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Excess ^{210}Pb inventories and fluxes along the continental slope and basins of the Gulf of Papua

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[1] Sediment samples were collected from continental slopes and marginal basins in the Gulf of Papua and analyzed for excess ^{210}Pb to elucidate transport processes of fine-grained particles to this region. Estimated excess ^{210}Pb fluxes of $1.0\text{--}12.8\text{ dpm cm}^{-2}\text{ a}^{-1}$ were derived from measured seabed inventories. Highest sediment accumulation rates ($0.28\text{--}0.35\text{ cm a}^{-1}$) were measured along the northeastern shelf edge, and they decrease in seaward directions and along isobaths to the southwest. The excess ^{210}Pb flux could result from either focused deposition of high- ^{210}Pb activity sediments from the continental shelf and upper slope or scavenging of ^{210}Pb brought landward from deep-sea waters. This sediment flux is concentrated in the northeastern Gulf of Papua, where the shelf is narrow and calcium carbonate contents are lowest. Analysis of sedimentary fabric and ^{210}Pb distributions in cores suggests sediment delivery to the slope occurs on a 100-year timescale as both diffuse hemipelagic deposition as well as turbidity flows. The flux of sediment in turbidity flows is not well constrained but may be producing additional deep-sea accumulation in the Moresby Trough, as well as export from the study area.

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1. Introduction

[2] The MARGINS Source-to-Sink (S2S) initiative aims to develop a quantitative understanding of sediment transport and accumulation across and along continental margins [NSF MARGINS Science Plans, 2004]. The uplands of southern Papua New Guinea (PNG) and the adjacent Gulf of Papua (GOP) (Figure 1) were chosen because of large production of siliciclastic and carbonate sediments, clearly defined sediment sources and sinks, variable sediment fluxes over a range of timescales, and a relatively pristine environment [NSF MARGINS Science Plans, 2004].

[3] A series of rivers deliver some $200\text{ to }365 \times 10^6\text{ t a}^{-1}$ of siliciclastic material (e.g., quartz, feldspars, clays) to the inner shelf of the northern GOP annually [Harris et al., 1993; Milliman, 1995; Milliman et al., 1999]. This massive sediment discharge is largely due to voluminous rainfall

($4\text{--}10\text{ m a}^{-1}$ in many places) on young and very steep (erodible) mountains where landslides frequently occur [Pickup, 1984; Harris et al., 1993]. Of the total mass discharged, some $100 \times 10^6\text{ t a}^{-1}$ accumulate on the inner to middle shelf in an extensive prograding delta system composed of clinoforms [Harris et al., 1993; Walsh et al., 2004]. Another $10\text{--}15 \times 10^6\text{ t a}^{-1}$ may collect in mangrove covered coastal regions [Wolanski et al., 1998; Walsh and Nittrouer, 2004]. Assuming that these estimates are correct, huge amounts of siliciclastic sediment ($>100 \times 10^6\text{ t a}^{-1}$) escape the GOP shelf, at least on the 100-year timescale. However, Walsh and Nittrouer [2003] indicate that very little fluvially derived sediment ($<15 \times 10^6\text{ t a}^{-1}$) accumulates beyond the shelf break.

[4] Sedimentary facies across the northern GOP shelf can be generally divided into a silt/sand-rich region of clinoform topsets on the inner shelf, a fine-grained region of clinoform foresets on the middle shelf, and a region of relict lithogenic and carbonate sands on the outer shelf [Harris et al., 1993; Brunskill et al., 1995; Walsh and Nittrouer, 2003]. This distribution suggests that very little sediment presently crosses the shelf directly south of the major river inputs. However, some sediment from fluvial sources either bypasses the clinoform or is resuspended from the clinoform to be transported clockwise around the GOP shelf [Harris et al., 1993; Wolanski et al., 1995; Walsh et al., 2004; Ogston et al., 2008]. A portion of this suspended-sediment flux crosses the outer shelf in the northeastern GOP, where the shelf narrows dramatically (Figure 1). Walsh and Nittrouer [2003] suggest that nepheloid layers,

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Figure 1. Map of the Gulf of Papua (GOP) region showing the continental slope, major basins, and reefs. The hashed line on the inner shelf is the 60 m isobath showing the extent of prograding clinoform. Black dots show locations of multicores examined in this study, whereas gray squares and white triangles show locations of cores examined by *Walsh and Nittrouer* [2003] and *Brunskill et al.* [2003], respectively. Bathymetric contour interval is 200 m; the 100 m contour is also shown.

created by northeastward along-shelf advection, transport the sediment. In any case, thick prograding sediment packages extend to the shelf edge in this area [*Wolanski and Alongi*, 1995, *Slingerland et al.*, 2008a], likely supplying siliciclastic sediment to adjacent deep-water environments.

[5] Slope and basin settings in the GOP (Figure 1) can be divided into three large physiographic regions: Pandora Trough, Ashmore Trough and, Moresby Trough [*Winterer*, 1970; *Francis et al.*, 2008]. Pandora Trough is a SW-NE trending basin confined by Portlock Reef, Boot Reef, and Ashmore Reef to the west, Eastern Plateau to the southeast, and the shelf edge to the north and east. The southwestern end is a ~1700 m deep semienclosed basin without a clear modern outlet, whereas the northeastern end is connected to Moresby Trough at ~2000 m water depth. Ashmore Trough is a SSW-NNE trending feature rimmed by active carbonate reef systems on two sides and a water depth of less than 1000 m. The third deep region, Moresby Trough, is a NW-SE trending basin confined by the narrow shelf of the Papuan Peninsula to the northeast, and Eastern Plateau and Eastern

Fields Reef to the southwest. It is connected to the Coral Sea Basin via Moresby Canyon in the southeast.

[6] The three deepwater structural basins collectively constitute a major depocenter for sediment escaping the GOP shelf (Figure 1). Sedimentation in Ashmore Trough is presently dominated by neritic carbonate input from surrounding reefs [*Francis et al.*, 2008]. Pandora and Moresby troughs, however, receive enough terrigenous material so that siliciclastic contents of surface samples can exceed 80% [*Febo et al.*, 2006]. This material presumably escapes from the shelf to the slope, principally in the northeast GOP where the sediment packages cross the shelf edge.

[7] Lead 210 (^{210}Pb) is a naturally occurring radionuclide in marine sediments, and has been widely used to study sediment dispersal and accumulation on century timescales [e.g., *Goldberg and Koide*, 1962; *Nittrouer et al.*, 1979]. Particles, organic or inorganic, scavenge ^{210}Pb from the water column and deposit ^{210}Pb in excess of that produced within sediments by the decay of its parent isotope, ^{226}Ra (half-life of 1602 years). As sediments are buried, ^{210}Pb decays with a half-life of 22 years [*Appleby and Oldfield*,

1992]. Bioturbation can significantly affect ^{210}Pb distributions in sediment [Benninger *et al.*, 1979; Thomson *et al.*, 1988]. However, measurement of ^{210}Pb activities at selected intervals can yield an activity-versus-depth relationship that can be used to estimate an apparent sediment accumulation rate [e.g., Koide *et al.*, 1972; Nittrouer *et al.*, 1984].

[8] Two studies have examined ^{210}Pb in sediment from the outermost shelf, slope and basin regions in the GOP. Walsh and Nittrouer [2003] studied 10 box cores (Figure 1) in the northeast GOP, and determined apparent accumulation rates of $\leq 0.2 \text{ cm a}^{-1}$ ($0.08 \text{ g cm}^{-2} \text{ a}^{-1}$). Brunskill *et al.* [2003] examined Kasten cores from the base of the slope where mass accumulation rates were between 0.05 and $0.07 \text{ g cm}^{-2} \text{ a}^{-1}$. Core locations used for the above cited studies were selected without the aid of detailed geoacoustic seabed surveys.

