension is conceivable considering that the bed slope over the depth region of 30 m - 55 m is close to the critical minimum, \( \theta_{\text{auto}} \), required to sustain downslope transport via autosuspension. Although autosuspension may have contributed significantly to downslope acceleration of gravity-driven flows, support by current and wave induced bed stresses in shallower regions of the shelf was undoubtedly implicated in the initiation of the flows.

Conceptually, we can imagine an across-shelf progression of the dominant forcings responsible for maintenance of hyperpycnal near-bed conditions (Fig. 15). In the immediate vicinity of the river mouth, the hyperpycnal outflow itself is likely to dominate but, without waves and currents, hyperpycnal river effluents would experience very rapid extinction over low gradient inner shelves. Wave support over the inner shelf can be expected to transition to combined wave and current support and eventually to current-dominated support as the flows move into progressively deeper water. In the
Waiapu case, the current-supported flows were, at least theoretically, able to transition to autosuspending flows on the mid-shelf because the critical slope, \(\theta_{auto}\), occurred at a relatively shallow depth. This was not the case on the mid-shelf off the Eel River.

6.4 Implications for modeling subaqueous clinoform morphodynamics

The Waiapu shelf is not unique in being influenced by strong non-tidal parabathic currents. Such currents are less predictable not only because they do not follow the highly deterministic behavior of astronomical tides, but also because they are not always directly correlated with, or forced by, wind stresses at the sea surface. Depending on local circumstances, they may be baroclinic, they may be manifestations of episodic shoreward intrusions or eddy shedding of large scale ocean currents (e.g. the Gulf of Mexico Loop Current), or they may be related to the propagation of continental shelf (Rossby) waves. Recent studies have shown that the continental shelf off the Mississippi Delta is influenced by all of the above in addition to more “well behaved” locally-forced wind driven currents (e.g. Schmitz et al., 2005; Jarosz and Murray, 2005).

Our observations over the mid-shelf off the Waiapu River clearly show that along-shelf currents there dominate the support of turbid near-bed layers and probably control the flux gradients associated with those layers. These currents also facilitate the transition of gravity-driven transport from a wave-supported inner shelf regime to an autosuspending regime in deeper water. In this way, the currents presumably control the morphodynamic development of the deeper parts of the Waiapu submarine deposits. What we do not know is why the observed currents behaved the way they did. The physical oceanography of that shelf is only superficially described and understood. If we
wish to develop realistic models of the depositional morphodynamics of the Waiapu shelf or other current-swept shelves like it, we must first get the full picture of the physical oceanographic regimes involved.

7. Summary and conclusions

The results just described add further evidence to the growing body of literature indicating that gravity-driven transport within near-bed turbid layers is a fundamental mode of across-shelf sediment dispersal on river-nourished shelves. The Waiapu data also suggest that hyperpycnal conditions within the river itself are not sufficient to cause benthic transport beyond the inner shelf. The major points from this study can be summarized as follows.

• Sediment discharged by the Waiapu River during floods were inferred to be rapidly transported offshore and to the north within both surface and bottom layers.

• Seaward flow within the surface layer coincided with maximum river discharge and is inferred to have been related to a hypopycnal plume.

• The most significant seaward transport of sediment within the turbid near-bed layer did not coincide with the time of maximum sediment discharge from the river mouth but rather with the time of maximum current-induced bed stress.

• The strong seaward flows within the bottom layer also coincided with the time of the highest observed near-bed suspended sediment concentrations.

• Although current-supported, the resulting downslope velocities and concentrations were still reasonably consistent with the dynamics of
critically-stratified sediment gravity flows as previously applied to wave-supported cases.

- High bed slopes and strong ambient currents both contributed to gravity flows on the Waiapu being significantly thicker than those previously observed in wave-supported cases over more gently sloping shelves.

- Because current-supported gravity flows are thicker than wave-supported flows, a similar load for a given ambient velocity results in proportionately lower suspended sediment concentrations.

- The increasing thickness and velocity of the gravity flow with distance offshore along with generally high bed slope suggest that autosuspension may have contributed to downslope transport over the steeper regions of the mid-shelf.

- The observed benthic across-shelf flows carried suspended sediment seaward beyond the 60-m isobath; no clear evidence of flux convergence was seen between 40 and 60 m during the event period;
Acknowledgements

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### Table 1. Summary of tripod instrumentations.

<table>
<thead>
<tr>
<th>Instruments</th>
<th>40-m Tripod</th>
<th>60-m Tripod</th>
<th>Function</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sontek Acoustic Doppler Velocimeter (ADV)</td>
<td>18 cmab</td>
<td>26 cmab*</td>
<td>Measures 3-D turbulence at fixed elevations above bed</td>
</tr>
<tr>
<td></td>
<td>40 cmab</td>
<td>42.5 cmab</td>
<td></td>
</tr>
<tr>
<td>Sontek Pulse-Coherent Acoustic Doppler Profiler (PCADP)</td>
<td>67 cmab</td>
<td>70 cmab</td>
<td>Downward-aimed, measures velocity profiles very near the bed</td>
</tr>
<tr>
<td>RDI Acoustic Doppler Current Meter (ADCP)</td>
<td>No</td>
<td>217 cmab</td>
<td>Upward-aimed, velocity profile through water column</td>
</tr>
<tr>
<td>Sontek Acoustic Doppler Profiler (ADP)</td>
<td>230 cmab*</td>
<td>No</td>
<td>Upward-aimed, velocity profile through water column</td>
</tr>
<tr>
<td>Aquatec Acoustic Backscatter Sensor (ABS)</td>
<td>85 cmab*</td>
<td>62 cmab*</td>
<td>Near-bed suspended sediment concentration</td>
</tr>
<tr>
<td>YSI Water quality sensor</td>
<td>41 cmab*</td>
<td>91 cmab</td>
<td>Turbidity, temperature and salinity</td>
</tr>
<tr>
<td>Sediment Trap</td>
<td>87 cmab</td>
<td>46 cmab</td>
<td>Collecting near-bed suspended sediment</td>
</tr>
<tr>
<td></td>
<td>88 cmab</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* Failed during deployment.
Figure 1. Site map of the Waiapu shelf, North Island, New Zealand, showing locations of tripod deployments at 40-m and 60-m isobaths (solid dots). Throughout the paper, along-(v) and across-shelf (u) velocities correspond to flows parallel and perpendicular to the local bathymetric contours. Star and triangle in the left lower panel indicate locations of wind and river discharge measurements.
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Figure 4. General configuration of the tripods deployed on the Waiapu shelf.
Figure 5. Conditions during the observation period: time-series of (a) surface wind speed and direction (meteorological convention), (b) Waiapu River discharge, (c) total suspended sediment concentrations within the lower course of the river, and (d) r.m.s. wave height as determined from the 40-m ADV records. Wind and discharge data were provided by the Gisborne District Council (20 km upstream from the river mouth). Sediment concentration is based on the sediment rating curve of Hicks et al. (2004).
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Figure 7. Across-shelf and along-shelf flows from ADCP (upper part) and PC-ADP (lower part) from the 60-m tripod. The first pair of profiles was collected on June 30 as the flood was rising to its peak (— & ….). The second pair was collected on July 1, about 20 hours after the flood peak ( - - - & -- -- --).
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Figure 12. Time-series of bed shear stress estimated by (a) the inertial dissipation (ID) method and (b) a one-dimensional boundary layer model.
Figure 13. Time series of (a-b) suspended sediment concentration based on calibrated YSI transmissometer records (91 cmab) at the 60-m isobath and ADV backscatter readings at the 40-m isobath (solid and dashed lines indicate 18 and 40 cmab respectively), (c) gradient Richardson number estimated using measured velocity and backscatter data from 40-m ADVs, and (d-e) very near bed (z ~10 cm) across-shelf mean velocity from PC-ADP at 60- and 40-m water depth, respectively. (Only the 60-m tripod had an operational transmissometer for accurately estimating fine sediment concentration, and only the 40-m tripod had two operational ADVs for roughly estimating Ri.)
Figure 14. Variation of gradient Richardson number with absolute velocity shear calculated at 40-m isobath.
Figure 15. Conceptual model showing the across-shelf progression of the dominant forcings responsible for maintenance of near-bed hyperpycnal flows off of the Waiapu River. Tripod locations are indicated by triangles.
Chapter 3: Current-supported Gravity Flows on the Continental Shelf offshore of the Waiapu River, New Zealand: Observations and Modeling

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Abstract

Tripod observations and model calculations reveal that current-supported gravity flows dominated across-shore sediment dispersal on the Waiapu continental shelf during a winter field experiment in 2004. Three transport events were identified based on waves, currents, and sediment concentration data obtained from two tripods that were deployed on the Waiapu middle shelf at 40- and 60-m water depths. Each event involved two distinct phases: a flood phase with fluvial sediment input and a resuspension phase during which sediment were mobilized. The flood phases were characterized by large sediment input from the Waiapu River, coupled with moderate to strong waves but weak currents. No strong sediment signals were recorded by the tripods during these flood periods. During the post-flood resuspension phases, the tripods both observed strong seaward near-bottom turbid flows, with down-slope velocities up to 0.5 m/s. A one-dimensional boundary layer model supported the inference that these flows were dynamically similar to wave-supported gravity flows observed on Eel and Po River shelves, except that currents rather than waves were responsible for sediment resuspension. In contrast to thin (~ 10 cm) and dense (c_s >10 kg/m^3) wave-supported gravity flows, the current-supported gravity flows on the Waiapu shelf were significantly thicker (1~2 m) and more dilute, with depth-averaged concentrations of 2 to 4 kg/m^3. Total depth-integrated sediment fluxes within the turbid layers were on the order of magnitude of 10^5 kg/m over the course of the three events. Most sediment flux occurred during resuspension phases, induced by current-supported gravity flows.
Keywords: Waiapu shelf; New Zealand; Current-supported gravity flow; Across-shelf sediment transport; Bottom boundary layer; Flood event

1. Introduction

During the past decade, gravity flows have increasingly interested scientists studying sediment transport across continental shelves. In contrast to classical turbidity currents, the gravity flows discussed here rely on velocity shear generated by ambient waves and/or currents to maintain sediment in suspension. Moreover, to sustain a gravity flow, the steepness of a continental shelf need not exceed 0.012, the minimum shelf slope at which a turbidity current can maintain itself via auto-suspension (Wright et al., 2001; 2002).

Sediment gravity flows were first documented on the Amazon delta front by Sternberg et al. (1996) as ‘downslope migration of fluid muds’ supported by ‘boundary shear stress generated by waves and currents’. Such fluid mud migrations, however, were not fully described until they were observed as part of the STRATAFORM program on the Eel River continental shelf (Ogston et al., 2000; Traykovski et al., 2000). Facilitated by high-resolution boundary-layer-measurements within 10 cm of the bed, Traykovski et al. (2000) discussed the important role that wave-induced density-driven flows played in cross-shelf sediment transport on the Eel shelf. Wright et al. (2001, 2002) introduced a simple analytic theory of gravity flow dynamics and related it to previous field data from the Gulf of Bohai offshore of the Yellow River, the Eel shelf, the Louisiana shelf west of the Mississippi River, and the Middle Atlantic Bight offshore of Duck, NC.
Since then, research has continued to highlight the significance of sediment gravity flows on the transport of fine sediment across continental shelves. For example, Scully et al. (2002, 2003) extended gravity flow theory to include analytical expressions for deposition, and elucidated the importance of bathymetry to the formation of flood deposition on the Eel shelf. Friedrichs and Wright (2004) further demonstrated that the equilibrium bathymetric profiles offshore of many river mouths were highly related to gravity-flow transport. The subsequent Euro STRATAFORM project, conducted in the Adriatic Sea, showed that wave-supported gravity flows also influenced flood deposition offshore of the Po River mouth, and south-eastward sediment transport on the Po subaqueous delta (Friedrichs and Scully, 2007; Traykovski et al., 2007; Harris et al., 2008). In addition to direct field observations, many numerical models have been developed or modified to address the importance of gravity flows in sediment transport, including one-dimensional models by Traykovski et al. (2007), two-dimensional models by Scully et al. (2003, Eel shelf), Friedrichs and Scully (2007, Po shelf), and Hsu et al. (2006), as well as a three-dimensional model applied to the Eel (Harris et al., 2004; 2005) and Waiapu River shelves (Kniskern, 2007).

The prerequisites for gravity-driven flow to be important include i) sufficient sediment available for resuspension, ii) strong wave- or current-generated turbulence to maintain sediment in suspension, iii) a sloping seabed to produce a sufficient downhill gravity force. Most of the collective studies of sediment gravity flows on Eel and Po shelves have focused on wave-supported cases, because waves were the dominant turbulence-generating mechanism. Current-supported gravity flows, on the contrary, have seldom been documented. Thus, tripod observations along with coring studies on the
Waiapu shelf in the winter of 2004 broke new ground in the study of gravity flows and sediment transport processes. Focusing on a two-day sediment resuspension event, Ma et al. (2008) analyzed tripod data and showed direct evidence of current-supported gravity flows on the Waiapu shelf. This paper further explores the significance of current-supported gravity flows by (i) analyzing field data from two tripods (40- and 60-m depth) to describe the oceanographic conditions during a 45-day period, (ii) characterizing three flood and transport events, (iii) applying a one-dimensional sediment transport model to estimate the velocity and thickness of gravity flows, and (iv) identifying the mechanisms of cross-shelf sediment transport during three events with respect to the relative importance of current- and wave-supported gravity flows as compared to dilute suspended flows.

