Late Pleistocene and Holocene sedimentation, organic-carbon delivery, and paleoclimatic inferences on the continental slope of the northern Pandora Trough, Gulf of Papua

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[1] We investigated sediment and organic-carbon accumulation rates in two jumbo piston cores (MV-54, MV-51) retrieved from the midslope of the northeastern Pandora Trough in the Gulf of Papua, Papua New Guinea. Our data provide a first assessment of mass fluxes over the past ~33,000 14C years B.P. and variations in organic-carbon sources. Core sediments were analyzed using a suite of physical properties, organic geochemistry, and micropaleontological measurements. MV-54 and MV-51 show two periods of rapid sediment accumulation. The first interval is from ~15,000 to 20,400 Cal. years B.P. (MV-51: ~1.09 m ka−1 and ~81.2 g cm−2 ka−1) and the second occurs at >32,000 14C years B.P. (~2.70 m ka−1 and ~244 g cm−2 ka−1). Extremely high accumulation rates (~3.96 m ka−1; ~428 g cm−2 ka−1) characterize 15,800–17,700 Cal. years B.P. in MV-54 and likely correspond to early transgression when rivers delivered sediments much closer to the shelf edge. A benthic foraminiferal assemblage in MV-51 from ~18,400 to 20,400 Cal. years B.P. indicates a seasonally variable flux of organic carbon, possibly resulting from enhanced contrast between monsoon seasons. The oldest sediments, >32,000 14C years B.P., contain TOC fluxes >200 g cm−2 ka−1, with >50% of it derived from C3 vascular plant matter. Magnetic susceptibility values are 2 to 3 times higher and benthic foraminiferal accumulation rates are 6 times higher during this interval than at any younger time, indicating a greater influence of detrital minerals and labile organic carbon. The MS data suggest more direct dispersal pathways from central and eastern PNG Rivers to the core site.


1. Introduction

[2] Rivers deliver large amounts of organic carbon (~0.4 × 1013 g C a−1) from land to continental margins, with approximately 30% of it originating from tropical rain forests alone [Schlesinger and Melack, 1981; Hedges et al., 1997]. Understanding the ultimate fate of this important carbon pool is crucial not only for understanding climatic changes, but also for accurately modeling the broader global carbon cycle [Schlünn and Schneider, 2000; Berner, 2003]. Within this context, the Gulf of Papua (GoP) (Figure 1) is a particularly intriguing region to examine the flux and fate of terrestrial organic carbon over glacial to interglacial timescales.

[3] The GoP is a marine depocenter for large and very active river basins. In the GoP, three large rivers, the Fly, Kokori and Purari, and several smaller rivers currently deliver an estimated 300–400 megatonnes a−1 (Mt) of terrigenous siliciclastic material (e.g., clay, quartz, feldspars) from a young, mountainous and tropical region to the inner shelf each year [Milliman, 1995]. This is an annual flux greater than that of the Mississippi River, and comes from drainage basins with combined areas ~3% the size of the Mississippi drainage basin [Milliman and Syvitski, 1992; Milliman, 1995]. With this immense siliciclastic load comes approximately 4 Mt a−1 of terrestrial particulate organic carbon [Bird et al., 1995]. Today, most of the fluvial discharge accumulates on the inner shelf of the GoP and in an extensive mangrove-covered deltaic system along the northern coast (Figure 1); very little of the siliciclastic material and terrestrial organic carbon escapes the shelf to deeper water at present [Bird et al., 1995; Walsh and Nittrouer, 2003].
The modern shelf edge of the GoP lies about 125 m below sea level (Figure 1). However, during late Quaternary sea level lowstands, water depths were ~60 to 125 m below those at present. Consequently, river mouths were at or near the shelf edge, and large amounts of siliciclastic material and terrestrial organic carbon presumably accumulated in deep water settings, as is known for other large river systems [e.g., Flood and Piper, 1997, and references therein]. However, there is no information to evaluate whether massive terrestrial input indeed occurred in GoP slope and basin environments during glacial times, and, if so, what was the fate of that material with respect to burial, transformation, or remineralization.