[9] Recent studies in the GOP have highlighted the importance of designing seabed sampling based on multi-beam and subbottom surveys so that depositional processes can be associated with known seabed geomorphology [Slingerland *et al.*, 2008a]. Building on previous studies, we undertook a more extensive coring program in conjunction with detailed seabed mapping [Daniell, 2008; Francis *et al.*, 2008]. These studies allow us to identify and map major pathways (gullies, canyons and channels) for sediment transport from shelf and bank top to deep water (by gravity flows and/or sediment plumes), and constrain ^{210}Pb inventories, ^{210}Pb depositional fluxes, and rates of sediment accumulation associated with the depositional environments. These observations are then used to examine possible sediment-transport pathways and processes in the GOP.

2. Materials and Methods

2.1. Field Sampling

[10] Thirty multicores were collected from shelf edges and slope basins of the northern GOP during the PANASH cruise of the R/V *Melville* during March–April 2004 (Figure 1). Core sites were selected based on seabed morphology and sedimentary facies determined from geoacoustic seabed mapping on the same cruise. Positioning was accomplished by Differential Global Positioning System and the ship's dynamic positioning system. The use of a multicorer allowed recovery of undisrupted sediment samples extending from the surface of the seabed. The presence of clearly defined, oxidized core tops and in situ arborescent foraminifera and ophiuroids in some cores suggests that our multicores indeed recovered relatively undisturbed sediment samples.

[11] Immediately after recovery, one sediment-filled core tube from each multicorer deployment was extruded at 1 cm intervals for the top 6 cm, and at 2 cm intervals from 6 cm to the base. These samples were sealed in plastic bags and refrigerated until analysis in an onshore laboratory. One additional sediment-filled tube from each multicore deployment was prepared for X-radiography. These were sliced into 2 cm thick axial slabs, and imaged onboard using a Thales Flashscan 35 digital X-ray detector panel, illuminated with a Medison Acoma PX15HF X-ray generator. Images were archived as 14 bit gray scale images with $127 \mu\text{m}$ pixel resolution.

2.2. Laboratory Measurements and Theory

[12] Samples for radionuclide measurements were dried for 24 hours at 60°C , ground with a porcelain mortar and pestle, and sealed with glue in $50 \times 9 \text{ mm}$ petri dishes. Weighed masses were then counted for 24 hours on Canberra low-background planar gamma detectors; correction for self-absorption of ^{210}Pb was done using the method of Cutshall *et al.* [1983]. Total ^{210}Pb was determined by measurement of the 46.5 keV ^{210}Pb gamma peak. Supported ^{210}Pb from the decay of ^{226}Ra within the seabed was determined by measurement of the granddaughters of ^{226}Ra , ^{214}Pb (at 295 and 352 keV) and ^{214}Bi (at 609 keV). Excess (unsupported) ^{210}Pb was determined by subtracting total ^{210}Pb activity from supported ^{210}Pb activity for each interval. All values of excess ^{210}Pb were decay corrected to the date of collection. Activities are reported in decays per minute (dpm) (with $1 \text{ dpm} = 1/60 \text{ Becquerels}$). Activity errors are derived from detector control software, and represent goodness of fit of a Gaussian curve to the observed spectral peak near 46.5 keV. Minimum detectable activity for excess ^{210}Pb is 1 dpm g^{-1} , which is approximately three times the counting error for supported activity.

[13] Apparent sediment accumulation rates (SARs) can be derived from ^{210}Pb observations using a one-dimensional, two-layer, steady state model in which mixing of sediments occurs only in the upper layer [Goldberg and Koide, 1962; Nittrouer *et al.*, 1979]. Such modeling assumes constant depositional rates for sediment and unsupported ^{210}Pb . It also assumes that the thickness of the mixed layer remains constant with time. The general equation for this steady state model is

$$dA/dt = D_b(\partial^2 A/\partial z^2) - S(\partial A/\partial z) - \lambda A \quad (1)$$

where A is activity of radionuclide (dpm g^{-1}), D_b is particle mixing coefficient ($\text{cm}^2 \text{ a}^{-1}$), λ is decay constant for radionuclide (a^{-1}), S is apparent sediment accumulation rate (SAR) in centimeter per year (cm a^{-1}), and z is depth in sediment column (cm).

[14] Once sediments are deposited, they may undergo biological mixing by benthic organisms. If biological mixing can be ignored, then $D_b = 0$ in equation (1). The solution of equation (1) is then given by [Krishnaswami *et al.*, 1980]

$$A_z = A_0 e^{(-\lambda z/S)} \quad (2)$$

where A_0 is excess ^{210}Pb activity extrapolated to the sediment surface, and A_z is activity at depth z . Apparent accumulation rates (i.e., values of SAR that assume negligible effects of biological mixing) are then calculated using least squares regressions of the ^{210}Pb profiles and application of equation (2). In addition to SAR, we calculated sediment mass accumulation rate (MAR) according to $\text{MAR} = (1 - \Phi)\rho_s \text{ SAR}$, where Φ is the average porosity for the core, and ρ_s is the density of sediment grains assumed to be 2.65 g cm^{-3} .

[15] Under steady state conditions of sediment accumulation and bioturbation, the distribution of excess ^{210}Pb activity in a core commonly delineates two distinct zones [e.g., Nittrouer *et al.*, 1979; Kuehl *et al.*, 1986]: a surface layer of uniform and high ^{210}Pb activity, where bioturbation

Figure 2a. Cores from the northern Pandora Trough. Excess ^{210}Pb activity (dpm g $^{-1}$; decay per minute per gram) is plotted on the x axis, and depth is plotted on the y axis. Solid lines represent best fit through the data points. Core locations are shown in Figure 1, and water depths are given in Table 1.

thoroughly mixes sediments; and a lower zone of exponentially decreasing excess ^{210}Pb activity, where the activity gradient is influenced by sediment burial and ^{210}Pb decay. Below this depth, all detected ^{210}Pb is supported by decay of the parent radioisotope ^{226}Ra . Sediment accumulation rates herein are estimated from the log linear slope for the lower interval of decreasing excess ^{210}Pb activity. Some cores appear to have a distinct surface mixed layer (e.g., MV-18, MV-21), but most do not. Nevertheless, only basal

portions of ^{210}Pb profiles were used to estimate accumulation rates, to minimize the influence of bioturbation.

[16] These sediment accumulation rates are upper limits because our calculations ignore the potential influences of bioturbation in this region. The minimum SAR we report is 0.05 cm a $^{-1}$, which is near the practical lower limit for ^{210}Pb geochronology and may be strongly influenced by slow, shallow bioturbation. This rate represents effectively negligible sediment accumulation over 100-year timescales.

Figure 2b. Same as Figure 2a except for cores from the Ashmore Trough and southwestern Pandora basin.

However, the same rate extended over millennial timescales represents a significant sediment flux.

[17] Other radioisotopes, including ^{137}Cs and ^{234}Th , have been used to constrain ^{210}Pb -derived SARs. However, ^{137}Cs was not detectable above our detection limit of 0.05 dpm g $^{-1}$ for this radioisotope. The cruise duration and shipping times for samples precluded measurement of excess ^{234}Th , due to its short half-life (24 d).

[18] The effects of bioturbation undoubtedly introduce significant uncertainty into some of our SAR and MAR estimates. However, the use of ^{210}Pb inventories to constrain depositional fluxes is less influenced by the effects of bioturbation. The seabed inventory of excess ^{210}Pb is calculated as the product of depth-integrated radionuclide activity and sediment dry bulk density. This seabed inventory is supplied by ^{210}Pb flux from the atmosphere and ^{210}Pb production in the water column. Sediment inventories of excess ^{210}Pb (dpm cm $^{-2}$) at the seabed can be calculated by

$$I = \sum \rho_s \Delta z (1 - \phi_i) A_i \quad (3)$$

where I is inventory, ρ_s is mineral density, Δz is thickness (cm) of the sampled interval i (1 or 2 cm), ϕ is porosity

(measured by water loss at 60°C), and A is the excess ^{210}Pb activity (dpm g $^{-1}$) of the sampled interval. Sediment ^{210}Pb inventories reported here are minimum values because not all cores penetrated to depths below the zone of excess ^{210}Pb . However, additional contributions from deeper, unsampled intervals should be minimal, because excess ^{210}Pb activities at core bases were generally more than 1 order of magnitude lower than at the surface. Errors for radionuclide concentrations are propagated accordingly to obtain error for inventory values.