2. Site description

The Waiapu River (Fig. 1) is located on the east coast of the North Island of New Zealand. With headwaters in the Raukumara Range, it drains northeastwards through a mountainous and tectonically active landscape, and eventually empties into the Pacific Ocean (Eden et al., 2001). The Waiapu drainage basin receives high rainfall (~2.4 m/yr), has a steep channel gradient given the head water elevation of ~1500 m and short channel length of 130 km, and contains easily erodible Tertiary mudstone. All of these lead to one of the world’s highest sediment yields of ~20,000 T/km²/yr. Following European settlement in the late 1820s, the indigenous forest was widely cleared, dramatically increasing erosion and sediment transfer. Nowadays, the Waiapu annually supplies about $35 \times 10^6$ tonnes of suspended sediment to the ocean (Hicks and Griffiths,
1992; Hicks et al., 2000), which accounts for about 0.2% of the total sediment discharged globally by rivers (Milliman and Syvitski, 1992). Most of the sediment is debouched into the ocean during storm events (Hicks et al., 2004), and the Waiapu typically produces several hyperpycnal river discharge events during a typical winter flood season (June - September). Due to the small size of the catchment (~1700 km²), the river-ocean system is strongly coupled in that the weather systems that bring heavy rains to the river catchment also create strong waves and wind-driven currents on the shelf.

The Waiapu River dispersal system lies on an active margin. The shelf is narrow in width (~ 20 km) and steep in slope. The bathymetric profile (Fig. 2) of the shelf is characterized by two slight humps centered at the 35- and 100-m isobaths, respectively. Near the river mouth, the slope of the subaqueous prodelta (0~30-m water depth) is slightly gentler than the critical slope of 0.012 required to sustain autosuspending turbidity flows (Wright et al., 2001). Moving into deeper water the seabed steepens, with a slope of 0.01-0.013 between 30-60 m isobath. Seaward of this the seafloor becomes relatively flat until the 100-m isobath where the slope steepens and again exceeds the critical value of 0.012. A striking sedimentary feature of the Waiapu shelf is the Holocene mud wedge resting in the mid-shelf basin with maximum sediment thickness reaching 110 m around 110-m isobaths (Lewis et al., 2004). This significant Holocene deposition has created a convex-upward across-shelf profile as is commonly seen over a progradational clinoform shelf. Studies of sediment cores and Chirp sub-bottom profiles revealed that high-density event layers were well preserved throughout the entire shelf (except the northern rocky area as indicated by Wadman et al., 2008) as well as further offshore on the shelf edge (Kniskern, 2007; Addington et al., 2007).
Oceanographic forces on the Waiapu shelf are very energetic, especially during storms. The inner shelf is subject to the northward flowing Wairarapa Coastal Current (WCC) (Chiswell, 2000), while the outer shelf and upper slope are swept by the southward flowing East Cape Current (Stanton et al., 1998). The prevailing waves along the east coast of New Zealand arrive from the southeast, probably due to the refraction of the easterly Antarctic Circumpolar Current. Based on newly-collected tripod data located on the 40- and 60-m isobaths offshore of the Waiapu River mouth (Ma et al., 2008), typical significant wave height was on the order of 1.3 m with a mean period of 11.4 s. During regular storm periods, significant wave height exceeded 3-4 m with typical wave period of 11 s. These storm waves were able to penetrate an 80-m thick column of water to resuspend fresh muddy sediment on the seabed. Observed currents on the Waiapu continental shelf were fairly strong. At the 60-m tripod location, mid-water current speeds reached 1.3 m/s, and near-bed current speeds was as high as 0.5 m/s. Local wind forcing was significant, especially during high energy events. At 60-m depth, about half of the variance in the near-bed along-shelf current was correlated with hindcast along-shelf wind stress (Chapter 4).

Tripod data analyzed by Ma et al. (2008) provided the first direct evidence that the seaward flows on the Waiapu shelf were related to gravity-driven hyperpycnal transport. Unlike processes observed offshore of the Eel and Po Rivers, currents appeared to play as important a role in supporting the gravity flows here, as did waves. High and episodic sediment loads, the steep slope of the shelf, and energetic marine conditions together made the Waiapu area an ideal site to study current-supported gravity flows on the continental shelf.
3. Field observations and results

A field investigation was carried out on the Waiapu shelf during the storm season (May-October) of 2004. Two instrumented tripods (Fig. 3) were deployed on the 40- and 60-m isobaths of the Waiapu shelf directly offshore of the river mouth to observe hydrodynamics and sediment transport. Each tripod was equipped with one upward-looking Acoustic Doppler Profiler (ADP, 40-m tripod) or Acoustic Doppler Current Profiler (ADCP, 60-m tripod) to monitor the mean currents in the water column from 4.21 m above the bed (mab) to the sea surface. The 300-kHz ADCP measured mean current in 10-min bursts every hour with sampling frequency of 1 Hz and cell size of 2 m. The 1.5 MHz ADP recorded data in 15-min bursts every half hour with cell size of 1 m.

Two Acoustic Doppler Velocimeters (ADVs) were placed on the extension arm of each tripod to measure three-dimensional turbulent velocities and acoustic backscatter intensity. These data were sampled at 5 Hz for 5-min bursts every two hours and had a single point sampling volume of 10 cm$^3$. A downward-looking Pulse Coherent-Acoustic Doppler Profiler (PC-ADP) was located above the ADVs on the same arm to observe the bottom boundary layer velocity profile. The PC-ADP measured mean velocity at 1 Hz in 15-min bursts every hour, and with sampling bins of 4.7 cm, much smaller than that of the upward-looking ADCP. Each tripod also held one YSI that measured salinity, temperature and near-bed turbidity; one sediment trap that collected suspended sediment samples; and one Acoustic Backscatter Sensor (ABS) that recorded near-bed suspended sediment concentration profiles.
Some instruments failed to operate during the deployment or were damaged during deployment or retrieval (Table 1), including both ABSs, and the ADP and YSI at the 40-m site. A few instruments malfunctioned after July 2004, such as the PC-ADP and ADV located on the 60-m tripods. Because the data obtained before the middle of July were most reliable, in this paper we present observed data for June-July (45 days) and focus on events during this two-month period.

3.1 YSI calibration

To estimate sediment concentration, turbidity readings from the YSI were calibrated in a laboratory using native suspended material collected by sediment traps mounted 46 cm above bed (cmab) on the 60-m tripod. The calibration was executed in a 2-m high chamber equipped with a motor to maintain a well-mixed water column. Sediment samples were added to the fluid gradually to increase its concentration. For each measurement, fluid samples were collected through a tube inserted into the chamber wall, and were filtered to measure concentration. This was then compared to the YSI reading (Fig. 4) to obtain a linear calibration between the turbidity readings and sediment concentrations ($R^2$ as high as 0.99).

A possible source of error in this calibration was that strong turbulence within the chamber might disaggregate flocculated particles and increase the turbidities measured by the YSI because optical instruments are more sensitive to finer materials (Traykovski et al., 2000). However, during periods of high bed stress when most of sediment transport occurred, waves and/or currents likely generate intense turbulence that breaks down large fragile aggregates. Another potential problem in applying this calibration was that this
YSI was maxed out at concentrations above about 3 kg/m$^3$. It tended to underestimate sediment concentrations if concentrations were actually higher than this critical value for example during event periods.

3.2 River discharge and oceanographic forces

Three floods, with durations of two to three days, occurred during June and July (around June 5, June 19 and July 1) (Fig. 5a). High near-bed suspended sediment concentrations, however, were observed one to four days after flood peaks at the 60-m tripod (Fig. 5b). Those time lags were much longer than the estimated six hours needed for the river plume to travel from the river mouth to the tripod site (Ma et al., 2008). Three events (E1, E2 and E3, shaded areas) were delineated based on the river hydrograph and tripod data. Each event began when water discharge increased and ended when sediment concentration fell back to its normal value. Each event was further divided into two phases: Flood (F) and Resuspension (R). This definition tended to separate initial delivery of sediment by the river from the remobilization of sediment by oceanographic forces. Root Mean Square (RMS) wave heights and near-bed current speeds show that both waves and currents were strong during the three events, especially during the resuspension phases (Fig. 5c). The “flood” periods of the events were associated with low sediment concentrations at the tripods, though these were the times when sediment was delivered to the shelf. Interestingly, all of the resuspension phases were associated with high sediment concentrations at the tripods and strong seaward across-shelf currents (Fig. 5d). Table 2 summarizes the event characteristics.
A medium intensity flood occurred on June 5, 2004 having peak water discharge of about 670 m$^3$/s. High waves and strong currents, however, did not impact the shelf until three days later, during which time the bottom across-shelf currents turned seaward, and the YSI, located 91 cmab on the 60-m tripod, observed sediment concentrations as high as 2.5 kg/m$^3$. Later, the Waiapu experienced two major floods on June 18 (E2) and June 30 (E3), respectively. During these two discharge pulses, Waiapu water discharge exceeded 2000 m$^3$/s, more than twenty times of the annual mean value (85 m$^3$/s), and total sediment concentration within the lower course of the river was estimated to have reached 50 kg/m$^3$ (Ma et al., 2008), making the river water distinctly hyperpycnal. The flood phase E2-F was characterized by 2.5-m high waves and landward bottom currents. Sediment concentrations increased four days later during the resuspension phase, E2-R, that included medium waves (1.5 m) and strong seaward currents. The third flood, the biggest one during the tripod deployment period, was addressed in detail by Ma et al. (2008) (Chapter 2). During that event, waves dominated the marine forces in E3-F, but extremely strong seaward currents characterized the resuspension phase (E3-R) with velocities as high as 0.46 m/s. In the following analysis we characterized these three events and examined possible mechanisms that influence sediment dispersal on the Waiapu shelf.

3.3. Velocity profiles

Currents throughout the water column are represented in Fig. 6 by velocity vectors observed at the water surface (5 m below the sea surface), in mid-water (30 mab), in the bottom boundary layer (4.2 mab) and near the bed (48 cmab). The surface currents
correlated reasonably with the winds measured at Hicks Bay with $R^2$ of about 0.55. Currents in the upper and lower water column tended to be veered by Ekman spiral to the left and right respectively, as expected in the southern hemisphere. The maximum current exceeded 1.3 m/s in the surface of the water column, while the near-bed current exceeded 0.5 m/s (60-m PC-ADP ceased recording after July 9, 2004). Although the wind direction was highly variable, westerly winds appeared to dominate all flood event periods.

The along- and across-shelf velocity profiles during event periods at the 60-m site are shown in Fig. 7. Each profile was averaged over the six sub-events and separated into two segments: near-bed and upper-water profiles measured by the PC-ADP and ADCP, respectively. The lowest bin of ADCP was 4.2 mab and the uppermost bin of 60-m PC-ADP was 48 cmab. Unlike the relatively calm conditions seen during flood phases, all of the resuspension phases were characterized by strong along- and across-shelf flows. By visually connecting the two segments of the velocity profiles, zones where seaward flows were highest are estimated to be at about 40 cmab, with weaker offshore flows above and below. Detailed near-bed across-shelf velocity profiles for both the 40- and 60-m tripods (Fig.8) clearly exhibited the convex-seaward shapes during resuspension phases, while the along-shelf velocity profiles notably did not. Convex-seaward velocity profiles suggest gravity-driven turbid flows as recently described by Traykovski et al. (2000, 2007), Ogston et al. (2000), Wright et al. (2001), and Hsu et al. (2006), among others. The turbid flows documented here, however, were at least 50 cm, much thicker than those expected for wave-supported flows (~10 cm).
3.4 Near-bed conditions

Fig. 9 shows the conditions observed near the bed at the 40- and 60-m tripods, including time series of representative bottom wave periods \( T \), RMS amplitude of orbital velocities \( u_w \) and near-bed along- \( v \) and across- \( u \) shelf mean currents. Wave period, \( T \), was determined from ADV velocity records via power spectral analysis following Madsen (1994). The mean period was about 11.4 s, representing long-period swell. With this typical period, 1.5- and 2.5-m high waves are strong enough to resuspend muddy sediment from the sea bed at 40 and 60-m water depths, respectively. Wave orbital velocity was calculated as \( u_w = \sqrt{2 \langle (u-<u>)^2 +(v-<v>)^2 \rangle} \), where \( <> \) indicates a burst average, and \( u \) and \( v \) are instantaneous horizontal velocities recorded by the ADV (Wiberg and Sherwood, 2008). Wave orbital velocities ranged from 0.02 to 0.25 m/s at both sites, with the averaged orbital velocity on the 40-m isobath slightly higher than that seen at the 60-m isobath. Near-bed current velocities were almost twice as high as \( u_w \) at both sites, with a maximum speed of 0.55 m/s occurring at the 60-m site.

3.5 Bed shear stress

Bed shear stress, \( \tau_s \), is a critical parameter that influences sediment resuspension. On continental shelves, total bed stress \( \tau_{cw} \) is the sum of current- and wave-induced components, \( \tau_c \) and \( \tau_w \), respectively, each of which influences the other (Grant and Madsen, 1986; Smith, 1977). The bed stress above the wave boundary layer \( \tau_c \) can be estimated directly with field tripod data in several ways (Sherwood et al., 2006; Kim et al., 2000). In this paper the current-induced bed shear stress was calculated using two
methods: the covariance method (COV) based on ADV readings (Grant and Madsen, 1986), and the log profile method (LP) based on PC-ADP velocity profiles (Soulsby, 1983; Dyer and Soulsby, 1988). As discussed below, each of these methods has limitations and at times provides very different values. Following discussion of each method, we described a criteria used to derive our final estimate of current-induced bed stresses from these measurements. The total bed stress including waves (\( \tau_{cw} \)), was then calculated by a two-dimensional boundary layer model based on the roughness selected by matching the observed and modeled estimates of \( \tau_c \).