This study aims to evaluate sediment accumulation in the northeastern GoP since the last glacially induced sea level lowstand. Specifically, we determine: (1) sediment accumulation rates at two representative locations on the middle slope, (2) the late Pleistocene to Holocene organic-carbon contents and fluxes at these locations, and (3) the proportions of terrestrial and marine organic carbon.

2. Gulf of Papua

The GoP is a tropical mixed siliciclastic-carbonate sedimentary system that is presently affected by a warm, wet monsoon-dominated tropical climate and a young,
mountainous continental terrain. The dry season occurs during April to November when SE trade winds dominate, setting up moderate wavefields (~1.3 m significant wave heights) and invigorating the clockwise Coral Sea Coastal Current. In contrast, the wet season occurs during December to March and is characterized by milder NW monsoonal winds that bring significant amounts of rain and lower (~0.3 m) significant wave heights [Thom and Wright, 1983].

[7] The Papua New Guinea highlands receive 10—13 m of annual rainfall that ultimately drains into the GoP [Wolanski et al., 1984]. Coarse sediments discharged by the Fly, Kikori, and Purari Rivers initially accumulate in deltaic sand bodies, whereas suspended sediments are mostly advected clockwise to the northeast by geostrophic flow [Wolanski et al., 1995; Wolanski and Alongi, 1995] and stored along the inner shelf in waters generally shallower than 60 m [Bird et al., 1995; Brunskill et al., 1995; Harris et al., 1996; Walsh et al., 2004; Keen et al., 2006]. Further remobilization may occur during the wave-dominated trade-wind season [Hemer et al., 2004].

[8] The Pandora Trough is a broad basin beyond the shelf edge covering >8000 km² and is thought to receive a limited modern terrigenous sediment flux [Milliman et al., 1999; Walsh and Nittouer, 2003]. The Pandora Trough is rimmed by the GoP shelf that changes from broad and low gradient (1:1000) in the northwest to a narrow and higher gradient (1:133) in the northeast (Figure 1). Sediments escape into the northern Pandora Trough by nepheloid-layer transport downslope [Walsh and Nittouer, 2003] and episodic turbidity currents in some locations [Bentley et al., 2006]. An estimated 2—3% of the sediments entering the GoP each year may be transported off the shelf and deposited in the Pandora Trough [Walsh and Nittouer, 2003; Muhammad et al., 2008].

3. Materials and Methods

3.1. PANASH Cruise and Sediment Cores

[9] The Pandora and Ashmore Troughs are two large sediment sinks that ultimately store sediments shed from the PNG highlands and coastal plain. The PANASH cruise, which took place from March to April 2004, was part of the NSF MARGINS Source-to-Sink Program to sample and quantitatively study this important sedimentary system. During the 2004 field season aboard the R/V Melville, we collected a total of 30 multicores and 33 jumbo piston cores.

[10] Core MV-54 was collected from the slope seaward of the broad shelf (~150 km wide) bordering the central Pandora Trough; core MV-51 was collected from the narrow shelf (~20 km wide) along the northern Pandora Trough (Figure 1). We selected cores on the basis of achieving comparable geologic setting, core length, and water depths to compare and contrast sediment accumulation in different regions of the Pandora Trough. Cores MV-54 and MV-51 are both ~12 m in length and were obtained from 923 and 804 meters water depth in open slope, nonchannelized locations that are not likely to be inundated directly by turbidity currents (based on our multibeam surveys) [Francis et al., 2008] (Figure 1 and Table 1). MV-54 is located on the central Pandora midslope region (Figure 1) just NE of slump scars and between unfilled, relict channels as noted in the work of Francis et al. [2008, Figure 8]. MV-51 was collected on top of a NW-SE trending tectonic ridge that forms a local topographic high in the midslope [Francis et al., 2008, Figure 9]. The numerous tectonic ridges in the northeastern Pandora Trough slope also have troughs between them that serve to trap or divert sediments moving downslope. The ridges are probably sufficiently elevated above troughs to avoid direct inundation by gravity flows (~100 m) (Figure 1, bathymetric profile B), yet still receive suspended sediment transported off-shelf or downslope. Consequently, MV-51 potentially holds a high-quality, high-resolution time-stratigraphic record, owing to the anticipated minimal influence from gravity flows.