[19] From this measured inventory of ^{210}Pb , we can estimate the annual flux of excess ^{210}Pb required to support the observed inventory at steady state. We refer to this as the core-calculated flux (F), determined by:

$$F = \lambda I \quad (4)$$

where F is the annual flux (dpm cm $^{-2}$ a $^{-1}$) and λ is the radioactive decay constant for ^{210}Pb (0.031 a $^{-1}$).

3. Results

[20] Core profiles of excess ^{210}Pb activity for the study area are shown in Figure 2, and measured excess ^{210}Pb

Figure 2c. Same as Figure 2a except for cores from the Moresby Trough channel.

seabed inventories are given in Table 1. X-radiographs for most multicores display a pervasively bioturbated sedimentary fabric. A typical example is core MV-14, located in the northern Ashmore Trough at a water depth of 760 m. Mottling without primary stratification characterizes sediment throughout the entire core (Figure 3). A prominent exception to this pattern is core MV-24, taken from a well-developed channel in Moresby Trough at a water depth of 2102 m. This core displays stratification at three depths (Figure 3), consistent with turbidity-current deposition [Bentley *et al.*, 2006a].

[21] Across the entire study area, ^{210}Pb sediment accumulation rates are generally highest near the shelf edge in the northern Pandora Trough (Figure 2), and decrease with increasing water depth (Figure 4), toward the southwest (i.e., toward Ashmore Trough) (Figure 2). In the Pandora Trough, SAR and MAR drop from 0.35 and 0.20 g cm $^{-2}$ a $^{-1}$ for core MV-42 at the shelf edge (water depth of 91 m), to 0.21 and 0.11 g cm $^{-2}$ a $^{-1}$ for core MV-53 on the middle slope (water depth of 1111 m) (Figure 2). These values compare favorably with those determined by Walsh and Nittrouer [2003] (≤ 2 mm a $^{-1}$, 0.1 g cm $^{-2}$ a $^{-1}$; Table 1). A similar pattern is evident for upper slope cores MV-18 and MV-19, and middle slope core MV-38 (Figure 2 and Table 1). All three cores are located along one channel system, with MAR decreasing from ~ 0.22 g cm $^{-2}$ a $^{-1}$ at 230–270 m water depth to 0.06 g cm $^{-2}$ a $^{-1}$ at 961 m water depth.

[22] Significant variability exists across depositional facies at comparable depths, however. Pandora Trough core MV-56 (water depth of 450 m; Figure 2) was collected on the upper slope within a channel (based on multibeam data) (Figure 2). The seabed at this core location consists of coarse carbonate sands (>250 μm), and the SAR for this core is less than 0.1 cm a $^{-1}$ (Table 1). Core MV-47, collected from soft mud deposits at a comparable water depth (399 m) farther to the northeast, yields a similar SAR of 0.28 cm a $^{-1}$. In the southwestern Pandora Trough at water depths exceeding 1600 m, accumulation rates measured for two cores, MV-60 and MV-70, are 0.07 cm a $^{-1}$ (0.04 g cm $^{-2}$ a $^{-1}$) and 0.12 cm a $^{-1}$ (0.08 g cm $^{-2}$ a $^{-1}$), respectively (Figure 2).

[23] In the Ashmore Trough and southwestern Pandora Trough, upper limits of accumulation rates measured are between 0.05 and 0.23 cm a $^{-1}$. ^{210}Pb profiles show simple distributions where excess ^{210}Pb activities decrease with depth. Core MV-59 (water depth of 577 m; Figure 1) was taken on the upper slope within a channel (based on multibeam data). The seabed at this core location consists of coarse carbonate sands (>250 μm) and the penetration depth was only ~ 10 cm.

[24] In the Moresby Trough, cores taken from water depths greater than 2000 m in a prominent channel and the mouth of Moresby Canyon display similar excess ^{210}Pb profiles, and yield upper limits for accumulation rates of 0.07–0.14 cm a $^{-1}$ (0.04–0.07 g cm $^{-2}$ a $^{-1}$; Figure 2). Primary stratification evident in core MV-24 (Figure 3) suggests that either bioturbation has been slower at this locale than elsewhere, or that sediment accumulation here has been more rapid (thus allowing preservation of primary fabric).

[25] Excess ^{210}Pb inventories in the northeastern Pandora Trough range between 29 and 267 dpm cm $^{-2}$ with highest inventories generally on the northeastern shelf, decreasing to the southwest (Table 1 and Figure 5). One exception is core MV-18, collected from the upper slope near the mouth of a submarine canyon at 267 m water depth, which yielded the highest inventory at 413 dpm cm $^{-2}$. Inventories in the Ashmore Trough and southwestern Pandora Trough range between 31 and 83 dpm cm $^{-2}$, with the highest values generally occurring closest to the shelf edge, and lower values in deeper water, farther from the shelf edge. Excess ^{210}Pb inventories in the Moresby Trough channel are relatively uniform, covering a range of 63–87 dpm cm $^{-2}$.

4. Discussion

4.1. Sediment Mass Accumulation Rates

4.1.1. Northern Pandora Trough

[26] Patterns of sediment accumulation in the Pandora Trough seabed are consistent with this region's intermediate location in the GOP dispersal system, between shelf depocenters, and deep-sea fans farther south in the Coral Sea Basin (the ultimate sinks in this dispersal system; Figure 1) [Winterer, 1970]. Accumulation rates are highest on the

Table 1. Excess ^{210}Pb Surface Activities, Supported Activities, Measured Inventories, Estimated Fluxes, and Upper Limits for Sediment Accumulation Rates

Core	Water Depth (m)	Supported ^{210}Pb Activity ^a (dpm g ⁻¹)	Measured Excess ^{210}Pb inventory (dpm cm ⁻²)	Theoretical Flux ^b (dpm cm ⁻² a ⁻¹)	R (Equation (4)) ^c	SAR ^d (cm a ⁻¹)	MAR ^e (g cm ⁻² a ⁻¹)
MV-10	584	0.64 ± 0.17	43	0.5	2.5	0.05	0.05
MV-12	372	0.52 ± 0.15	50	0.4	4.0	0.09	0.10
MV-14	760	1.1 ± 0.53	72	0.7	3.4	0.23	0.20
MV-16	686	1.3 ± 0.21	68	0.6	3.7	0.05	0.05
MV-18	267	0.97 ± 0.12	413	0.4	34.9	0.26	0.22
MV-19	229	0.86 ± 0.19	164	0.4	14.2	0.18	0.21
MV-20	661	1.2 ± 0.28	78	0.6	4.4	0.12	0.11
MV-21	1001	1.7 ± 0.22	83	0.8	3.1	0.05	0.04
MV-24	2102	2.2 ± 0.19	63	1.9	1.0	0.14	0.07
MV-26	2232	3.4 ± 0.44	76	2.0	1.2	0.07	0.04
MV-28	2426	3.7 ± 0.33	85	2.2	1.2	0.10	0.05
MV-30	2023	3.5 ± 0.33	87	1.9	1.5	0.10	0.05
MV-32	1620	4.2 ± 0.27	77	1.5	1.6	0.05	0.03
MV-38	961	3.1 ± 0.34	78	0.8	3.0	0.10	0.06
MV-39	690	1.4 ± 0.25	136	0.6	6.8	0.13	0.08
MV-42	91	1.4 ± 0.35	98	0.3	9.7	0.35	0.20
MV-43	62	0.53 ± 0.16	117	0.3	11.7	0.34	0.28
MV-44	99	1.4 ± 0.29	136	0.3	13.4	0.28	0.17
MV-47	399	0.89 ± 0.27	267	0.4	21.2	0.28	0.14
MV-50	795	2.2 ± 0.37	245	0.7	11.3	0.13	0.06
MV-53	1111	2.5 ± 0.38	140	0.9	4.8	0.21	0.11
MV-56	450	1.5 ± 0.20	29	0.4	2.1	0.08	0.08
MV-59	577	1.6 ± 0.19	74	0.5	4.4	NA	NA
MV-60	1671	3.5 ± 0.33	78	1.5	1.6	0.07	0.04
MV-65	1012	2.1 ± 0.26	31	0.8	1.1	NA	NA
MV-67	1147	2.2 ± 0.22	62	1.1	1.8	NA	NA
MV-69	1638	3.7 ± 0.28	83	1.5	1.7	0.05	0.03
MV-70	1758	4.6 ± 0.29	62	1.6	1.2	0.12	0.08
MV-72	1320	1.2 ± 0.28	45	1.2	1.2	0.07	0.06
MV-76	1551	2.3 ± 0.26	39	1.3	0.9	0.07	0.06
TC1						0.14 ^f	0.07 ^f
TC2						0.17 ^f	0.1 ^f
TC3						0.17 ^f	0.1 ^f
TC4						0.16 ^f	0.08 ^f
TC6						0.1 ^f	0.05 ^f
TC7						0.09 ^f	0.06 ^f
TC8						0.15 ^f	0.08 ^f
TC9						NA ^f	NA ^f
TC10						0.16 ^f	0.08 ^f
237	923						0.03 ^g
241	1117						0.05 ^g