3.5.1 Log Profile (LP) method

Based on the ‘law of the wall’, the mean velocity near a seafloor boundary varies linearly with the logarithm of elevation and can be expressed by Von Karman-Prandtl equation as \( u = \frac{u_*}{\kappa} \ln \left( \frac{z}{z_0} \right) \), where \( \kappa \) is the von Karman constant (\( \sim 0.41 \)), \( z_0 \) is hydraulic roughness, and \( u_* \) is shear velocity (\( u_* = \sqrt{\tau_b / \rho} \)). Mean velocities measured at several elevations within the bottom boundary layer can be used to fit a semi-logarithmic regression of velocity versus elevation. By applying a linear least-square fit to the data, one can determine \( u_* \) as the inverse of slope of the velocity profile. If the measurements are above the wave boundary layer, but within the log-layer, the shear velocity can be used to calculate the current-induced bed shear stress via \( \tau_c = \rho u_*^2 \). Limitations of this method include that i) there might be poor correlation coefficient for the semi-log regression; ii) all measurements should be within the log-layer, which may be thin; iii)
velocity acceleration and sediment stratification tend to add curvature to the log-linear profile within the bottom few meters of the water column.

Velocity data from PC-ADPs deployed at the 40- and 60-m sites were analyzed using this method. Four near-bed data points, at ~13, 18, 23, and 28 cmab, were used to determine the velocity profile for each burst; and any profiles for which the correlation coefficient \( R^2 \) was less than 0.8 was discarded. The largest source of error in applying this method was that it may overestimate \( \tau_c \) when suspended sediment concentration is high; because sediment induced stratification tended to dampen turbulent mixing and added a convex shape to the velocity profile. The upper panels in Fig. 10 (gray lines) display the calculated \( \tau_c \) for 40- and 60-m water depths. Average bed shear stress at the 40-m site was 0.078 Pa, while it was 0.083 Pa at the 60-m site. For some events, shear stress was higher at the 60-m site because of a stronger current. Shear stresses based on these regressions were especially high during the event periods.

3.5.2 Covariance (COV) method

Assuming the sampling volume is close enough to the seabed, measured turbulent velocities can be used to derive a Reynolds' stress estimate of bed shear stress. The measured instantaneous velocity vector \( \mathbf{u} \) can be separated into mean \( \langle \mathbf{u} \rangle \) and turbulent \( \mathbf{u}' \) velocity components such that \( \mathbf{u} = \langle \mathbf{u} \rangle + \mathbf{u}' \). The COV method estimates bed shear stress, i.e. the Reynolds stress, utilizing direct measurement of those turbulent velocities as

\[
\tau_{zx} = -\rho \langle u'w' \rangle, \tag{1}
\]

\[
\tau_{zy} = -\rho \langle v'w' \rangle, \tag{2}
\]
where $u$, $v$ and $w$ are horizontal and vertical velocities. The upper ADVs mounted on the 40- and 60-m tripods at elevations of 40 and 42.5 cmab were used in this method because they provided long and reliable records. A problem with this approach is that even small tilts in the velocity sensor can allow wave orbital motions to contaminate the estimate of turbulent fluctuations and skew Reynolds stress estimates (Shaw and Trowbridge, 2001). When strong waves are present, instantaneous velocity vectors should be rewritten as $\mathbf{u} = <\mathbf{u}> + \mathbf{u}' + \mathbf{u}_o$, etc., where $\mathbf{u}_o (u_o, v_o, w_o)$ represents the wave orbital component. Using a residual velocity to estimate the turbulent velocity will result in an overestimated Reynolds stress by mistakenly including wave energy in the turbulent component as $(\mathbf{u}' + \mathbf{u}_o)$. If the sensor is perfectly aligned with the vertical, the non-turbulent vertical velocity components ($<w>$ and $w_o$) should be very small, and the orbital components will not corrupt the Reynolds stress estimates. Any small tilt in the sensor, however, will cause some of the horizontal orbital velocity ($u_o$ and $v_o$) to appear as an apparent vertical orbital velocity component ($w_o$). When the residual velocities are then time-averaged in the covariance method (Eqs. (1) - (2)), this adds a non-zero term as $<u_o w_o>$ that causes the Reynolds stress estimate to be overestimated whenever waves are large. Another source of anomalously high apparent turbulent velocities could be wobbling of the instrument itself, which could also induce spurious vertical velocities in the recordings.

To apply the covariance method, ADV measurements were therefore rotated in an attempt to correct for instrument tilting via the method of Kim et al. (2000):

$$\mathbf{u}_r = [R] \mathbf{u},$$

(3)

where $\mathbf{u} (u,v,w)$ is velocity vector measured by the ADV in instrument coordinate system, and $\mathbf{u}_r (u_r,v_r,w_r)$ is velocity vector after rotation and $[R]$ is a rotation matrix:
\[
[R] = \begin{bmatrix}
\cos \theta \cos \beta & \cos \theta \sin \beta & \sin \theta \\
-\sin \beta & \cos \beta & 0 \\
-\sin \theta \cos \beta & -\sin \theta \sin \beta & \cos \theta
\end{bmatrix}.
\]

The horizontal rotation angle \( \theta \) was chosen to minimize the variance of \( v \), thereby aligning the rotated vectors with the dominant horizontal velocity. Next, the vertical angle, \( \beta \), was chosen to minimize the variance of \( w \), so that the instrument vertical axis aligns with the real vertical direction where the vertical energy (square of variance of \( w \)) was the least. The vertical angle \( \beta \) ranged from \( 11^\circ \) to \( -11^\circ \), clustering around \( 0^\circ \), while \( \theta \) was more scattered, with the majority (~80%) between \( 180^\circ \) and \( 0^\circ \). Burst averages were then calculated from the rotated vectors and the residual values applied in equations (1) and (2) to estimate bed shear stress.

For the 60-m site, bed shear stresses estimated by the COV method were highly correlated with the variance of \( w \) (\( \text{var}(w) \)), and were usually much higher than those estimated by LP method (Fig. 10). This might be caused by wave contamination or instrument wobble that remained in spite of the rotation used.

3.5.3 Estimate of bed shear stress

We next derived the observed bed shear stress using a combination of the estimates obtained using the LP and COV methods. The LP method was more reliable when sediment concentration fell below a threshold where sediment stratification would be expected to be small. The COV was more reliable when \( \text{var}(w) \) was below a threshold, suggesting that wave contamination and/or instrument wobble did not corrupt the estimate. Figure 11a shows the velocity shear determined by the LP (x axis) versus COV (y axis) methods. The data points did not appear to cluster around 1:1 line but showed a
wide scatter at high values of bed stress. The data points that fall far above and below the line were associated with high variance in the vertical velocity, and high sediment concentrations \((c_s)\), respectively. Different thresholds of \(c_s\) and \(\text{var}(w)\) were tried to restrict the scattered points. The final result is shown in Fig. 11b with threshold values of \(c_s\) and \(\text{var}(w)\) around 0.014 kg/m\(^3\) and 0.008 m\(^2\)/s\(^2\), respectively. The data points colored red were from bursts that had a large \(\text{var}(w)\) (i.e. > 0.008 m\(^2\)/s\(^2\)); hence LP method was used to estimate the bed shear stress for these data. Green points were associated with high \(c_s\) (i.e. >0.14 g/l); therefore the COY method worked best for these data. Blue points indicated bursts for which both \(\text{var}(w)\) and \(c_s\) were high, so that neither the LP nor COY methods were reliable. Finally, for points colored black, both methods seem reasonable and the average of the COY and LP methods was used. After application of the above criteria, the final bed shear stress is shown in Fig. 12b (blue line).

Similar criteria could not be applied to the 40-m ADV, however, because the YSI mounted on the 40-m tripod was damaged and did not provide measured sediment concentration. Nonetheless, the shear stresses estimated by the two methods for the 40-m tripod were comparable (Fig. 10), and both provide a somewhat reliable estimate of \(\tau_c\). The final bed shear stress was then determined based on the comparisons of COY and LP results with the estimates by a one-dimensional boundary layer model described below.
4. One-dimensional boundary layer model

Though field observations provided estimates of sediment flux and compelling evidence of current-induced gravity flows, their resolution was too coarse to allow for estimates of total sediment transport within the gravity layer. For that purpose, a bottom boundary layer model (Wiberg et al., 1994) was used in conjunction with gravity-flow theories as described by Wright et al. (2001) and Traykovski et al. (2007). Through this exercise, we demonstrated that current-supported flows were more important for sediment flux than were wave-boundary layer gravity flows at this site, and better characterized the gravity-flow layers.

4.1 The model

A theoretically-based model complemented field observation by extrapolating measurements of velocities and sediment concentration made at a few points to cover the bottom few meters of the flow. The model was based on boundary layer theory addressed by Smith (1977) as well as Grant and Madsen (1979), and was developed and modified primarily by Wiberg and colleagues (e.g. Wiberg and Smith, 1983; Wiberg et al., 1994; Harris and Wiberg, 1997). It used a continuous eddy-viscosity profile, thereby avoiding a discontinuity near the top of the wave boundary layer. With current velocity provided as input at one level, the model solved for the complete velocity profile by assuming that it is acted upon by the Coriolis force, bed friction, and turbulence generated by wave orbital motions and quasi-steady currents.

The required model inputs included wave orbital velocity, wave period, and along- and across-shelf current velocities at a given reference level. For sediment
transport, the model also required as input characteristics of sea-bed sediment including
the fraction of sediment, critical shear stress, and settling velocity for each of several
sediment classes. The model included essential processes from boundary layer theory,
including sediment-induced stratification (McLean, 1992), moveable bed roughness
(Wiberg and Rubin, 1989; Wiberg and Harris, 1994) and bed armoring (Wiberg et al.,
1994; Harris and Wiberg, 1997). Previous applications of the model have shown it
capable of reproducing field observations of suspended sediment concentration profiles
and bed shear stresses (Wiberg et al., 1994; Cacchione et al., 1999)

The model has also been applied to study wave supported gravity-flows on the Po
shelf (Traykovski et al. 2007). When both suspended sediment-induced stratification and
bed armoring were taken into account, the modeled sediment concentration profiles
matched ABS-measured near-bed concentration profiles offshore of the Po during times
of wave-supported gravity flows. This demonstrated that the model’s estimates can be
used to represent sediment concentration profiles for high concentration events associated
with sediment gravity flows.

4.2 Model input

Field data were used to derive inputs to the one-dimensional model, including wave
information (orbital velocity, \( u_w \), and period, \( T \)), current velocity (\( u \) and \( v \)) provided at a
single elevation (\( z_r \)), and bottom sediment characteristics for several sediment types or
sizes (grain size, settling velocity, critical shear stress, fraction of seabed, and density,
\( \rho_s \)). Wave characteristics \( u_w \) and \( T \) were determined via spectral analysis of ADV
measurements, while the reference current (\( u \) and \( v \)) were specified as the across- and
along-shelf velocities from the same ADV (~ 40 cm ab). Sediment cores were collected at both tripod locations at the beginning of the deployment to characterize the sea bed. Size distributions of the top two centimeters of the sediment cores were analyzed using the pipette method. Data from non-flood seasons indicated that the spatial distribution of surface sediment on the Waiapu shelf varied widely. For example, sand accounted for 10% of the total material at 60-m isobath, but represented about 40% of total material at 40-m isobath (Kniskern, 2007; Wadman et al., 2008). For this model, however, we assumed that sediment distributions during the event periods were more similar to riverine material and did not vary significantly between water depths of 40 and 60 m. We therefore used the sediment distribution from the 60-m site to represent the initial sediment bed for both the 40- and 60-m sites, because it was more consistent with the sediment distribution within the river (Hicks et al., 2000). It seemed more likely to represent conditions on the shelf during times of fluvial delivery of material, which was the focus of our modeling exercise.

With \( \rho_s \) of 2,650 kg/m\(^3\), settling velocities (\( w_s \)) and critical shear stresses (\( \tau_{cr} \)) of coarse particles (\( \Phi < 4.5 \)) were determined using an empirical equation (Dietrich, 1982) and non-dimensional Shields curve (Smith, 1977), respectively. We assumed that fine sediment would be partially consolidated, and would likely have been incorporated into flocs. For particles with \( \Phi > 4.5 \), \( w_s \) was assigned to be 0.6 mm/s, and \( \tau_{cr} \) was selected to be 0.08 Pa, which was consistent with the apparent \( \tau_{cr} \) from looking at the observed time series of \( c_s \). A hydraulic roughness of 0.005 cm was chosen to better match the observed bed shear stress. This value is consistent with those previously used.
for the model, such as Cacchione et al. (1999) on the Eel shelf and Traykovski et al. (2007) on the Po shelf.

The model included the effect of bed armoring. If the surface layer of the seabed is composed of sediment with mixed grain sizes, then finer sediment, associated with smaller critical shear stress and lower settling velocity, is more likely to be mobilized from the bed and kept suspended in the water column. Once fine material is removed, the remaining coarse fraction armors the bed surface, making underling finer sediment inaccessible for resuspension. As the result, less fine sediment can be suspended than the flow is capable of carrying. The surface layer with mixed grain sizes is referred as an "active" or "mixed layer", within which sediment is available for resuspension (Wiberg et al., 1994). The model accounted for bed armoring by limiting the volume of suspendable sediment to be the amount within the active layer. Previous implementations of the model tried to calculate the thickness of the active layer ($\delta_{mix}$) based on movable bed theory, and typically used $\delta_{mix}$ was in the order of millimeter (Wiberg et al., 1994; Harris and Wiberg, 1997; Cacchione et al., 1999). This paper, however, solved for the appropriate $\delta_{mix}$ needed to match the observed data during events.

4.3 Bed shear stress

The model estimated wave-induced ($\tau_w$), current-induced ($\tau_c$) and wave-current combined ($\tau_{cw}$) bed stresses. Modeled $\tau_c$ was compared in Fig. 12 to the observed shear stresses derived from measured values as described in section 4.5. They matched well for the 60-m case. For the 40-m tripod, the modeled $\tau_c$ was more consistent with
the value determined using the LP than the COV method. Therefore, \( \tau_c \) from the LP method was regarded as our most reliable observed value of shear stress for the 40-m tripod. Total shear stress, which was compared to critical shear stress to evaluate whether sediment can be mobilized and resuspended, is also shown as black lines in Fig. 12.