[11] Interpretations presented in this manuscript apply specifically to the core locations under investigation. However, the geologic settings of these cores suggest that the cores should contain records of regional processes because of (1) the placement of MV-54 away from surrounding channels and slump scars, (2) because MV-51 is positioned on top of a tectonic ridge, and (3) because both are relatively close to river-influenced shelf edges.

3.2. Shipboard Analyses

[12] After core retrieval, shipboard measurements were made on piston cores for bulk density and magnetic susceptibility using a GeoTek Multi Sensor Core Logger (MSCL). Bulk density was calculated from the attenuation of gamma rays emitted from a 10 milli-Curie 137Cs source through sediment cores. GeoTek MSCL-derived bulk density units are g cm⁻³ and referred to as gamma-ray density (GRD) to separate them from traditional wet and dry bulk density measurements [Weber et al., 1997]. Magnetic susceptibility (MS) was measured using a Bartington MS2C Loop Sensor attached to the MSCL. MS values were corrected with respect to density and units are χ reported in 10⁻⁴ m³ kg⁻¹ [Thompson and Oldfield, 1986]. Both MS and GRD were measured on the MSCL at 1-cm intervals.

[13] Selected sections of MV-51 were imaged via X-radiography. Core subsections 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 grayscale images with 127-micron pixel resolution.

3.3. Radiocarbon Analyses

[14] Radiocarbon dates are based primarily on woody organic matter because of insufficient absolute concentrations of planktonic foraminifera. Nine wood samples from cores MV-54 (n = 4) and MV-51 (n = 5) were analyzed for their radiocarbon content at three labs (Table 2) using accelerator mass spectrometry (AMS). Bulk sediment samples were washed on a 63-μm sieve with distilled water and then at least 10 mg of woody organic matter were hand picked from each for analysis. Only large wood particles (>250-μm) that clearly demonstrated fibrous wood texture.
Table 2. Radiocarbon Ages, Analytical Errors, and Corrections for Each Sample

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Laboratory</th>
<th>Material</th>
<th>$^{14}$C Age</th>
<th>Error (years)</th>
<th>Median Calendar Age, B.P.</th>
<th>Range (years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MV-51 1.20</td>
<td>1</td>
<td>wood</td>
<td>15,565</td>
<td>±50</td>
<td>18,600</td>
<td>650</td>
</tr>
<tr>
<td>MV-51 1.34</td>
<td>2</td>
<td>wood</td>
<td>15,390</td>
<td>±100</td>
<td>18,405</td>
<td>505</td>
</tr>
<tr>
<td>MV-51 3.51</td>
<td>1</td>
<td>wood</td>
<td>20,400</td>
<td>±90</td>
<td>20,405</td>
<td>185</td>
</tr>
<tr>
<td>MV-51 7.40</td>
<td>3</td>
<td>Benthic Foraminifera</td>
<td>29,500</td>
<td>±220</td>
<td>28,017$^c$</td>
<td>220</td>
</tr>
<tr>
<td>MV-51 7.94</td>
<td>1</td>
<td>wood</td>
<td>29,940</td>
<td>±180</td>
<td>out of range</td>
<td>180</td>
</tr>
<tr>
<td>MV-51 10.00</td>
<td>3</td>
<td>wood</td>
<td>34,100</td>
<td>±250</td>
<td>32,617$^c$</td>
<td>250</td>
</tr>
<tr>
<td>MV-51 12.00</td>
<td>1</td>
<td>wood</td>
<td>33,360</td>
<td>±320</td>
<td>out of range</td>
<td>320</td>
</tr>
<tr>
<td>MV-51 1.90</td>
<td>1</td>
<td>wood</td>
<td>13,320</td>
<td>±25</td>
<td>15,850</td>
<td>700</td>
</tr>
<tr>
<td>MV-54 4.75</td>
<td>1</td>
<td>wood</td>
<td>14,160</td>
<td>±35</td>
<td>17,000</td>
<td>500</td>
</tr>
<tr>
<td>MV-54 8.30</td>
<td>1</td>
<td>wood</td>
<td>14,175</td>
<td>±40</td>
<td>17,000</td>
<td>500</td>
</tr>
<tr>
<td>MV-54 9.94</td>
<td>2</td>
<td>wood</td>
<td>14,810</td>
<td>±80</td>
<td>17,740</td>
<td>440</td>
</tr>
</tbody>
</table>