^aSupported $^{210}\text{Pb} = ^{210}\text{Pb}_{\text{total}} - ^{210}\text{Pb}_{\text{excess}}$.^bThe theoretical flux is the sum of atmospheric fallout (0.3 dpm cm⁻² a⁻¹) and ^{226}Ra water column production from GEOSECS western Pacific data [Chung and Craig, 1980].^cRatio of core-derived flux (dpm cm⁻² a⁻¹) estimated from measured $^{210}\text{Pb}_{\text{excess}}$ inventory (dpm cm⁻²) to theoretical flux (dpm cm⁻² a⁻¹) given as R .^dLead 210-derived apparent sediment accumulation rates calculated by least squares fit of the analytical solution to the equation, assuming accumulation is dominant over bioturbation ($D_b = 0$). These accumulation rates, therefore, represent a maximum estimate.^eMAR (g cm⁻² a⁻¹) = $(1 - \phi) \rho_s \text{ SAR}$ (cm a⁻¹) where ϕ is the average porosity for the entire length of each core, and ρ_s is the density of sediment grains (assumed to be 2.65 g cm⁻³).^fSAR and MAR from Walsh and Nittrouer [2003].^gMAR from Brunskill et al. [2003].

outer shelf and upper slope region of the northeastern Pandora Trough (Figure 2), and decrease beyond the upper slope toward deeper water (Figure 4). Cores taken from the shelf break (MV-42, MV-43, and MV-44) exhibit measured SAR values at least twice as large as SAR on the middle slope (MV-38, MV-39, MV-50, and MV-53). This decrease in accumulation rate from the shelf break to the middle slope could be in part due to a decrease in particle flux with increasing distance from the river source. However, SAR and MAR appear to decrease along isobaths to the south-

west, closer to the Fly River, suggesting that the northeast shelf edge is a location of sediment focusing.

[27] Cores MV-60 (SAR = 0.07 cm a⁻¹) and MV-32 (SAR = 0.05 cm a⁻¹) located on the lower open slope suggest that accumulation of terrigenous sediment is negligible on a 100 year timescale.

[28] All cores collected from the Pandora Trough display wholly biogenic sedimentary fabric and lack any physical stratification, suggesting that sediment accumulation occurs slowly and steadily, allowing bioturbation to readily over-

Figure 3. X-radiographs from multicores show sedimentary structure of seabed from GOP (scale in centimeters). (left) Core (MV-14; water depth of 760 m) from Ashmore Trough showing bioturbation and absence of any physical structures. This is typical of biogenic fabric observed in X-radiographs of most cores. (right) Core (MV-24; water depth of 2102 m) taken from the Moresby Trough showing possible turbidite layers partially destroyed by bioturbation.

print primary depositional fabric [e.g., Bentley *et al.*, 2006b].

4.1.2. Ashmore Trough and Southwestern Pandora Trough

[29] Apparent MARs are generally low for cores in Ashmore Trough and the southwestern Pandora Trough ($0.03\text{--}0.10\text{ g cm}^{-2}\text{ a}^{-1}$). A MAR value of $0.20\text{ g cm}^{-2}\text{ a}^{-1}$ for core MV-14 is likely an upper limit due to bioturbation. The geomorphology of this region is dominated by reefs and carbonate platforms with intervening basins. Although channels are present in several places, seismic and piston core data show uniformly thin pelagic and hemipelagic drapes that are primarily carbonate in composition [Francis *et al.*, 2008]. The low SARs and MARs for this area (Table 1 and Figures 1 and 2) suggest terrigenous sediment accumulation is negligible on a 100 year timescale. X-radiographs from Ashmore Trough (e.g., MV-14) display wholly biogenic sedimentary fabric (Figure 3), consistent

with low accumulation rates and thorough bioturbation. The shelf reaches a maximum width of 150 km near the Ashmore Trough (Figure 1), and outer shelf sediments include calcareous algal bioherms, suggesting that little terrigenous sediment is crossing or accumulating in this region at present [Harris *et al.*, 1996]. Collectively, all information suggests that limited fluvial sediment reaches Ashmore Trough now.

4.1.3. Moresby Trough

[30] To evaluate potential gravity-driven sediment delivery from the shelf into the deepest basin of the study area, we took cores from a prominent submarine channel in the Moresby Trough (Figure 1). This channel was identified during our multibeam and seismic surveys, and extends from the base of the northeastern Pandora Trough slope to the Moresby Canyon in the southeastern part of the GOP [Francis *et al.*, 2008]. This channel is a likely conduit for long-term sediment delivery into the Moresby Canyon, as indicated by thick sequences of siliciclastic sediment in the Coral Sea Basin [Ewing *et al.*, 1970; Winterer, 1970].

[31] Cores collected from a channel in the Moresby Trough, downslope from the Pandora Trough, have low apparent MARs between 0.04 and $0.07\text{ g cm}^{-2}\text{ a}^{-1}$ (Figure 2). One core (MV-24) from this channel displays stratification suggesting possible turbidity current deposition (Figure 3). Excess ^{210}Pb was detected in this core in samples at 12–14 cm, suggesting that the uppermost stratified bed could have been deposited in the past century or so. However, the presence of crosscutting burrows in the X-radiograph of this core (Figure 3) also suggests that this excess ^{210}Pb activity could have been transported to these sediment depths via bioturbation, meaning that the bed is older. A jumbo piston core (MV-22) collected nearby from the same channel-levee complex has been studied by Patterson *et al.* [2006] and is composed of similar muddy turbidites in the upper ~ 2 m of core. The ages of the uppermost muddy turbidites in MV-22 have not been established, but the highly porous muddy sediment texture suggests that the turbidites are relatively recent [Patterson *et al.*, 2006]. These observations suggest that this channel is probably a conduit of sediment transport southward from the shelf edge by means of turbidity currents, and further suggest that such sediment delivery could continue farther into the deep basin via Moresby Canyon.