Overall, during all of the resuspension phases, bed stresses were sufficient to mobilize freshly delivered sediment. The total shear stresses during E1-R and E3-R were much higher than those during corresponding flood phases, and these high stresses were accompanied by strong seaward across-shelf currents (Fig. 5). The second event differed in terms of total shear stress and sediment flux. Shear stresses in E2-R were lower than (40-m case) or as high as (60-m case) those during E2-F. However the across-shelf currents in E2-F were landward and the near-bed sediment concentration was low. This implied that fluvial sediment delivered during E2-F was retained on the inner shelf until current direction changed.

4.4 Sediment concentration

Suspended sediment concentration at 0.91 mab estimated by the model for the 60-m site was compared to observations by that tripod’s YSI (Fig. 13a). They matched very well, though the model tended to underestimate \( c_s \) during event periods. Peak concentrations exceeded 2 kg/m\(^3\) in the first two events, and reached 3.2 kg/m\(^3\) in the third one. Based on Figs. 5 and 11, these high concentrations coincided with strong seaward across-shelf currents as well as high bed shear stresses. Modeled sediment concentration for the 40-m site at the same elevation (Fig. 13b) was slightly higher than the 60-m site in response to higher bed stresses.
Figure 13c displays the active layer thickness ($\delta_{\text{mix}}$) used in the model. During the second phase of each event, $\delta_{\text{mix}}$ was selected to provide agreement between the modeled and observed $c_s$, whereas during all the other times $\delta_{\text{mix}}$ was calculated by the model based on Harris and Wiberg (1997). During the resuspension phases $\delta_{\text{mix}}$ values of up to 1.6 cm were chosen, which was significantly larger than used for other implementations of this model (1~2 mm) (e.g., Wiberg et al 1994). The larger $\delta_{\text{mix}}$ values needed for resuspension phases implied that more sediment was available for resuspension during those periods, which was consistent with the inferred delivery of new fluvial mud.

The ADV observations were used to verify the settling velocity used in this model for fine-grained sediment (0.6 mm/s). Assuming the near-bed flows were steady and uniform, conservation of suspended sediment mass can be stated as the balance between upward diffusion by turbulence and downward settling,

$$< w' c_s '>= w_s < c_s >,$$  \hspace{1cm} (5)

where primes and $<$ > indicate turbulences and burst average, respectively. Burst-averaged backscatter from the 60-m site’s ADV (42.5 cmab) were related to the suspended sediment concentration modeled for the same elevation. Figure 14a shows the fit as an exponential relationship with a correlation coefficient of 0.79. Based on this, ADV backscatter within each burst was further calibrated into instantaneous sediment concentrations, from which $c_s'$ was calculated. Diffusive sediment flux ($< w' c_s '>$) was then plotted versus $< c_s >$ (Fig. 14b), where $w'$ was determined in the same way as in Eqs. (1) and (2). The slope of the regression line (0.56 mm/s) represented an observed
value of the effective settling velocity of suspended sediment, which validated our choice of 0.6 mm/s as the settling velocity of fine sediment.

4.5 Gravity flow velocity and thickness

Assuming a well-mixed turbid layer of thickness $\eta$ overlies a sloping seabed with gentle slope $\theta$, then the downhill effective gravitational force acting on the turbid layer per unit seabed area can be expressed as

$$\gamma (\rho - \rho_w) g \eta \cdot \sin \theta,$$

where $\rho$ and $\rho_w$ are turbid layer and seawater density. The first and second terms of Eq. (6) represents total gravitational force of the layer and the buoyancy due to immersed weight of sediment in seawater. The density of the turbid layer $\bar{\rho}$ can be calculated based on sediment density $\rho_s$ and mass concentration of the turbid layer $c_\tau$ as

$$\bar{\rho} = c_\tau + \rho_w (1 - c_\tau / \rho_s).$$

Under non-homogeneous conditions, $c_\tau$ can be estimated by vertically averaging the sediment concentration:

$$c_\tau = \frac{1}{\eta} \int_0^\eta c_\tau(z) dz.$$

Neglecting the large-scale pressure gradient force as well as the drag force from the overlying water, the force resisting reduced gravity is friction as described by the drag equation:

$$\rho_w C_D u_{\text{graw}}|u_{\text{max}}|.$$
Here $C_D$ is the frictional bottom drag coefficient ($= 0.003 \sim 0.005$), and $u_{grav}$ is the velocity of the gravity-driven flow. The magnitude of the velocity at the top of the layer, $u_{\text{max}}$, is the combination of the wave orbital velocity amplitude ($u_w$), along-shelf current speed ($v_c$), and the gravity-driven current, $u_{\text{max}} = \sqrt{u_w^2 + v_c^2 + u_{grav}^2}$ (Scully et al., 2002).

Combining Eqs. (7) - (9) provides a force balance within the turbid layer,

$$B \sin \theta = C_D u_{grav} |u_{\text{max}}|,$$

where $B$ is depth integrated buoyancy anomaly formulated as

$$B = \frac{g (\rho_s - \rho_w)}{\rho_s \rho_w} \int \eta c_s(z)dz.$$

Assuming $C_D = 0.003$, $g = 9.8 \text{ m/s}^2$, $\rho_s = 2,650 \text{ kg/m}^3$, $\rho_w = 1,025 \text{ kg/m}^3$, by utilizing measured $u_0$ and $v_c$ from Fig. 9 and sediment concentration profiles from the one-dimensional model, the velocity of gravity-driven flow with thickness of $\eta$ was estimated via an iterative method based on Eqs. (10) and (11). The thickness of the turbid layer, $\eta$, could be as thin as the wave boundary layer, which was estimated through $\delta_w = u_{sw} / 2 \omega$, but should be larger if current-supported gravity flows are suspected as they are here (Ma et al. 2008).

Figure 15 compares observed across-shelf velocities and estimated gravity flow velocities estimated using assumed turbid layer thicknesses that ranged from the wave boundary layer ($\sim 2 - 3 \text{ cm}$) to 2.0 m. The values estimated for a wave boundary layer gravity flow ($u_{gw}$) were much smaller than velocities estimated for thicker turbid layers. During the event periods, $u_g$ estimated for 1.5-m thick (40-m site) and 2-m thick (60-m
case) seemed to explain the observed large seaward velocities. In contrast, $u_{gw}$ estimated for wave boundary layer gravity flows was small enough to be negligible (Fig. 15a).

The gravity flows on the Waiapu shelf between 40- to 60-m water depths did not appear to be trapped within thin wave boundary layers ($< 10$ cm) but were spread up into much thicker layers ($> 1$m) by strong currents. Intense turbulence caused by currents broke up potential lutoclines at the top of the wave boundary layer and mixed sediment and presumably momentum upward, which supported the movement of these turbid layers down-slope. The greater thickness of gravity flows assumed for the 60-m site compared to the 40-m site was consistent with the fact that the deeper location experienced stronger currents during event periods than did the shallower site. At the shallower location, peaks in the current velocity did not coincide with wave peaks. Given the thickness of gravity flow layers, the vertically-averaged $c_s$ within the layers and $c_s$ at the top of those layers was estimated based on the modeled concentration profiles. For the 40- and 60-m locations the vertically-averaged $c_s$ ranged from 2.2~4.3 kg/m$^3$ and 2.2~3.5 kg/m$^3$, respectively; $c_s$ at 1.5 mab for 40-m depth and 2 mab for 60-m depth ranged from 1.5~2.5 kg/m$^3$ and 1~2 kg/m$^3$, respectively. This implied that turbid layers having sediment concentrations as low as 2~4 kg/m$^3$ appear sufficient to support current-driven gravity flows, much more dilute than those observed in wave-supported gravity flows offshore of the Eel and Po Rivers.
4.6 Cumulative sediment flux

Time series of sediment fluxes within gravity flow layers (Fig. 16) were calculated based on sediment concentrations and velocities estimated by the boundary layer model. During the study period, the along- and across-shelf cumulative fluxes estimated for the 40-m site were $1.3 \times 10^5$ and $1.1 \times 10^5$ kg/m, slightly greater than those for the 60-m site ($1.1 \times 10^5$ and $1.1 \times 10^5$ kg/m, respectively). This implies that very little deposition occurred between these two isobaths, and that most of the fine sediment presumably bypassed the 60-m isobath, driven seaward by gravity flows and northeastward by mean shelf currents.

A striking feature of Fig. 16 is that flux mainly occurred during the resuspension phases that followed actual flood pulses (i.e. E1-R, E2-R, E3-R). Based on our analysis of gravity-driven velocities (Fig. 15), current-supported gravity flow contributed substantially to these high fluxes. These gravity flows transported $2.5$, $1.5$ and $4.5 \times 10^4$ kg/m sediment offshore during the resuspension phases of events 1-3, respectively. For a single pulse (2~3 days) of high flow associated with a strong wave/current event, the sediment flux caused by gravity flows on the Waiapu shelf was on the same order of magnitude as that estimated for the Eel shelf ($\sim 5 \times 10^4$ kg/m) but higher than seen offshore of the Po River ($\sim 0.8 \times 10^4$ kg/m) (Traykovski et al., 2007). This implied that current-supported gravity flows can carry sediment loads comparable to or even higher than those transported by wave-supported gravity flows, though the depth-averaged concentrations in current-supported gravity flow layers (2.2~4.3 kg/m$^3$) are much lower than in wave boundary layers (>10 kg/m$^3$). The greater thickness of current-supported
gravity flows compensated for the lower sediment concentration, thus increasing the capacity to carry suspended sediment.

5 Event interpretations and conclusions

Sediment dispersal of flood material from the Waiapu River occurred within two distinct phases on the middle continental shelf: a flood phase characterized by a relatively calm ocean and tremendous fluvial sediment input, and a resuspension phase associated with fast currents, strong bed shear stresses and high near-bed sediment concentrations. Short time lags of only a few days separated the original sediment delivery from subsequent sediment resuspension. During resuspension phases PC-ADP velocity profiles exhibited convex-seaward shapes that were associated with strong seaward bottom velocities and high near-bed sediment concentrations. Dynamically, this seaward transport of sediment within the near-bed turbid layer appeared similar to those gravity flows observed on the Eel shelf and Po subaqueous delta except that the gravity flows observed on the Waiapu shelf were much thicker and more dilute. Unlike the Eel and Po sites, it was currents, not waves, that dominated the turbulence of the turbid layer here. Based on gravity flow theory and estimation from a one-dimensional boundary layer model, these current-supported gravity flows greatly contributed to across-shelf velocities and appeared to be significantly thicker (1-2 m) than those previously observed in wave-supported cases. Suspended sediment concentrations of 2-4 kg/m³ seemed sufficient for supporting current-supported gravity flows on the Waiapu shelf. Total depth-integrated sediment fluxes within these current-supported gravity flows were comparable to those seen offshore of the Po and Eel Rivers within wave boundary layers.
For resuspension phases of events, our one-dimensional model needed to assume much thicker active layers (1.6 cm) than have usually been used (1 - 2 mm) in order to estimate suspended sediment concentrations that matched observed values. This implied that sediment availability during resuspension phases was higher than normal due to the delivery of fresh fluvial material by the preceding flood. In a sense, application of a thick $\delta_{mix}$ provided an ad-hoc method for including advection of flood material from the inner shelf to the tripod locations. Because gravity-driven velocities peaked during these resuspension phases, this further implied that gravity-driven flows would be most important on the Waiapu shelf during resuspension phases of flood dispersal periods, when there was sufficient sediment available.

The sequence of processes that contributed to the dispersal of Waiapu delivered sediment during the tripod observation periods can be summarized as follows. Tremendous sediment was input on the Waiapau inner shelf during discharge pulses. Some of the sediment was exported by transport within the river plume and/or by local currents. Other sediment was deposited temporarily on the inner shelf because the sediment was discharged at a rate faster than they could be transported by water column suspension. This could be because currents and waves were too weak to maintain sediment in suspension, or the currents were landward and trapped the sediment on the inner shelf, or river plume and ocean currents produced convergent flow. With intensifying waves/currents a few days later, the sediment was resuspended from the seabed. Some of the suspended sediment was transported northeastward by along-shelf currents (Kniskern, 2007), whereas the remaining sediment was trapped near the bed, producing a bottom turbid layer. When the currents turned seaward, this turbid layer was
transported offshore in the mode of gravity-driven flow. In the near-shore area (shallower than 40-m isobath), the gravity flow was probably supported mainly by waves like those observed on the Eel and Po shelves (Kniskern, 2007). As the flow moved into deeper water, waves attenuated quickly, while currents maintained their energy or even intensified, for example from the 40- to 60-m water depth in Fig. 9. These strong currents diffused suspended sediment upward into a thick turbid layer, which was 1.5 - 2 m thick on the middle shelf. In this manner, the gravity flow was maintained, even as wave velocities attenuated with depth, and continued to transport sediment seaward.

Seaward of 60-m water depth, the fate of the gravity flows could not be as easily surmised due to lack of field observations. Recent studies on the Waiapu shelf have found event layers in sediment cores collected at the mid-shelf depocenter, centered at the 100-m isobath (Kniskern, 2007), as well as on the shelf break and outer-shelf (Addington et al., 2007). Based on gravity flow theory and collective field experience from wave-supported cases, we inferred that current-supported gravity flows might travel seaward continually or in pulses until they reach long-term depositional sites, where the sea-bed slope decreases and/or the strength of the currents decreased (Fig. 2). In addition, some sediment might escape from the continental shelf past the shelf break as turbidity flows that are triggered by current supported gravity flows on the continental shelf.
References


Harris, C.K., Wiberg, P.L., 1997. Approaches to quantifying long-term continental shelf sediment transport with an example from the Northern California STRESS mid-shelf site, Continental Shelf Research 17, 1389-1418.


<table>
<thead>
<tr>
<th>Instruments</th>
<th>40-m Tripod</th>
<th>60-m Tripod</th>
<th>Function</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sontek Acoustic Doppler Velocimeter (ADV)</td>
<td>18 cmab</td>
<td>26 cmab*</td>
<td>Measures 3-D turbulence at fixed elevations above bed</td>
</tr>
<tr>
<td></td>
<td>40 cmab</td>
<td>42.5 cmab</td>
<td></td>
</tr>
<tr>
<td>Sontek Pulse-Coherent Acoustic Doppler Profiler (PCADP)</td>
<td>67 cmab</td>
<td>70 cmab</td>
<td>Downward-aimed, measures velocity profiles very near the bed</td>
</tr>
<tr>
<td>RDI Acoustic Doppler Current Meter (ADCP)</td>
<td>No</td>
<td>217 cmab</td>
<td>Upward-aimed, velocity profile through water column</td>
</tr>
<tr>
<td>Sontek Acoustic Doppler Profiler (ADP)</td>
<td>230 cmab*</td>
<td>No</td>
<td>Upward-aimed, velocity profile through water column</td>
</tr>
<tr>
<td>Aquatec Acoustic Backscatter Sensor (ABS)</td>
<td>85 cmab*</td>
<td>62 cmab*</td>
<td>Near-bed suspended sediment concentration</td>
</tr>
<tr>
<td>YSI Water quality sensor</td>
<td>41 cmab*</td>
<td>91 cmab</td>
<td>Turbidity, temperature and salinity</td>
</tr>
<tr>
<td>Sediment Trap</td>
<td>87 cmab</td>
<td>46 cmab</td>
<td>Collecting near-bed suspended sediment</td>
</tr>
<tr>
<td></td>
<td>88 cmab</td>
<td></td>
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</table>

* Failed during deployment.

Table 1. Summary of tripod instrumentations.
<table>
<thead>
<tr>
<th>Event</th>
<th>Sub-event</th>
<th>Flood /Resuspension Peaks</th>
<th>Descriptions</th>
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<tbody>
<tr>
<td>E1</td>
<td>E1-F</td>
<td>05 June 04, 12:00</td>
<td>weak waves and currents</td>
</tr>
<tr>
<td></td>
<td>E1-R</td>
<td>07 June 04, 21:00</td>
<td>strong waves and current, across-shelf currents seaward</td>
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<tr>
<td>E2</td>
<td>E2-F</td>
<td>19 June 04, 07:00</td>
<td>waves and currents were medium at the beginning, then turn to weak</td>
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<tr>
<td></td>
<td>E2-R</td>
<td>23 June 04, 21:00</td>
<td>strong waves and current, across-shelf currents seaward</td>
</tr>
<tr>
<td>E3</td>
<td>E3-F</td>
<td>30 June 04, 06:00</td>
<td>Strong waves, weak currents</td>
</tr>
<tr>
<td></td>
<td>E3-R</td>
<td>01 July 04, 05:00</td>
<td>very strong waves and current, across-shelf currents seaward</td>
</tr>
</tbody>
</table>

Table 2. Characteristics of three events occurred during field observations on the Waiapu shelf.
Figure 1. Site map of the Waiapu shelf, North Island, New Zealand. Triangles show the locations of tripod deployments at 40-m and 60-m isobaths. Star and circle in the left lower panel indicate locations of wind and river discharge measurement.
Figure 2. Bathymetric profile of the continental shelf offshore of the Waiapu River. Triangles show the locations of tripod deployments at 40-m and 60-m isobaths.
Figure 3. Configuration of the tripods deployed on the Waiapu shelf.
Figure 4. Calibration data of YSI mounted on the 60-m tripod.
Figure 5. Time series of a) river (black) and sediment (gray) discharge, b) near-bed sediment concentration, c) RMS wave height (black) and current speed (gray), and d) across-shelf current velocity. Water discharge was provided by the Gisborne District Council (20 km upstream from the river mouth). Sediment discharge is based on the sediment rating curve of Hicks et al. (2004). Near-bed sediment concentration calibrated from turbidity readings of 60-m YSI (Fig. 4). RMS wave height based on linear-wave theory as \( H_{rms} = 2u_w \sinh(kh)/\omega \), where wave orbital velocity \( u_w \) and radian frequency \( \omega \) were determined from the 60-m ADV velocity recordings via power spectral analysis of following Madsen (1994) (Ma et al., 2008). Current speed and across-shelf velocity were all from 60-m ADV. Positive numbers represent seaward currents.
Figure 6. Time series of a) wind vectors and velocity vectors at the b) surface (5.79 m below sea surface), c) middle (29.79 mab), d) lower (4.21 mab), and e) bottom (48 cmab) water column. Wind data from Hicks Bay wind station (see Fig. 1). Velocity vectors of surface, middle and lower water column recorded by 60-m ADCP. Bottom velocity vector was measured by 60-m PC-ADP.
Figure 7. a) Along- and b) across-shelf velocity profiles throughout the 60-m water column. Profiles averaged for six sub-events. The upper and lower parts of the profiles were from ADCP and PC-ADP, respectively.
Figure 8. PC-ADP measured near-bed across-shelf velocity profiles during resuspension phases of the three events for both a) 40- and b) 60-m sites.
Figure 9. Time series of near-bed wave and current conditions for both a) 40-m and b) 60-m sites, including wave period ($T$), wave orbital velocity ($u_w$), along- ($v$) and across-shelf ($u$) current velocities. Positive $v$ and $u$ mean northeastward and seaward, respectively.
Figure 10. Time series of bed shear stresses estimated by COV (black) and LP (gray) methods (upper panels), and variance of vertical velocity (lower panels) estimated with COV method.
Figure 11. Scatter plots of shear velocity estimated by LP versus COV methods for the 60-m site.
Figure 12. Time series of observed (black) and modeled (dark gray) current shear stress as well as modeled total shear stress (light gray) for both a) 40- and b) 60-m sites.
Figure 13. Time series of a) observed (black) and modeled (gray) sediment concentrations at 91 cmab for 60-m water depth, b) modeled concentration for 40-m water depth at 91 cmab, and c) active layer thickness used by the model.
Figure 14. Scatter plot of a) backscatter measured by 60-m ADV versus sediment concentration estimated by model for the 60-m site, and b) $<c_s>$ versus $<c_s,w>_t>$, where $<>$ indicates burst average and primes indicate turbulent component.
Figure 15. Estimated velocities of wave-supported gravity flow (black lines in upper panels), and current-supported gravity flow (black lines in lower panels) compared with observed across-shelf velocities (gray lines in lower panels, positive numbers indicate seaward direction).
Figure 16. Cumulative along- and across-shelf fluxes estimated for both the a) 40- and b) 60-m sites.
Chapter 4: Modeling sediment transport and deposition by wave- and current-supported gravity flows on the Waiapu continental shelf

Yanxia Ma, Carl T. Friedrichs, Courtney K. Harris, L. Donelson Wright
Abstract

A two-dimensional model for wave- and current-supported sediment gravity flows was used to represent sediment transport and deposition on the Waiapu shelf, New Zealand over an annual cycle of storm events and associated Waiapu River floods. The model was run from September 2003 to August 2004, with benthic tripod data used to calibrate wind and wave forcing from the NOAA WAVEWATCH III global hindcast. The 12-month run was divided into two portions: a lower energy portion (September to May) with weak waves and currents and low river discharge, and a higher energy portion (May to August) with stronger waves and wind-driven currents and more frequent river floods. Model results indicated that during the low energy period, sediment delivered by the Waiapu River was trapped between the 20- and 80-m isobaths. During the following high-energy period, sediment was deposited obliquely across the shelf between the 60- and 120-m isobaths. The predicted deposit locations for the low- and high-energy portions, respectively, match well with short- and long-term observed accumulation patterns based on $^7\text{Be}$ and $^{210}\text{Pb}$ activity (Kniskern, 2007; Kniskern et al. 2009). Sensitivity analysis indicated that the flows were mainly wave-supported landward of the 40-m isobath, but became increasingly current-supported as wave orbitals decayed in deeper water. Wave-supported gravity currents were sensitive to local water depth and favored deposition parallel to isobaths as depth increased. In contrast, current-supported gravity currents were more sensitive to seabed slope, with decreasing slope favoring transport convergence. We conclude that the longer term (~100 yr), shelf-oblique mud deposit on the Waiapu shelf is mainly the result of current-supported gravity flows responding to local variations in the slope of the shelf.
1. Introduction

Until the 1990s, sediment gravity flows were largely considered insignificant for the dispersal of riverine sediment across continental shelves because few shelves are steep enough to provide sufficient gravity forces to maintain autosuspension, and few river effluents are turbid enough to directly produce hyperpycnal discharges. This opinion was fundamentally challenged a decade ago by new field observations and hydrodynamic theory and reexamination of existing data indicated that gravity flows can occur on more gently sloping continental shelves when waves and/or currents provide sufficient turbulence to develop near-bed hyperpycnal layers by sediment resuspension (Traykovski et al., 2000; Ogston et al., 2000; Wright et al., 2001).

Three prominent examples of shelves impacted by sediment gravity flows are found seaward of the Eel, Po and Yellow Rivers. Physical and geological studies were conducted on the Eel shelf in the late 1990s as part of the STRATAFORM program (Nittrouer, 1999). One of the fundamental conclusions of STRATAFORM was that across-shelf transport of Eel sediment was dominated by wave-supported gravity flows (Traykovski et al., 2000; Harris et al., 2005). A significant fraction (~25%) of Eel delivered sediment initially settled from the river plume along the coast just north of the river mouth (Sommerfield and Nittrouer, 1999). Large wave energy then facilitated across shelf sediment transport in the form of near-bed gravity flows, allowing 25-30% of the sediment to join the mid-shelf flood deposit, with the remainder escaping along-shelf, further offshore or buried in nearshore sands (Crockett and Nittrouer, 2004; Harris et al.,
EuroSTRATFORM collected observations on the Po subaqueous delta from the winter 2000 to the spring 2003 (Nittouer et al., 2004). Field and modeling results again showed that wave-supported gravity flows played an important role in moving sediment downslope offshore of the river mouth (Traykovski et al., 2007; Friedrichs and Scully, 2007).

The Yellow River is characterized by extremely high sediment concentrations and is one of relatively the few rivers from which hyperpycnal discharge regularly occurs. However, the direct hyperpycnal river flow extinguished itself very close to the Yellow River mouth, and most of the sediment was initially trapped within near-shore areas (Bornhold et al., 1986; Wright et al., 1990). Based on revised sediment gravity flow theory, Wright et al., (2001) reanalyzed field data collected offshore of the Yellow River in the 1980s and concluded that the turbid flows maintained by tidal currents could potentially move sediment downslope during slack tide when eddy viscosity was temporally relaxed.

Sediment gravity flows can be wave- (e.g., off the Po and Eel) or current-supported (e.g., off the Yellow River), based on the main source of turbulence. Wave-supported gravity flows have been well studied in the last decade, in part, because they were observed in both the STRATFORM and EuroSTATAFORM projects, and also because the predictability of bottom orbital velocities from wave heights simplified theoretical applications. Wave-induced flows are supported by velocity shear associated with wave orbital velocity within the thin wave boundary layer. Reduced shear at the top of the layer restricts the suspension from diffusing upward, potentially leading to very high sediment concentrations. Fluid mud with sediment concentrations as high as 80
kg/m³ were inferred within the wave boundary layer on the Eel shelf (Traykovski et al., 2000; Wright et al. 2001). The capacity of such flows to hold suspended sediment is limited by a balance between velocity shear and sediment induced stratification as indicated by the Richardson Number (Wright et al., 1999; 2001). Deposition from these flows is then determined by the gradient of near-bed orbital velocity with depth (Traykovski et al., 2000; Scully et al., 2002).

Conclusive evidence of current-supported gravity flows in a mid-shelf environment, on the other hand, has only been documented very recently. Tripod observations collected on the Waiapu shelf, New Zealand, in 2004 provided the first evidence that current-supported gravity flows could be an important mode of across-shelf sediment dispersal in relatively deep water (Ma et al., 2008). In contrast to thin (< 10 cm) and dense (concentration >10 kg/m³) wave-supported gravity flows, the current-supported gravity flows observed on the Waiapu shelf were significantly diluted (depth-averaged concentration ~ 4 kg/m³), and extended over a thicker layer of 1~2 m (Chapter 3). Compared with wave orbital velocities (based on wave buoys or wave hindcasts), near-bed sub-tidal currents are often less predictable. As a result, predicting the locus of deposition caused by current-supported gravity flow presents additional challenges (Wright and Friedrichs, 2006).

This paper extends the analysis of sediment transport and depositional processes from a vertical dimension (considered in Chapter 2) to a lateral domain. It utilizes a two-dimensional numerical model for sediment gravity-driven flows (vertically integrated over the benthic boundary layer) to investigate sediment deposition on the Waiapu shelf. The objectives of the paper are to determine whether gravity flows may play a key role in
forming the muddy deposits on the Waiapu shelf and to investigate the effects of currents versus waves on gravity flows and resulting shelf deposition.

2. Site description

The Waiapu River drains a small mountainous basin (~1700 km²), characterized by steep terrain, heavy rainfall (~2.4 m/yr), and unconsolidated soft Tertiary mudstone and siltstone. As a result, its sediment yield is tremendous (~21,000 t km²/yr) (Milliman and Syvistki, 1992), far higher than most rivers. Its water discharge is very episodic over both inter- and intra-annual timescales, and almost all of the discharge is associated with floods brought on by cyclonic storms. Hyperpycnal plumes may occur in the Waiapu system during flood periods, i.e. total suspended solids in the river can exceed 40 kg/m³ (Mulder and Syvitski, 1995). At the river mouth, fine-grained materials (>90%) dominate the sediment load (Hicks et al., 2000), favoring high flocculation. Due to the small drainage basin of the Waiapu River, the river-ocean system is strongly coupled in that the weather systems that produces heavy rains also creates strong waves and wind-driven currents in the coastal ocean. The floods usually last a few days, and the associated sediment is introduced onto the inner shelf rapidly, overwhelming the sediment dispersal capacity of the coastal ocean.