4.1.4. Sediment Accumulation in the Pandora Trough

[32] In order to estimate apparent annual mass accumulation across the Pandora Trough (the region for which we have the most extensive core coverage), we have calculated a regionally weighted average for mass accumulation rates. A region extending from core MV-43 on the northeastern GOP shelf edge to the northeast corner of Eastern Fields Reef and then to Portlock Reef and the adjacent shelf edge was outlined using GIS software. MAR values from Table 1 were gridded at 5 km resolution using an inverse square distance interpolation algorithm, to produce a region of $19,800\text{ km}^2$ with a spatially averaged MAR of $0.12\text{ g cm}^{-2}\text{ a}^{-1}$, and total annual apparent mass accumulation rate of $\sim 23.7 \times 10^6\text{ t a}^{-1}$. If the carbonate fraction of our total mass accumulation is $\sim 40\%$ of total sediment mass (Figure 6), the terrigenous fraction would thus account for $\sim 14 \times 10^6\text{ t a}^{-1}$, or $\sim 5\%$ of the estimated

Figure 4. Sediment accumulation rate (SAR, cm a^{-1}) and mass accumulation rate (MAR, $\text{g cm}^{-2} \text{a}^{-1}$) of samples versus water depth in the northern Pandora Trough. The SAR and MAR decrease with increasing distance from the shelf and increasing water depth.

annual fluvial sediment flux to the GOP ($\sim 300 \times 10^6 \text{ t a}^{-1}$) [Harris et al., 1993; Milliman, 1995; Milliman et al., 1999].

[33] We also estimated the sediment budget for the northern Pandora Trough using more localized averaging of MAR (Figure 7), to explore MAR variability associated with seabed morphology and locale. The surface area of each accumulation rate region in Figure 7 was determined using ArcGIS (ESRI) software. Spatially averaged MARs were determined using linear inverse distance weighting. Using the localized MAR approach, we estimate a total sediment budget of $\sim 11.3 \times 10^6 \text{ t a}^{-1}$, yielding a terrigenous fraction of $\sim 7 \times 10^6 \text{ t a}^{-1}$ over an area 60% the size of the regional budget area described above. This method highlights the fact that approximately 85% of terrigenous sediment delivery occurs in a relatively small area, along the northeast shelf and upper slope of the study area. Possible sources of error include the values of ^{210}Pb accumulation rates, which we acknowledge to be maximum estimates, as well as MAR variability in regions we did not sample, or sampled only sparsely.

[34] Using a mass accumulation rate of $0.08 \text{ g cm}^{-2} \text{a}^{-1}$ over a smaller region ($\sim 8500 \text{ km}^2$), Walsh and Nittrouer [2003] estimated a regional accumulation rate of about $6.8 \times 10^6 \text{ t}$ of sediment per year, or $\sim 2.3\%$ of the annual terrigenous sediment flux. Considering the uncertainties involved with the above calculations, we consider our results broadly consistent with those of Walsh and Nittrouer [2003]. However, we have chosen to exclude areas with observed sediment accumulation in the Moresby Trough (Figure 2) because of the small number of cores spanning a large and heterogeneous seafloor region. If that region were included, the total quantity of annual sediment accumulation in the deep basins of the Gulf of Papua would be greater than the above estimates.

4.2. The ^{210}Pb Fluxes

[35] Inventories of excess ^{210}Pb vary markedly across the GOP, although they generally decrease with increasing water depth (Figure 5). This spatial variation can be understood by comparing the sources of ^{210}Pb in the region with the fluxes of ^{210}Pb required to support excess ^{210}Pb inventories at steady state conditions (equation (4)).

[36] The primary sources of ^{210}Pb to waters above the shelf break and slope are atmospheric fallout of ^{210}Pb to surface water, and production from decay of ^{226}Ra [Cochran, 1982; Cochran et al., 1990]. By comparison, inputs of dissolved ^{210}Pb from rivers are small, mainly because of rapid scavenging of dissolved Pb by suspended sediments [Benninger, 1978]. Assuming efficient trapping of river-borne sediments on the inner shelf, the total expected inventory of excess ^{210}Pb deposited on the distal seabed is the sum of the two primary sources mentioned above. We refer to the sum of flux from these two sources as the theoretical flux (T).

[37] The mean annual atmospheric flux of ^{210}Pb , measured in rainfall at Townsville, northeastern Australia, between 1964 and 1970, was $0.22 \text{ dpm cm}^{-2} \text{a}^{-1}$ [Bonnyman and Molina-Ramos, 1975]. This average atmospheric ^{210}Pb flux is in good agreement with the ^{210}Pb flux of $0.23 \pm 0.07 \text{ dpm cm}^{-2} \text{a}^{-1}$ reported by Turekian et al. [1977] from the same area. However, J. Pfitzner (unpublished data, 2001) measured ^{210}Pb fluxes in rain collections at the Australian Institute of Marine Sciences (Townsville) and reported a ^{210}Pb flux of $0.33 \text{ dpm cm}^{-2} \text{a}^{-1}$. We note that temporal and spatial variations in rainfall cause fluctuations in the atmospheric ^{210}Pb flux, and we use a value of $0.3 \text{ dpm cm}^{-2} \text{a}^{-1}$ [Brunskill et al., 2003; Pfitzner et al., 2004] as representative atmospheric ^{210}Pb flux for our study area.

[38] We did not measure ^{226}Ra activities in the water column from the study area. Consequently, we estimate water column inventory of ^{226}Ra based on measurements from GEOSECS Station 263 located at $16^\circ 40'\text{S}$; $167^\circ 05'\text{W}$ in the southwestern Pacific, integrated over the depth range at each core location. The profile of ^{226}Ra at this station increases to a maximum above 1500 m water depth, decreases below 2000 m, and activity of ^{226}Ra ranges between 6.0 ± 0.7 and $27.3 \pm 0.6 \text{ dpm } 100 \text{ kg}^{-1}$ from surface to 2500 m water depth [Chung and Craig, 1980]. Water column generation of ^{210}Pb from ^{226}Ra is calculated as the product of ^{226}Ra water column inventory and the ^{210}Pb decay constant (assuming negligible loss of the intermediate decay product ^{222}Rn).

Figure 5. Excess ^{210}Pb inventory versus water depth in the northern Pandora Trough shelf break and slope region in GOP.

Figure 6. Map of calcium carbonate concentrations (% by mass) from core tops after *Febo et al.* [2006]. Lowest carbonate concentrations occur in the region with the highest sediment accumulation rates and most enriched ^{210}Pb inventories (i.e., northernmost continental slope).

[39] For the present discussion, we will define the ratio between the core-calculated flux (F) (equation (4)) and theoretical flux (T) as $R = F/T$. If $R = 1$, then observed inventories are consistent with vertical scavenging of ^{210}Pb from the water column inventory, whereas values of $R > 1$ suggest focusing or lateral import of ^{210}Pb to the area. Similarly, a value of $R < 1$ indicates lateral export of ^{210}Pb from that area [e.g., Buesseler *et al.*, 1985; also see Baskaran and Santschi, 2002].

[40] The highest values of R occur along the upper slope and shelf edge of the northeastern Pandora Trough, where $R \geq 10$ for six cores in water depths of 90–1100 m, and $R > 5$ for three additional cores in the area (Figure 2 and Table 1). In the Ashmore Trough, intermediate values of $2 \leq R \leq 5$ suggest modest enrichment of ^{210}Pb flux with respect to vertical scavenging of in situ production. These cores (MV-10, MV-12, MV-14, and MV-16) were collected seaward of escarpments of Portlock, Boot, and Ashmore reefs. Cores from the Moresby Trough, southwestern Pandora Trough, and the Eastern Plateau display the lowest values of R ($0.9 \leq R \leq 2$), suggesting that in these areas, approximate equilibrium exists between in situ water column production, and seabed inventories. These cores are from the deepest basins,

and are the most remote from the shelf edge and escarpments, in our study (Table 1 and Figure 1).