Over the Waiapu shelf, our tripod observations indicated that long-period swell from south Pacific was significant, with mean wave heights on the order of 1.3 m and mean wave period of about 11.4 s. During storms, significant wave heights exceeded 3-4 m with typical wave period of 11s (Ma et al., 2008). In addition, currents on the middle shelf (40-60 m) were strong, exceeding 1 m/s at the surface and reaching 0.5 m/s near the
seabed. Large scale circulations influences the outer shelf, including the southwestward East Cape Current (ECC) and northeastward Wairarapa Coastal Current (Chiswell, 2000) inshore of the ECC.

The Waiapu continental shelf (Fig. 1) is located on the leading edge of the Australian plate under which the Pacific plate is subducting. This subduction combined with tremendous sediment load contribute to the high subsidence rate (0.4 cm/y) of the Waiapu basin (Lewis et al., 2004). The shelf is composed of a Holocene sediment wedge that is over 110 m thick centered on the 110-m isobath directly offshore of the river mouth. This massive Holocene deposition has created a convex-upward shelf profile as is commonly seen over the landward portion of progradational clinoform shelves. Preserved deposits facies on the Waiapu inner shelf exhibits significant spatial variation, with river-plume and event deposits dominating north and south of the river mouth, respectively (Wadman et al., 2008). On the mid-outer shelf the average accumulation rate is about 1.5 cm/yr, with the highest accumulation rate at the Holocene depocenter of about 3.5 cm/yr, which is significantly higher than the typical global range of 0.1 - 1 cm/yr (Kniskern, 2007). The head of the Ruatoria Indentation defines the Waiapu shelf break (Lewis et al., 1998; 2004). Event layers have been found in the sediment cores collected on this shelf break, suggesting the Ruatoria Indentation is a potential conduit through which sediment escapes to deeper water (Kniskern et al., 2009).

Kniskern (2007) and Kniskern et al. (2009) measured the $^7$Be and $^{210}$Pb excess activities of sediment cores collected throughout the Waiapu shelf in August 2003 and May 2004. The half-life of $^7$Be and $^{210}$Pb are 53 days and 22 years, respectively, hence they are used to infer deposition on time scales of several months and decades,
respectively. Figure 2 displays the $^7$Be activities measured in surface sediment (0–2 cm) in May 2004 (upper), just before the tripod deployment of Ma et al. (2008), and the long-term accumulation rate based on excess $^{210}$Pb profiles (lower). The short-term accumulation rate just before May 2004 was highest on the inner shelf (<50 m water depth) but became close to zero on the middle-to-outer shelf, whereas the long-term accumulation trend was reversed in that $^{210}$Pb was mainly found seaward of 50 m. The results indicated that rapid deposition occurred adjacent to the Waiapu River mouth in the months leading up to May 2004, but temporary shallow water deposits were probably moved seaward across the shelf over longer time scales. Based on X-radiographs of the sediment cores, Kniskern et al. (2009) characterized that long-term sedimentation on the Waiapu shelf was by event deposits related to hyperpycnal flows.

3. The two-dimensional model

3.1 Theoretical development

Within a gravity flow layer, the momentum balance can be represented using an upslope frictional drag and down-slope pressure gradient force induced by suspended sediment (Wright et al., 2001),

$$B \sin \theta = C_D |u_{\text{max}}| u_g.$$  \hspace{1cm} (1)

In Eq. (1) $\theta$ is the seabed slope, $C_D$ is the frictional drag coefficient (~0.003), and $u_g$ is the velocity of gravity-driven flow. The quantity $u_{\text{max}}$ is the magnitude of the instantaneous velocity at the top of the gravity layer, which can be expressed as

$$u_{\text{max}} = \sqrt{u_w^2 + v_c^2 + u_g^2},$$  \hspace{1cm} (2)
where $u_w$ is the root-mean-square (RMS) amplitude of wave orbital velocity (equal to $2^{1/2}$ times the standard deviation of the instantaneous orbital velocity), $u_g$ is the down-slope directed, gravity-driven current speed, and $v_c$ is the non-gravity driven, mainly along-shelf directed ambient current. The depth-integrated buoyancy anomaly $(B)$ induced by suspended sediment with mass concentration $c_s$ can be estimated as

$$B = \rho_s^{-1} g s \int_{z=0}^{\delta} c_s dz,$$

in which $g$ is the acceleration of gravity, $s = (\rho_s - \rho_w) / \rho_w$ is the submerged weight of the sediment relative to seawater, $\rho_s$ and $\rho_w$ are the density of the sediment and water, and $\delta$ is the turbid layer thickness.

A key scaling parameter that controls the capacity of the gravity flow to hold suspended sediment is the Richardson number,

$$Ri = B / u_{\text{max}}^2.$$

It expresses the importance of the stabilizing effects caused by stratification relative to the destabilizing effects induced by velocity shear. Bed stresses induced by intense waves and currents can generate strong bottom turbulence and can also resuspend bottom sediment causing strong stratification. The suppression of turbulence by the stratification leads to settling of the suspended sediment; but settling out of sediment then reduces stratification, which enhances the turbulence once more. This negative feedback keeps the $Ri$ in the neighborhood of a critical value (of order $Ri \approx 1/4$), as has been demonstrated by field observations of bottom turbid layers on the Amazon shelves.
Combining Eqs (1), (3), and (4) yields expressions for \( u_g \) and for the depth-integrated sediment concentration for periods when the turbid layer has its maximum sediment capacity:

\[
\begin{align*}
    u_g &= \beta \cdot u_{\text{max}} = \beta \cdot \sqrt{\frac{u_w^2 + v_c^2}{1 - \beta^2}}, \\
    \int c_s dz &= \frac{R_i \rho_s}{g_s} \left[ \frac{u_w^2 + v_c^2}{1 - \beta^2} \right],
\end{align*}
\]

where \( \beta = R_i \sin \theta / C_D \). Thus the maximum down-slope rate of sediment transported by gravity flows can be estimated by as the product of Eq. (5) and Eq. (6):

\[
Q_g = u_g \int c_s dz = \frac{\rho_s R_i \omega^2}{g_s C_D} \left( \frac{u_w^2 + v_c^2}{1 - \beta^2} \right)^{3/2}.
\]

The rate of down-slope sediment transport increases with bed slope, \( \theta \), as well as with the magnitude of wave orbital velocity, \( u_w \), and near-bed ambient current speed, \( v_c \). The rate of deposition is determined by the gradient in sediment transport, \( D = -dQ_g / dx \), and hence is controlled by the gradients in bed slope and wave/current energy. It is reasonable to assume \( dv_c / dx << du_w / dx \), since mean currents decay with water depth more slowly than waves; thus

\[
D = Q_g \left\{ -\frac{3u_w}{u_w^2 + v_c^2} \frac{\partial u_w}{\partial x} + \frac{1 + 2\beta^2}{(1 - \beta^2)\alpha} \frac{\partial \alpha}{\partial x} \right\}.
\]

Equation (8) shows that the deposition is controlled by the wave orbital velocity, along-shelf current, and shelf slope. Since wave orbital velocity attenuates with water
depth, the first term in the brackets always favors deposition. The second term in the brackets either increases deposition if the seafloor profile is concave upward ($\frac{\partial \alpha}{\partial x} < 0$), or decreases deposition if the profile is convex upward ($\frac{\partial \alpha}{\partial x} > 0$).

3.2 Model implementation

The two-dimensional implementation of the above equations was developed by Scully et al. (2003) and has been applied successfully on the Eel (Scully et al., 2003) and Po (Friedrichs and Scully, 2007) shelves to represent patterns of muddy flood deposit formation in response to wave-supported gravity flows. Here we extend the two-dimensional model to include the effect of significant ambient currents. Assuming there are sufficient river-delivered sediment available to allow $Ri=Ricr$, the model solves the aforementioned equations to predict gravity flow velocity and sediment deposition. Near-bed orbital velocity, $u_\omega$, is calculated from surface wave height and period using linear wave theory, and near-bed along-shelf velocity, $v_c$, is input directly. For each time step, $t$, and each grid cell, $i$, the model calculates the theoretical suspended sediment capacity ($C_{t,i}$) of the gravity flow based on Eq. (6) and then compares $C_{t,i}$ with the amount of existing suspended sediment from the previous time step ($S_{t-1,i}$). New deposition occurs if $C_{t,i}$ is smaller than $S_{t-1,i}$. But if $C_{t,i}$ exceeds $S_{t,i}$ and $u_{max}$ is greater than a critical value ($u_{erode}$), erosion occurs ($E_{t,i}$). The amount of eroded sediment is determined as $E_{t,i} = C_{t,i} - S_{t,i}$, if there is enough previously deposited sediment available to reach $C_{t,i}$. Otherwise, any previously deposited sediment is eroded. The suspended sediment for time step $t$, $S_{t,i}$, is estimated based on the above arguments, and the depth-integrated buoyancy anomaly $B$ is calculated through Eq. (3). The model then calculates the gravity-driven flow velocity
based on Eq. (1) via an iterative method and determines the downslope sediment flux for each grid cell. The sediment flux is then used to adjust $S_{li}$ for the new calculation at the next time step. For all of the model runs presented in this paper, $R_{icr} = 1/4$, $C_d = 0.003$ and $u_{erosion} = 0.35$ m/s, the same values used for the Po shelf by Friedrichs and Scully (2007).

For the present application, the active model domain covered a region extending along shore from 11 km northeast of the Waiapu River mouth to 19 km southwest of the river mouth and from the coastline to offshore near the 170-m isobath (Fig. 1). The x and y axes of the domain corresponded to the across- and along-shelf directions, respectively. The domain was composed of 145×127 rectangular element grids, and each grid cell represented an area of about 207 m in along-shelf direction and 193 m in the across-shelf direction. Bathymetry, which is shown in Fig. 1, was provided by NIWA (New Zealand National Institute of Water & Atmospheric Research). The shorelines were digitized from a world coastlines database (http://www.ngdc.noaa.gov/mgg/shorelines/shorelines.html), and the water depth along the shoreline was assigned to be zero. Bathymetry was smoothed in both the along- and across-shelf directions first via a two-dimensional triangle filter and then by a cubic spline on every seventh grid point. Finally, the shelf slopes in the along- and across-shelf directions were calculated based on the smoothed bathymetry.
3.3 Model inputs

The two-dimensional model was applied to the Waiapu shelf for the 12-month period extending from 1 September 2003 to 31 August 2004. This period was chosen to cover both storm and calm weather seasons, to encompass the observation of sediment accumulation conducted by Kniskern (2007) and Kniskern et al. (2009), and to overlap the tripod observations reported by Ma et al. (2008). Wave period and height, shelf current speed, and riverine sediment discharge were the required model inputs. Since the simulation period was longer than the available tripod observations, hindcast data for wave height, wave period, and wind velocity were derived from the NOAA WAVEWATCH-III global ocean wave model (NWW3). NWW3 is a publicly available, third generation global wave model with a resolution of 1° latitude and 1.25° longitude (Tolman, 2002). NWW3 wave heights then were calibrated based on wave heights inferred from the tripod by Ma et al. (2008), and NWW3 winds were related to the observed along-shelf currents. The most reliable tripod data were from May to July 2004; the observed waves and currents were all obtained within this time interval from an ADV mounted about 40 cm above the seabed on the tripod at the 60-m isobath.

3.3.1 Waves input

Although the NWW3 grid spacing (1.25 by 1 deg) is relatively coarse, the nearest NWW3 node is advantageously located, only 30 km offshore of the Waiapu River mouth (Fig. 1). The NWW3 hindcast reported significant wave height, whereas the two-dimensional model used RMS wave height ($H_{rms}$). To correct for this inconsistency and also help compensate for other potential systematic differences, the NWW3 wave
heights were multiplied by a constant such that the 90th percentile of all NWW3 wave heights reported from May to July 2004 equaled the 90th percentile of all $H_{rms}$ values ($H_{rms90}$) observed at the 60-m tripod site (Fig. 3, top). $H_{rms90}$ was suggested by Friedrichs and Wright (2004) to be an appropriate statistic for predicting wave-supported gravity flows offshore of a variety of rivers around the globe. A reasonably strong correspondence was found between observed waves at the tripod and the NWW3 hindcast with a correlation coefficient of 0.76 (Fig. 3, top). Efforts were made to relate NWW3 hindcast wave periods to those observed at the tripod, but no conclusive relationship was found (the correlation was only 0.06). Other researchers (Trembanis et al. 2007) have similarly found that comparisons between NWW3 hindcasts and observations were much better for wave height than for wave period. Following Friedrichs and Wright (2004), we instead estimated wave periods as a function of wave height for the NWW3 hindcast by assuming

$$T = C \cdot \sqrt{H_{rms}},$$

(9)
in which $C$ is a constant. Assuming the most reliable observations of wave period from our tripod data were from when waves were large (observed $T = 12$ s when $H_{rms} = 2.5$ m, see Chapters 2&3), we had $C = 7.6$.

3.3.2 Current input

During energetic times, tidally-averaged currents on the Waiapu shelf were often as strong as or stronger than wave orbital velocities (Ma et al., 2008; Chapter 2) and, in contrast to the Eel and Po cases, cannot be neglected in the calculation of $u_{max}$ (see Eq. 2).
If one assumes that the along-shelf current ($v_c$) dominates the across-shelf current, and that $v_c$ is predominantly driven by the along-shelf wind stress, it follows that

$$v_c = [v_{wind} \times (u_{wind}^2 + v_{wind}^2)^{1/2}]^{1/2},$$

where $v_{wind}$ and $u_{wind}$ are the along- and across-shelf wind velocity. The correlation between the observed along-shelf current and the right-hand-side of Eq. (10) using NWW3 hindcast winds for May to July 2004 was a reasonably strong 0.70. Given this encouraging relationship, the NWW3 winds were converted to along-shelf currents based on a linear regression of equation Eq. (10) and then multiplied by a constant to ensure that the hindcast currents had the same energy (i.e., sum of the squared current velocity) as the observed $v_c$ (Fig. 3, bottom). Finally, the time-series for along-shelf current velocity was assigned uniformly onto each model grid point without considering energy decay with water depth or possible topographic effects.