[41] One potential source of additional ^{210}Pb is the northern extension of the Coral Sea Current [Wolanski *et al.*, 1995], which could provide ^{210}Pb through the process of “boundary scavenging” [Anderson *et al.*, 1994; Moore *et al.*, 1996; Baskaran and Santschi, 2002]. This tropical oceanic water mass is likely to be very low in suspended particulates that could scavenge dissolved ^{210}Pb for export to the seabed. Because GOP is a semienclosed basin, circulation is strongly influenced by the bathymetry and the coastal geometry, with substantial transport across depth contours [Wolanski *et al.*, 1995; Ogston *et al.*, 2008; Slingerland *et al.*, 2008b]. More abundant biogenic and lithogenic particles found closer to the GOP shelf could thus scavenge this dissolved ^{210}Pb advected from offshore waters, and then the particles could settle to the seabed. Boundary scavenging of ^{210}Pb from such horizontally advected water masses might be further enhanced by along-slope barriers to flow, such as the escarpments surrounding Ashmore Trough (Figure 2), and, in the northeastern Pandora Trough, the sharp bend in the shelf edge and pronounced steepening of the slope to the southeast

Figure 7. Pandora Trough sediment budget using MAR values divided into four regions. MAR values decrease with increasing water depth and reflect hemipelagic sedimentation seaward of the upper slope.

(Figures 1, 2, and 7), the two regions where the greatest relative enrichments of ^{210}Pb flux are observed. G. J. Brunskill and J. Pfitzner (unpublished data, 2006) measured an inventory of excess ^{210}Pb from sediment traps at the base of the continental slope that was 10–20 times the expected atmospheric input plus in situ decay supply rates, consistent with our core observations and calculations in Table 1.

[42] An additional potential source of ^{210}Pb to seabed inventories is sediment focusing, as observed in mud deposits off some rivers [e.g., *Bruland et al.*, 1974; *Krishnaswami et al.*, 1975; *Carpenter et al.*, 1981; *Buesseler et al.*, 1985; *Thorbjarnarson et al.*, 1986; *Smoak et al.*, 1996; *Kuehl et al.*, 2004]. SAR and ^{210}Pb inventories tend to display positive correlations in these settings, producing sediment depocenters with ^{210}Pb inventories elevated above values expected from the theoretical vertical flux. Sediment focusing is probably a factor in the northeast Pandora Trough (where SARs are relatively high), more so than in the Ashmore Trough (where SARs are very low). If in situ decay of ^{226}Ra in the water column is a dominant source of ^{210}Pb , inventories would be expected to increase progressively offshore, assuming accumulation rates were uniform. Therefore, from the Pandora Trough into the deeper Moresby Trough, the observed relative downslope decrease in estimated ^{210}Pb fluxes is interpreted to reflect a similar decrease in apparent sediment accumulation rates, resulting from sediment focusing and retention along the shelf edge.

[43] On the basis of recent studies, bathymetric data, and seismic profiles of the shelf and troughs [e.g., *Febo et al.*, 2008; *Francis et al.*, 2008; *Patterson et al.*, 2006; this study], the primary source of siliciclastic material for both Pandora and Moresby troughs appears to originate from the northeastern corner of the GOP. For example, calcium

carbonate content in multicore tops in the GOP illustrate that the lowest carbonate contents (<45 wt%) are observed on the shelf edge and upper slope of the northeast corner of Pandora Trough (pelagic production of carbonate), and along a channel in Moresby Trough (Figure 6) [*Febo et al.*, 2006]. In contrast, calcium carbonate exceeds 55–60% in the southwestern Pandora Trough and amounts to as much as 90% in the northern Ashmore Trough (sources of both neritic and pelagic carbonate) (Figure 6). This pattern is generally consistent with sediment focusing in the northeastern Pandora Trough and decreasing siliciclastic particle flux toward the carbonate-dominated Ashmore Trough (Figures 6 and 8). The most likely source of this sediment is the aforementioned extensive subaqueous clinoform of the GOP inner shelf (Figure 1), produced by fluvial sediment discharged from the Fly, Kikori, and Purari rivers, among others [*Walsh et al.*, 2004]. The shelf edge depocenter, described above near locations of cores MV-42 and MV-47 is probably fed by oceanic transport of inner shelf sediment from the northwest. Sediment accumulation near the shelf edge and steep upper slope may lead to episodic production of turbidity flows that move downslope, toward the channel in the Moresby Trough, as recorded in cores MV-24 (Figure 2) and MV-22 [*Patterson et al.*, 2006].

[44] A network of submarine channels in the Pandora and Moresby troughs, capable of delivering gravity-driven flows from the shelf edge to Moresby Canyon, is shown in Figure 8, along with values of R for individual cores (Table 1). The channel network was extracted from the bathymetric grid of *Daniell* [2008] using RiverTools[©] software (RIVIX LLC) for DEM analysis. The resulting channel network identifies potential pathways for sediment transport that link the locations of cores MV-42, MV-47,

Figure 8. Map of relative enrichment in excess ^{210}Pb inventories from equation (4) and Table 1, showing submarine channel network derived from DEM analysis (channel network data from *Patterson et al.* [2006] with permission). The largest dots indicate highest relative levels of ^{210}Pb enrichment, primarily on the shelf edge and upper slope of the northeast GOP. Patterns of ^{210}Pb enrichment and channel networks outline important sediment transport pathways from the shelf edge to deep basins.

and MV-24, among others, and gives a representation of possible flow pathways in shelf and slope settings where particles may move into the basin by gravity flow transport [*Bird et al.*, 1995].

4.3. A Sediment Transport Scenario

[45] Sediment focusing and boundary scavenging of oceanic ^{210}Pb are both likely sources for producing elevated ^{210}Pb inventories on the shelf break and upper slope. Both processes are probably active in the northeastern Pandora Trough, but boundary scavenging is probably dominant to the southwest, in regions of lower suspended-sediment flux. Spatial distribution of ^{210}Pb flux observed at the seabed can be attributed to downslope and lateral particle transport and variation in bottom topography. This phenomenon has been observed in a number of continental slope environments, e.g.: Quinault Canyon [*Carpenter et al.*, 1981; *Thorbjarnarson et al.*, 1986], California margin [*Huh et al.*, 1990; *Alexander and Venherm*, 2003] and the eastern and southeastern U.S. continental margin [*Biscaye et al.*, 1988; *DeMaster et al.*, 1991]. A certain proportion of particles may be transported from shelf to open slope waters and then down slope, and

the rest may be carried by along-shelf currents to canyon heads, settling into canyons, thus facilitating down canyon transport [e.g., *Biscaye and Anderson*, 1994].

[46] The northeastern GOP shelf edge is marked by several locations that could be major escape routes for sediments from the shelf to deeper basins (Figure 8). For example, cores MV-18 and MV-19 are situated at the head of an intraslope canyon as shown by analysis of seafloor bathymetries and channel networks (Figures 2 and 8). Further, the highest measured excess ^{210}Pb inventory from this study is located at MV-18, along with low surficial calcium carbonate content (Figures 6 and 8). Cores MV-42 and MV-47 also occur along a channel extending from shelf to basin, and both cores exhibit enriched ^{210}Pb inventories and relatively high SAR. Farther down the same channel network, core MV-50 (795 m water depth) exhibits the same characteristics, suggestive of sediment focusing and down-slope transport. Combined seafloor bathymetry and radio-nuclide data thus suggest that these locations are major conduits for sediment supply to deeper basins. These observations, combined with evidence from core MV-24,

further suggest that sediment transport occurs in both nepheloid layers and turbidity flows.

5. Conclusions

[47] Examination of ^{210}Pb activities on the continental margin and basin of the GOP has given some insight into the behavior of ^{210}Pb and the processes that control sediment transfer on the continental margin. Patterns of sediment accumulation ($0.05\text{--}0.35\text{ cm a}^{-1}$; $0.05\text{--}0.28\text{ g cm}^{-2}\text{ a}^{-1}$) and inventory-derived ^{210}Pb fluxes ($1\text{--}12.8\text{ dpm cm}^{-2}\text{ a}^{-1}$) display regional variations, decreasing seaward, and along isobaths away from the northeastern shelf edge. The amount of terrigenous sediment load being discharged annually from the shelf and accumulating in Pandora Trough is approximately $7\text{--}14 \times 10^6\text{ t}$, generally consistent with previous findings. However, the existence of possible turbidity current transport and deposition have been documented in a deep basin, Moresby Trough, suggesting that additional downslope export and sediment accumulation may be occurring in deeper parts of the Coral Sea. High excess ^{210}Pb fluxes estimated from seabed inventories at the shelf break and upper slope are consistent with the combined effects of sediment focusing and boundary scavenging of oceanic water masses. Sediments may be transported from inner shelf depocenters by oceanic processes, focused in depocenters near the northeastern GOP shelf edge, and distributed downslope through a combination of nepheloid layer flow and possible turbidity currents.