A year-long time-series of waves and currents was generated by applying the methods discussed above to the available NWW3 hindcast. The estimated RMS wave height and current velocity from September 2003 to August 2004 are shown in Fig. 4, based on which the period can be divided into two portions: 1) a low-energy portion (LEP) from 1 September 2003 to the mid-May 2004, and 2) a high-energy portion (HEP) from May to August 2004. During the LEP, mean RMS wave height, period and current speed were 0.81 m, 6.7 s and 0.08 m/s, respectively, and only one significant sediment-yield event occurred (Fig. 4c). During the HEP, wave height, period and current were 1.3 m, 8.6 s and 0.12 m/s on average, respectively. Several high discharge events occurred during the HEP, providing abundant sediment to feed gravity-driven flows (Fig. 4c)
3.3.3 Sediment input

The model assumed that no pre-existing erodible sediment was available on the seabed, and the only source of sediment input was fine-grained material delivered by the river. The sediment discharge time-series for the Waiapu River was provided by NIWA and was calculated by applying a rating curve from Hicks et al. (2004) to the water discharge measured by the Gisborne District Council. To produce realistic flood deposition, the sediment input of the model was set to be only 23% of the amount predicted by the rating curve. Sensible reasons for reducing the sediment input include the following: (i) The rating curve includes sands, while only fine sediment is transported by the modeled gravity flows. (ii) Riverine sediment are dispersed over the Waiapu shelf not only by gravity-driven flow, but also by other mechanisms, such as suspension, which likely carries significant amount of sediment out of the model domain (Kniskern et al., 2008). (iii) The river gauging station is 20 km upstream from the river mouth. Because of river channel aggradation, the amount of sediment delivered to the Waiapu coastal ocean is likely less than inferred 20 km upstream. (iv) According to Hicks et al. (2004) the rating curve used for Fig. 4c contains inherent uncertainty on the order of 40% and may overestimate the sediment concentration at high discharge because of the weighting given to concentrations associated with a few very high concentration, extremely high-yield events. In their application of the two-dimensional gravity flow model to the Po shelf, Friedrichs and Scully (2007) also reduced riverine sediment input to a constant fraction of that predicted by the Po rating curve for analogous reasons.

The sediment was introduced into the model along the coast as a line source. Two distinct line sources were considered, one with spatially uniform input all along the
model's landward edge, and one which decreased sediment discharge both north and south of the river mouth. In the latter case, a high energy period deposit quite similar in shape to the long-term deposit reported by Kniskern et al. (2009) was produced if the line source decreased linearly with distance to the south of the river mouth and exponentially with distance to the north of the river mouth (Fig. 5). The methodology here was again consistent with the approach of Friedrichs and Scully (2007), who employed both spatially uniform and spatially varying line sources off the Po. Based on the bulk density of cores reported by Kniskern et al. (2009), a seabed porosity of 0.6 was used here to estimate the ultimate deposit thickness.

4. Results and Discussion

4.1 Predicted versus observed gravity flow velocities

With the wave orbital velocity, current speed, and bed slopes known at each grid point, the two-dimensional model solved Eq. 1 via an iterative method to predict the across-shelf gravity flow velocity \( u_g \). Modeled \( u_g \) at the grid point nearest to the 60-m tripod site in response to NWW3 forcing was compared with observed \( u_g \) in Fig. 6. The observed \( u_g \) was defined as the near-bed across-shelf velocity from the second bin of 60-m PC-ADP \( (\sim 43 \text{ cmab}) \) minus the free stream across-shelf velocity from the middle bin of the 60-m ADCP \( (\sim 26 \text{ mab}) \) for the period of May to July 2004, when the most reliable tripod data were available.

The magnitudes and frequency of the observed and model estimated \( u_g \) match well (Fig. 6), which is key for successfully reproducing the general nature of the observed sediment deposits during the low- and high- energy hindcasts. The more imprecise time
lags between the observed $u_g$ and the model calculation forced by NWW3 reflected some discrepancy between the actual ambient waves and currents and those inferred from NWW3 timeseries.

4.2 Deposition during low-energy portion

The low-energy portion of the NWW3 hindcast covers 257 days, about 5 times the half life of $^7$Be. This period was divided into 10 segments, and at the end of each segment the modeled deposit was multiplied by a decay factor so that

$$D_{i+1} = D_i \cdot e^{-\Delta t / t_{1/2}},$$

(9)

where $D_i$ is the deposit thickness predicted at the end of segment $i$, $D_{i+1}$ is the deposit saved for the next segment $i+1$, $\Delta t$ is 25.7 days, and $t_{1/2}$ is the half life of $^7$Be (53.3 days). In this way, the final deposit is weighted toward more recent deposition in order to roughly represent the fraction of sediment potentially tagged with actively decaying $^7$Be.

Figure 7 displays the result of the calculated deposition for the LEP. The sediment was almost entirely deposited at water depths shallower than 80 m, with the maximum deposit thickness found along the 40-m isobath. The focusing of the deposit along the 40-m isobath was in response to a local maximum in bed slope at that depth. As discussed by Friedrichs and Scully (2007) in the context of the Po Shelf, greatest deposition by sediment gravity flows under low wave conditions tended to occur where steep slopes first stopped increasing with distance offshore. Pockets of high deposition along the 40-m isobath were somewhat patchy, most likely in response to local irregularities in bed slope. This is consistent with the results from Kniskern et al. (2009) who examined the penetration depths of Kasten cores and found that muddy sediment deposits landward of
the 80-m isobath were patchy. Maximum deposition tended to occur to the southeast of the river mouth, reflecting in part the weighting of riverine sediment input at that location. The deposition was likely further concentrated by a bathymetric low, which extends 10 km southward from the river mouth near the coast, narrows offshore and becomes indistinct at around 80-m water depth (Wadman et al., 2008). Gravity flows are expected to move sediment into that region from both northeast and southwest directions following the local bathymetry.

The markers (stars, crosses, triangles) superimposed on the estimated LEP deposition (Fig. 7) were sites of observed high $^7$Be activities from within the model domain based on the May 2004 survey of Kniskern (2007). The highest $^7$Be activity (shown as stars) generally matched modeled areas of high deposition, while medium (crosses) and relative lower (triangles) activities surrounded the isolated depocenters. The results showed that in the short-term (~ 5 months), sediment delivered by the Waiapu River during periods of low energy was mainly trapped between the 20- and 80-m isobaths.

4.3 Deposition during high-energy portion

The result of the 2-D model calculation for the high energy portion (HEP) is displayed in Fig. 8. The model formed two depocenters within the model domain. The main one was elongated and occurred between 60- and 120-m water depths southeast of the river mouth, trending obliquely to the bathymetry. Its locus matched remarkably with that of the observed highest long-term accumulation rate as shown in Fig. 2 (dark gray). The mean accumulation rate calculated by the model for this depocenter was about 4.9
cm/yr, similar to the observed long-term accumulation rate of 3.5 cm/yr. The location of this obliquely trending depocenters followed the along-shelf position of a local maximum in bed slope, consistent with the accentuation of deposition where slope first starts decreasing as described by Friedrichs and Scully (2007) for the Po. There was also a tendency for deposition to begin just offshore of the point where \( u_{\text{max}} \) can no longer transport all available sediment (~ 70 m), consistent with the main mechanism for deposition controlling gravity flow deposition offshore of the Eel (Scully et al. 2002, 2003).

Another distal depocenter predicted by the HEP model was located at the northeast corner of the model domain offshore of the 150-m isobath, with a mean deposition rate of about 4.5 cm/yr. This depocenters also coincided with a region of decreasing bed slope in the model. Since the depocenter was close to the model boundary, the model was unable to estimate how far offshore this deposit would extend. But immediately seaward of this depocenter is the Ruatoria Indentation described by Lewis et al. (1998; 2004). This sediment presumably would escape from the Waiapu shelf through the indentation into deeper water.

During the HEP simulation, of the 23% (7.5 MT) of sediment included into the gravity flow model, about 94.7% was deposited within the model domain; 0.6% was trapped in the water column; 3.7% escaped from the model domain into the deep sea; and only 1% went northward out of the model domain. Although this model could not evaluate the fate of sediment that was not associated with gravity flows, model results clearly suggested that it was during high-energy periods that the sediment delivered by
the Waiapu River was transported by gravity-driven flows into the region of long-term accumulation.

4.4. The role of currents vs. waves

The presence of strong currents highly influences sediment accumulation on the Waiapu shelf. Based on Eq. (8), currents can compensate for the decay of wave orbital velocity with depth, and hence can cause sediment to accumulate farther offshore and in deeper water. Figure 9 displays sediment deposition estimated by the model with the same settings used in Fig. 8, but neglecting currents. It represents the depositional pattern caused by wave-supported gravity flows alone on the Waiapu shelf. Sediment were collectively (>99%) deposited parallel to the bathymetry, centered around the 60-m isobath, with a mean depocenter thickness of about 6.7 cm. Through comparison with Fig. 8, it is clear that, on the Waiapu shelf, waves tended to favor accumulation of sediment on the middle shelf, whereas currents spread out sediment more effectively over the whole shelf. These results also implied that wave-supported gravity flows evolved into current-supported gravity flows as they moved into deeper water, which is consistent with the conceptual model presented by Ma et al (2007).

Figure 10 shows two deposition profiles along with associated seabed slopes. The positions of these profiles are indicated by the grey lines in Fig. 8. The across-shelf profile passes through the tripod sites, the northern end of the main depocenter and the southern end of the distal depocenter. With strong currents, significant deposition began around 70-, 110- and 150-m water depths. The latter two locations were just offshore of where bed slopes have begun to decrease, consistent with the arguments of Friedrichs and
Scully (2007) (Fig. 9b). This effect can be understood through Eq. (8), where a negative $\partial \alpha / \partial x$ (decreasing slope) in the second term on the right hand side produces positive deposition. Deposition increases further with the effect of this 2nd term on the right-hand-side of Eq. (8) as the magnitude of $\alpha$ decreases offshore in the denominator.

Without currents, however, much more sediment was trapped between the water depths of 50 and 70 m, where the decay of wave orbital velocities with depth caused wave-supported gravity currents to first lose their capacity to carry all of the available riverine sediment (highly analogous to the Eel shelf case) (Fig. 10a). Unlike wave-supported gravity-driven flows that converge quickly where increased depth causes wave orbital velocities to dramatically decrease (Friedrichs and Scully, 2007), the current-supported gravity flows can move the sediment farther seaward, because the currents compensate for the wave energy decay. This distinction is seen in the first term in the right-hand-side of Eq. (8). Under wave-dominated conditions, the net effect of $u_w$ is in the denominator, while under current dominated conditions, the net effect of $u_w$ is in the numerator. As $u_w$ decreases with depth, this first term becomes more important under wave domination, but becomes less important under current domination. Therefore, deposition by wave-supported gravity flows is relatively more sensitive to total depth, while deposition by current-supported gravity flows is relatively more sensitive to seabed slope. This is why the wave-only deposit more closely paralleled the bathymetry (following a pattern in $h$), while the current-influenced deposit diagonally crossed the bathymetry (following a pattern in $\alpha$).

The along-shelf profile displayed in Fig. 10 roughly paralleled the 70-m isobath, cutting through the center of the main depocenter. Significant sediment deposition was
estimated to occur south of the river mouth for both wave-dominated and current-influenced cases. Although the peak of sediment input was at the river mouth (Fig. 5), the highest deposition on the shelf occurred around 8 km south of the river mouth (Fig. 10c) for the current-supported case and 1-5 km south of the river mouth (red line) for the wave-supported case. Unlike the influence of the across-shelf profile, the along-shelf slope can encourage gravity flows go either up-coast or down-coast. Along with the effects of the along-shelf slope, the deposition pattern along this profile reflected the cumulative influence of gravity flows landward of this cross-section line. For the wave-supported case, except for small-scale fluctuations, the general trend of the deposition visibly followed the along-shelf sediment distribution (Fig. 5), further suggesting, in the large scale, that wave-supported gravity flows tend not to respond as strongly to small changes in bed slope (<0.002). Smoothed bathymetry showed a slight bathymetric low around 7km south of the river mouth, indicated by the negative slope in Fig. 10d, and this bathymetric low acted as a conduit for gravity flows only when strong currents were present.