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References

- Alexander, C. R., and C. Venherm (2003), Modern sedimentary processes in the Santa Monica, California continental margin: sediment accumulation, mixing and budget, *Mar. Environ. Res.*, **56**, 177–204.
- Anderson, R. F., M. Q. Fleischer, P. E. Biscaye, N. Kumar, B. Dittrich, P. Kubik, and M. Suter (1994), Anomalous boundary scavenging in the Middle Atlantic Bight: evidence from ^{230}Th , ^{231}Pa , ^{10}Be and ^{210}Pb , *Deep Sea Res., Part II*, **41**, 537–561.
- Appleby, P. G., and F. Oldfield (1992), Application of lead-210 to sedimentation studies, in *Uranium-Series Disequilibria: Applications to Earth, Marine, and Environmental Sciences*, edited by M. Ivanovich and R. S. Harmon, pp. 731–778, Clarendon, Oxford, U.K.
- Baskaran, M., and P. H. Santschi (2002), Particulate and dissolved ^{210}Pb activities in the shelf and slope regions of the Gulf of Mexico waters, *Cont. Shelf Res.*, **22**, 1493–1510.
- Benninger, L. K. (1978), ^{210}Pb balance in Long Island Sound, *Geochim. Cosmochim. Acta*, **42**, 1165–1174.
- Benninger, L. K., R. C. Aller, J. K. Cochran, and K. K. Turekian (1979), Effects of biological sediment mixing on the ^{210}Pb chronology and trace metal distribution in a Long Island sound sediment core, *Earth Planet. Sci. Lett.*, **43**, 241–259.
- Bentley, S. J., Z. Muhammad, L. J. Patterson, A. W. Droxler, G. R. Dickens, L. C. Peterson, and B. N. Opdyke (2006a), Modern and Pleistocene turbidite sedimentation in the Gulf of Papua S2S study area: Implications for modulation of sediment sources by oceanic and climatic processes, paper presented at 2006 Annual Convention, Am. Assoc. of Pet. Geol., Houston, Tex., 9–12 April.
- Bentley, S. J., A. Sheremet, and J. M. Jaeger (2006b), Bioturbation, event sedimentation, and preserved sedimentary fabric: Field and model comparisons in three contrasting marine settings, *Cont. Shelf Res.*, **26**, 2108–2124.
- Bird, M. I., G. J. Brunskill, and A. R. Chivas (1995), Carbon isotope composition of sediments from the Gulf of Papua, *Geo Mar. Lett.*, **15**, 153–159.
- Biscaye, P. E., and R. F. Anderson (1994), Fluxes of particulate matter on the slope of the southern Middle Atlantic Bight: SEEP-II, *Deep Sea Res., Part II*, **41**, 459–509.
- Biscaye, P. E., R. F. Anderson, and B. L. Deck (1988), Fluxes of particles and constituents to the eastern United States continental slope and rise: SEEP-I, *Cont. Shelf Res.*, **8**, 855–904.
- Bonnymann, J., and J. Molina-Ramos (1975), Concentrations of lead-210 in rainwater in Australia during the years 1964–1970, Commonwealth X-ray and Radium Lab., Dep. of Health, Melbourne, Australia.
- Bruland, K. W., M. Koide, and E. D. Goldberg (1974), The comparative marine geochemistries of lead-210 and radium-226, *J. Geophys. Res.*, **79**, 3083–3086.
- Brunskill, G. J., K. J. Woolfe, and I. Zagorskis (1995), Distribution of riverine sediment chemistry on the shelf, slope, and rise of the Gulf of Papua, *Geo Mar. Lett.*, **15**, 160–165.
- Brunskill, G. J., I. Zagorskis, and J. Pfitzner (2003), Geochemical mass balance for lithium, boron, and strontium in the Gulf of Papua, Papua New Guinea (Project TROPICS), *Geochim. Cosmochim. Acta*, **67**, 3365–3383.
- Buesseler, K. O., H. D. Livingston, and E. R. Sholkovitz (1985), $^{239,240}\text{Pu}$ and excess ^{210}Pb inventories along the shelf and slope of the northeast U.S.A., *Earth Planet. Sci. Lett.*, **76**, 10–22.
- Carpenter, R., J. T. Bennett, and M. L. Peterson (1981), ^{210}Pb activities in and fluxes to sediments of the Washington continental slope and shelf, *Geochim. Cosmochim. Acta*, **45**, 1172–1181.
- Chung, Y., and H. Craig (1980), ^{226}Ra in the Pacific Ocean, *Earth Planet. Sci. Lett.*, **49**, 267–292.
- Cochran, J. K. (1982), The oceanic chemistry of the U- and Th-series nuclides, in *Uranium Series Disequilibrium: Applications to Environmental Problems*, edited by M. Ivanovich and R. S. Harmon, pp. 384–430, Clarendon, Oxford, U.K.
- Cochran, J. K., T. McKibbin-Vaughan, M. M. Dornblaser, D. Hirschberg, H. D. Livingston, and K. O. Buesseler (1990), ^{210}Pb scavenging in the North Atlantic and North Pacific oceans, *Earth Planet. Sci. Lett.*, **97**, 332–352.
- Cutshall, N. H., I. L. Larsen, and C. R. Olsen (1983), Direct analysis of Pb-210 in sediment samples: Self-absorption corrections, *Nucl. Instrum. Methods Phys. Res.*, **206**, 309–312.
- Daniell, J. J. (2008), Development of a bathymetric grid for the Gulf of Papua and adjacent areas: A note describing its development, *J. Geophys. Res.*, **113**, F01S15, doi:10.1029/2006JF000673.
- DeMaster, D. J., D. C. Brewster, B. A. McKee, and C. A. Nittrouer (1991), Rates of particle scavenging, sediment reworking, and longitudinal ripple formation at the HEBBLE site based on measurements of ^{234}Th and ^{210}Pb , *Mar. Geol.*, **99**, 423–444.
- Ewing, J. I., R. E. Houtz, and W. J. Ludwig (1970), Sediment distribution in the Coral Sea, *J. Geophys. Res.*, **75**, 1963–1972.
- Feblo, L. A., S. J. Bentley, G. R. Dickens, A. W. Droxler, L. C. Peterson, and B. N. Opdyke (2006), Recent to late Pleistocene sedimentary organic matter in the Gulf of Papua, poster presented at 2006 Annual Convention, Am. Assoc. of Pet. Geol., Houston, Tex., 9–12 April.
- Feblo, L. A., S. J. Bentley, J. H. Wrenn, A. W. Droxler, G. R. Dickens, L. C. Peterson, and B. N. Opdyke (2008), Late Pleistocene and Holocene sedimentation, organic carbon delivery, and paleoclimatic inferences on the continental slope of the northern Pandora Trough, Gulf of Papua, *J. Geophys. Res.*, doi:10.1029/2006JF000677, in press.
- Francis, J. M., J. Daniell, A. W. Droxler, G. R. Dickens, S. J. Bentley, L. C. Peterson, B. N. Opdyke, and L. Beaufort (2008), Deep-water geomorphology of the mixed siliciclastic-carbonate system, Gulf of Papua, *J. Geophys. Res.*, doi:10.1029/2007JF000851, in press.
- Goldberg, E. D., and M. Koide (1962), Geochronological studies of deep sea sediments by the ionium/thorium method, *Geochim. Cosmochim. Acta*, **26**, 417–450.
- Harris, P. T., E. K. Baker, A. R. Cole, and S. A. Short (1993), Preliminary study of sedimentation in the tidally dominated Fly River Delta, Gulf of Papua, *Cont. Shelf Res.*, **13**, 441–472.
- Harris, P. T., C. B. Pattiaratchi, J. B. Keene, R. W. Dalrymple, J. V. Gardner, E. K. Baker, A. R. Cole, D. Mitchell, P. Gibbs, and W. W. Schroeder (1996), Late Quaternary deltaic and carbonate sedimentation in the Gulf of Papua foreland basin: Response to sea-level change, *J. Sediment. Res.*, **66**, 801–819.
- Huh, C. A., L. F. Small, S. Niemi, B. P. Finney, B. M. Hickey, N. B. Kachel, D. S. Gorsline, and P. M. Williams (1990), Sedimentation dynamics in the Santa Monica-San Pedro basin off Los Angeles: Radiochemical, sediment trap and transmissometer studies, *Cont. Shelf Res.*, **10**, 137–164.

- Koide, M., A. Soutar, and E. D. Goldberg (1972), Marine geochronology with ^{210}Pb , *Earth Planet. Sci. Lett.*, **14**, 442–446.
- Krishnaswami, S., B. L. K. Somayajulu, and Y. Chung (1975), $^{210}\text{Pb}/^{226}\text{Ra}$ disequilibrium in the Santa Barbara Basin, *Earth Planet. Sci. Lett.*, **27**, 388–392.
- Krishnaswami, S., L. K. Benninger, R. C. Aller, and K. L. Vondamm (1980), Atmospherically-derived radionuclides as tracers of sediment mixing and accumulation in near-shore marine and lake sediments: Evidence from Be-7, Pb-210, and Pu-239, Pu-240, *Earth Planet. Sci. Lett.*, **47**, 307–318.
- Kuehl, S. A., D. J. DeMaster, and C. A. Nittrouer (1986), Nature of sediment accumulation on the Amazon continental shelf, *Cont. Shelf Res.*, **6**, 209–225.
- Kuehl, S. A., G. J. Brunskill, K. Burns, D. C. Fugate, and T. Kniskern (2004), Nature of sediment dispersal off the Sepik River, Papua New Guinea: Preliminary sediment budget and implications for margin processes, *Cont. Shelf Res.*, **24**, 2417–2429.
- Milliman, J. D. (1995), Sediment discharge to the ocean from small mountainous rivers: the New Guinea example, *Geo Mar. Lett.*, **15**, 127–133.
- Milliman, J. D., K. L. Farnsworth, and C. S. Albertin (1999), Flux and fate of fluvial sediments leaving large islands in the East Indies, *J. Sea Res.*, **41**, 97–107.
- Moore, W. S., D. J. DeMaster, J. M. Smoak, B. A. McKee, and P. W. Swarzenski (1996), Radionuclide tracers of sediment-water interactions on the Amazon shelf, *Cont. Shelf Res.*, **16**, 645–665.
- Nittrouer, C. A., R. W. Sternberg, R. Carpenter, and J. T. Bennett (1979), Use of Pb-210 geochronology as a sedimentological tool: Application to the Washington continental-shelf, *Mar. Geol.*, **31**, 297–316.
- Nittrouer, C. A., D. J. DeMaster, B. A. McKee, N. H. Cutshall, and I. L. Larsen (1984), The effect of sediment mixing on Pb-210 accumulation rates for the Washington continental shelf, *Mar. Geol.*, **54**, 201–221.
- NSF MARGINS Office (2004), NSF MARGINS Program science plans, 170 pp., Lamont-Doherty Earth Obs., Columbia Univ., Palisades, N. Y.
- Ogston, A. S., R. W. Sternberg, C. A. Nittrouer, D. P. Martin, M. A. Goñi, and J. S. Crockett (2008), Sediment delivery from the Fly River tidally dominated delta to the nearshore marine environment and the impact of El Niño, *J. Geophys. Res.*, **113**, F01S11, doi:10.1029/2006JF000669.
- Patterson, L. J., S. J. Bentley, D. H. Henry, G. R. Dickens, A. W. Droxler, L. C. Peterson, and B. N. Opdyke (2006), Petrological and geochemical investigations of deep sea turbidite sands in the Pandora and Moresby troughs, source to sink Papua New Guinea focus area, poster presented at 2006 Annual Convention, Am. Assoc. of Pet. Geol., Houston, Tex., 9–12 April.
- Pfitzner, J., G. Brunskill, and I. Zagorskis (2004), ^{137}Cs and excess ^{210}Pb deposition patterns in estuarine and marine sediment in the central region of the Great Barrier Reef Lagoon, north-eastern Australia, *J. Environ. Radioact.*, **76**, 81–102.
- Pickup, G. (1984), Geomorphology of tropical rivers: I. Landforms, hydrology, and sedimentation in the Fly and lower Purari, Papua New Guinea, *Catena Suppl.*, **5**, 18–41.
- Slingerland, R. L., N. W. Driscoll, J. D. Milliman, S. R. Miller, and E. A. Johnstone (2008a), Anatomy and growth of a Holocene clinothem in the Gulf of Papua, *J. Geophys. Res.*, doi:10.1029/2006JF000628, in press.
- Slingerland, R., R. Selover, A. Ogston, T. R. Keen, N. W. Driscoll, and J. D. Milliman (2008b), Building the Holocene clinothem in the Gulf of Papua: An ocean circulation study, *J. Geophys. Res.*, doi:10.1029/2006JF000680, in press.
- Smoak, J. M., D. J. DeMaster, S. A. Kuehl, R. H. Pope, and B. A. McKee (1996), The behavior of particle-reactive tracers in a high turbidity environment: ^{234}Th and ^{210}Pb on the Amazon continental shelf, *Geochim. Cosmochim. Acta*, **60**, 2123–2137.
- Thomson, J., S. Colley, and P. P. E. Weaver (1988), Bioturbation into a recently emplaced deep-sea turbidite surface as revealed by $^{210}\text{Pb}_{\text{excess}}$, $^{230}\text{Th}_{\text{excess}}$ and planktonic foraminifera distribution, *Earth Planet. Sci. Lett.*, **90**, 157–173.
- Thorbjarnarson, K. W., C. A. Nittrouer, and D. J. DeMaster (1986), Accumulation of modern sediment in Quinault submarine canyon, *Mar. Geol.*, **71**, 107–124.
- Turekian, K. K., Y. Nozaki, and L. K. Benninger (1977), Geochemistry of atmospheric radon and radon products, *Annu. Rev. Earth Planet. Sci.*, **5**, 227–255.
- Walsh, J. P., and C. A. Nittrouer (2003), Contrasting styles of off-shelf sediment accumulation in New Guinea, *Mar. Geol.*, **196**, 105–125.
- Walsh, J. P., and C. A. Nittrouer (2004), Mangrove-bank sedimentation in a mesotidal environment with large sediment supply, Gulf of Papua, *Mar. Geol.*, **208**, 225–248.
- Walsh, J. P., C. A. Nittrouer, C. M. Palinkas, A. S. Ogston, R. W. Sternberg, and G. J. Brunskill (2004), Clinoform mechanics in the Gulf of Papua, New Guinea, *Cont. Shelf Res.*, **24**, 2487–2510.
- Winterer, E. L. (1970), Submarine valley systems around Coral-Sea basin (Australia), *Mar. Geol.*, **8**, 229–244.
- Wolanski, E., and D. M. Alongi (1995), A hypothesis for the formation of a mud bank in the Gulf of Papua, *Geo Mar. Lett.*, **15**, 166–171.
- Wolanski, E., A. Norro, and B. King (1995), Water circulation in the Gulf of Papua, *Cont. Shelf Res.*, **15**, 185–212.
- Wolanski, E., R. J. Gibbs, S. Spagnol, B. King, and G. Brunskill (1998), Inorganic sediment budget in the mangrove-fringed Fly River Delta, Papua New Guinea, *Mangroves Salt Marshes*, **2**, 85–98.

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