5. Model sensitivity

Recent studies have shown that patterns of gravity-flow deposition are influenced greatly by spatial variations in sediment supply and bed slope (Scully et al., 2002; 2003; Friedrichs and Scully, 2007). To examine the effects of these key factors on sediment deposition on the Waiapu shelf, sensitivity tests were conducted. Model runs for only the high energy portion of the year are presented below, because the model results for the LEP were found to be less sensitive to those factors.
5.1 Sediment supply

To produce a realistic deposition pattern, our model input specified less sediment put north than south of the river mouth. To test the sensitivity of model calculation to the along-shelf sediment distribution, the model was also run using a uniform sediment distribution that spread sediment input equally along-shore with the same total sediment load as the base model run in Fig. 8. The resultant deposition pattern was patchier than the base case (Fig. 11). The deposits directly offshore and 7 km south of the river mouth was smaller, because much less sediment was input around the river mouth. Figure 12 compares the along- and across-shelf deposition profiles (indicated in Fig. 8) of the base model run (solid lines) with those of the uniform sediment distribution (dashed lines). The uniform distribution resulted in greater deposition 10 ~ 15 km south of the river mouth, corresponding to a narrow region between the latitude of -37.88° ~ -37.95° (Fig. 11). It leads to similar offshore flux out of the model domain (3.9%), but more northward flux (6.3%) compared with those values for the base case. However, in the across-shelf direction, the influence of sediment distribution was mainly reflected in the deposition thickness. A key result was that even with uniform sediment input, the deposit created by the gravity flow model was oblique to the bathymetry, consistent with the observation of Kniskern et al. (2009). This supported the conclusion that shelf bathymetry played a key role in controlling the location and shape of deposition.
5.2 Along-shelf slope

The influences of along-shelf bathymetry on the Waiapu shelf sediment deposition has been addressed partially in previous sections. For example, the bathymetric low south of the river mouth acted to steer the sediment delivered by the Waiapu River southward. In addition, the along-shelf deposition tended to be inversely correlated with the along-shelf distribution of slope as shown in Fig. 10. To further examine the importance of the along-shelf bed slope to overall deposition, the model was run again, but with along-shelf bed slope set to be zero. Not unexpectedly, by neglecting the along-shelf components of gravity flows, the sediment tended to be deposited more directly offshore of the river mouth (Fig. 13), with the along-shelf deposition (Fig. 12) strongly reflecting the initial along-shelf distribution of sediment (Fig. 5). The across-shelf deposition pattern depicted in Fig. 12(a) showed less deposition in the distal depocenters, and the along-shelf distribution in Fig. 12(b) more directly reflected the distribution of riverine sediment input, with less sediment shifted southwards. This was because the along-shelf component of slope helped steer the sediment both offshore to the north and along-shore to the south.

6. Summary and Conclusions

Substantial effort has been devoted to understanding sediment transport and redistribution off high-yield rivers. The importance of wave-supported gravity flows to the dispersal of fluvial sediment has been recognized since they were directly observed on the Eel and Po shelves during the STRATAFORM and Euro-STRATAFORM projects, respectively. The geometries of the flood deposits on the Eel (Scully et al., 2003) and Po
(Friedrichs and Scully, 2007) shelves were previously shown to be highly consistent with emplacement by wave-supported gravity flows through application of a two-dimensional numerical model.

In this paper, the two-dimensional gravity-flow model developed by Scully et al. (2003) was applied to the Waiapu shelf by accounting for not only waves but also shelf currents. The purpose was to evaluate the likely significance that sediment gravity flows, especially the current-supported cases, played in the formation of the Holocene muddy deposits on the Waiapu shelf. The model was based on theory for gravity flows supported by ambient waves and currents as described by Wright et al. (2001) and assumed critically stratified flow when sufficient sediment was available. Basic model parameters, including the Richardson number (=0.25), the bottom drag coefficient (=0.003), and the critical velocity for seabed erosion (=0.35 m/s) were consistent with those used for the Eel (Scully et al., 2003) and Po (Friedrichs and Scully, 2007) shelves. To account for sediment suspension by wind-driven currents, winds hindcast by the NOAA WAVEWATCH III were correlated to shelf current speed. Tripod data collected at 60-m depth from May to August 2004 were used to calibrate the WAVEWATCH III forcing, including wave height, and along-shelf current.

The model successfully reproduced the depositional features of the riverine sediment on the Waiapu shelf by representing both the magnitude and location of the observed muddy deposit. During the fair weather season from September 2003 to May 2004, sediment tended to be trapped between the water depths of 30 and 60 m, consistent with the distribution of ⁷Be in May 2004 cores as reported by Kniskern (2007). During storm season in the months of May through August 2004, this recently deposited
sediment along with newly delivered riverine sediment were transported offshore, because during this time stronger waves and currents had an increased capacity to carry sediment. The resulting shelf-oblique deposit was consistent with the long-term (100 yr) accumulation pattern as indicated by $^{210}$Pb (Kniskern et al., 2009). Model simulations revealed that marine forcing (waves and currents), sediment supply and bathymetry were the three key factors that influenced patterns of shelf deposition.

Currents and waves were shown to play distinct roles in sediment transport and distribution on the Waiapu shelf. Currents tended to dominate the sediment gravity flows in water depths that exceeded about 40 m, where waves become secondary due to rapid decay of orbital velocities. When shelf currents were weak, deposits tended to occur in shallower water depths parallel to the isobaths during both high and low energy periods. This pattern occurred because transport convergence due to waves was driven by the strong dependence of near-bed orbital velocity on absolute depth. In contrast, current-supported gravity flows were less sensitive to absolute depth and could compensate for the decay of wave orbital velocity, moving sediment further offshore. Current-supported flows were more sensitive to the slope of the seabed, and deposition by current-supported gravity currents occurred where bed slope decreased with distance offshore.

Sensitivity tests further revealed the importance of the along-shelf distribution of sediment input and the role of the along-shelf component of bed slope. To produce a more realistic deposition pattern, the base case assumed a maximum in sediment input directly offshore of the river mouth. Using a uniform along-shelf sediment distribution resulted in a patchier deposit, but the trend remained oblique to the shelf bathymetry,
supporting the conclusion that subtle bathymetry strongly influences the pattern of current-induced accumulation. With the along-shelf bed slope was neglected, sediment tended to be deposited more directly offshore of the source with less steering of current-supported gravity currents to the north or south.

Across-shelf sediment transport on the Waiapu shelf resulted from the combination of both wave- and current-supported gravity flows. However, we concluded that the longer term (~100 yr), shelf-oblique mud deposit found between the 60- and 120-m isobaths mainly reflected current-supported gravity flows and their response to local variations in the slope of the seabed. Under high-energy conditions, wave-supported flows over the inner shelf (< ~40 m water depth) transitioned to current-dominated support as the flows moved progressively into deeper water. This conclusion confirmed the conceptual model derived in Chapter 2, and was also consistent with the analyses described in Chapters 2 and 3 which provided direct evidence of current-supported sediment gravity flows at the tripod sites. Overall, the results presented in this dissertation can help us better understand gravity-driven sediment transport, especially those flows supported by strong currents, and thereby provide new direction to the study in sediment dispersal of high-yield rivers, such as those reviewed in Chapter 1.
References


Figure 1. Study area of the Waiapu continental shelf (with actual bathymetry), northern island of the New Zealand. Two tripods at 40-m and 60-m isobaths are shown as triangles. The domain of 2-D gravity-driven flow model is shown as the bold square, which covers an area of ~29 km × 25 km for the coast areas to 160-m isobath. The location of the NOAA WAVEWATCH-III global model node used for wind and wave forcing is shown as a square.
Figure 2. Sediment accumulation patterns based on $^7$Be (upper, from Kniskern, 2007) and $^{210}$Pb (lower, from Kniskern et al., 2008) on the Waiapu continental shelf. $^7$Be indicates that sedimentation over 6-months previous to May 2004 was mainly confined to depths shallower than 80 m, but $^{210}$Pb shows a high long-term (100 year) sedimentation rate between the 60-m and 120-m isobaths in a pattern oriented oblique to the isobaths.
Figure 3. a) Root mean square wave height ($H_{rms}$) and b) current from ADV observation of the tripod at 60-m isobath (black), calibrated NOAA WAVEWATCH III ocean wave model hindcast (gray).
Figure 4. Calibrated a) root mean square wave height ($H_{rms}$) and b) current velocity ($V_c$) from NWW3; c) sediment discharge ($Q_s$) from the Waiapu River based on water discharge measured at Gisborne District Council and rating curve from Hicks et al. (2004).
Figure 5. Along-shelf distribution of fluvial sediment input off the Waiapu River mouth for the base-case model run. Note that there is an exponential decrease north of the mouth and a linear decrease south of the river mouth.
Figure 6. Time series of gravity-driven flow velocity from May to July 2004 of observed (solid line) relative to two-dimensional model forced by NWW3 output (dashed line).
Figure 7. Modeled sediment deposition thickness during the low-energy portion (LEP) from September 2003 to May 2004. The superimposed symbols are significant $^7$Be activities from Kniskern (2007), which are also shown in Fig. 2.
Figure 8. Modeled sediment deposition thickness during the high-energy portion (HEP) from May to August 2004. Two grey lines are along- and across-shelf profiles used in Figs. 10 and 12.
Figure 9. Modeled wave-supported (without currents) sediment deposition thickness during the high-energy portion (HEP) from May to August 2004.
Figure 10. Sediment deposition and corresponding slope across and along the shelf under the conditions with and without currents. Locations of profiles are shown in Fig. 8.
Figure 11. Modeled sediment deposition thickness during the high-energy portion (HEP) from May to August 2004 using a uniform along-shelf distribution of fluvial sediment input.
Figure 12. Across- and along-shelf sediment deposition during the high-energy portion (HEP) from May to August 2004 using the 'realistic' along-shelf distribution of fluvial sediment input (see Fig. 5), uniform along-shelf input, and no along-shelf component of shelf slope.
Figure 13. Modeled sediment deposition thickness during the high-energy portion (HEP) from May to August 2004 with no along-shelf component of shelf slope.
Chapter 5: Conclusions and future work

1. Conclusions

This dissertation comprehensively analyzed field data recorded by two instrumented tripods that were deployed in 2004 on the Waiapu continental shelf at 40- and 60-m water depths directly offshore of the river mouth. The instruments included Acoustic Doppler Current Profiler (ADCP), Acoustic Doppler Velocimeter (ADV), Pulse Coherent-Acoustic Doppler Profiler (PC-ADP) and YSI. A one-dimensional boundary layer model and two-dimensional gravity flow model used in this study complemented the field measurements and simulated sediment deposition on the Waiapu shelf, respectively. The main findings of this study included:

1) During the field observation period, the Waiapu River floods were characterized by two distinct phases: a flood phase with initial fluvial sediment input, and a resuspension phase a few days later. During the flood phases, neither tripod recorded high sediment concentrations; time-averaged near-bed along- and across-shelf velocities were fairly weak. During post-flood resuspension phases, however, sediment was mobilized and both tripods observed strong seaward near-bottom turbid flows.

2) These turbid flows coincided with intense seaward near-bed currents and high bed shear stresses. Time-averaged near-bed along-shelf profiles followed the classic ‘law of the wall’ with velocity decreasing to zero towards the seabed, whereas all of the across-shelf velocity profiles exhibited strong seaward-convex shapes, suggesting gravity-driven turbid flow as observed on the Eel and Po continental shelves.

3) With increasing velocity shear, the Richardson number ($R_i$), the crucial parameter that indicates the stability of the turbid flows, converged to its critical value of
one-quarter ($Ri_{cr} - 1/4$) as seen for other gravity-driven flows (Friedrichs and Wright, 2004; Fredrichs and Scully, 2007; Scully et al., 2002; 2003; Trowbridge and Kineke, 1994; Wright et al., 1999; 2002). This supported the assumption of gravity flow theory based on setting $Ri = Ri_{cr}$ when there are sufficient river-delivered sediment available, as was done in Chapters 2 and 4.

4) Turbid flows observed on the Waiapu shelf at 40- and 60-m water depths were dynamically similar to wave-supported gravity flows observed on Eel and Po River shelves, except that currents were as important for sediment resuspension as waves. These current-supported gravity flows were significantly thicker and more dilute than previously reported wave-supported gravity flows, with thicknesses on the order of 1~2 m and depth-averaged concentrations of 2 to 4 kg/m$^3$. The flows carried significant across-shelf flux and continued beyond 60-m water depth.

5) The depositional pattern estimated for the Waiapu shelf over an annual cycle was very different during low- (May to August) and high-energy (September to May) periods. Predicted depocenters for the low- and high-energy periods matched well with short- and long-term observed accumulation patterns reported by Kniskern (Kniskern, 2007; Kniskern et al., 2009) based on $^7$Be and $^{210}$Pb geochronology, respectively.

6) Across-shelf sediment transport on the Waiapu shelf was the combined result of both wave- and current-supported gravity flows. However, the longer term (~100 yr) mud deposit was mainly built by current-supported gravity flows, and its location represented a response to local variations in seafloor slope. Under high-energy conditions, wave-supported flows over the inner shelf (< ~40 m water depth) transitioned to current-dominated flows as they moved progressively to deeper water.
2. Future work

This dissertation represented a comprehensive analysis of field data, and application of numerical modeling tools to examine transport mechanisms and depositional products on the Waiapu River continental shelf. Efforts in the future could be aimed at improving these analyses in several ways:

1) The one-dimensional boundary layer model used in Chapter 3 ignored the drag force between the turbid layer and overlying water column, and, most importantly, neglected the downhill gravitation term in its governing momentum equation. Therefore, it failed to provide reliable turbulence structures of the gravity-driven flows. It is important in the future to account those effects into the model to better characterize bottom turbid flows.

2) Friedrichs and Wright (2004) assented that wave-supported gravity flows were responsible for the formation of the equilibrium profile formed offshore of many high-load river mouths. The current-supported gravity flows examined here, presumably control the development of the deeper parts of the Waiapu submarine deposits and should be included in future morphodynamic analysis.

3) The Waiapu shelf is not the only river-influenced continental shelf subject to strong currents. Depending on local circumstances, those currents may be tide induced, wind-driven, baroclinic, or related to the propagation of continental shelf waves. Future efforts should extend the main findings of this study to continental shelves offshore of other high-load rivers. This could be accomplished by literature reviews to identify other areas where current supported gravity flows may be important. Field programs could then
be developed to obtain more observations of this transport mechanism. In order to classify these continental shelves in terms of the sediment dispersal, future efforts also should be devoted to identify the key factors in shelf regimes controlling the occurring of gravity-driven flows.

4) Unanswered questions about the gravity-driven flows include: what are the internal structures of the wave and/or current-supported gravity-driven flow close to the seabed? What dynamic conditions determine the onset and extinction of gravity-driven flows on the continental shelves? Future analyses should be aimed at address these questions.
Reference


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