*a*Laboratory numbers are as follows: 1, Keck CCAMS Facility, UC Irvine; 2, Beta Analytic; and 3, National Ocean Sciences AMS at WHOI.

*b*Radiocarbon ages are calibrated to calendar ages (2 sigma 95% probability) using OxCal v. 3.10 [Bronk Ramsey, 1995] and the IntCal 98 calibration curve [Stuiver et al., 1998].

$^c$Corrected by subtracting age difference between planktic and benthic foraminifera in the Western Pacific (Average age difference: 1483 years [Broecker et al., 2004]).

were selected. Because wood and other organic matter may be reworked, particularly during periods of lower sea level, radiocarbon dates taken from wood may represent maximum ages for sediments.

Benthic foraminifera from two samples in MV-51 (Table 2) were also taken for radiocarbon analysis. These ages were reservoir corrected by subtracting the average age difference between planktonic and benthic foraminifera in the western Pacific Ocean (1483 years [Broecker et al., 2004]). Benthic foraminifera were analyzed because commonly used planktonic foraminifera (e.g., *Globigerinoides ruber*, *G. sacculifer*) were not sufficiently abundant.

For six samples, reported radiocarbon ages were corrected for past variations in cosmogenic production using OxCal v 3.1 with the IntCal98 correction curve for dates $\leq 20,000$ $^{14}$C years B.P. [Stuiver et al., 1998]. Five of seven dates from MV-51 remain “uncorrected” because internationally ratified calibration curves do not extend beyond 20,000 $^{14}$C years B.P. [Stuiver et al., 1998]. Calendar ages for these samples may be as much as 4000 to 6000 years older than 20,000–35,000 $^{14}$C years B.P. as a result of increased $^{14}$C production during changes in the Earth’s magnetic field intensity [Hughen et al., 2004; Fairbanks et al., 2005].

3.4. Magnetic Susceptibility

MS is a measure of the strength of induced magnetism experienced by mineral grains in sediments when placed in an applied field. Low-field MS was independently measured on piston-core samples at 5- to 20-cm intervals using a susceptibility bridge in a magnetically shielded room at the LSU Rock Magnetism Laboratory. The MS Bridge uses a balanced coil induction system and is sensitive to 1 ($\pm 0.2) \times 10^{-7}$ m$^3$ kg$^{-1}$ [Ellwood et al., 1996]. This method permits an independent cross check on data collected from the MSCL.

MS values are excellent indicators of the total iron-containing compounds in sediments [Nagata, 1961] and therefore MS is primarily a function of hinterland mineralogy and climate. Terrigenous magnetic components originate from multiple mineralogies including: (1) ferrimagnetic (such as magnetite and maghemite), (2) paramagnetic sources such as clays (e.g., chlorite and illite), iron sulfides (e.g., pyrite), and (3) ferromagnesian silicates (such as biotite, amphibole, and pyroxene). Weakly negative diamagnetic components in sediments include calcite, quartz, and organic matter. Diamagnetic contributions may reduce the overall sediment MS because of a dilution effect, although this is generally relatively small, given the susceptibility of even small amounts of detrital components [Ellwood et al., 2000].

3.5. Calcium Carbonate

Calcium-carbonate content (Figures 3 and 4) was analyzed in MV-51 and MV-54 using the vacuum-gasometric technique [Jones and Kaiteris, 1983] at the LSU Rock Magnetism Laboratory. Dried sediment samples were ground to a fine powder and then a 0.25 g subsample was reacted under a vacuum in a reaction vessel with phosphoric acid for one hour. After complete reaction, the pressure difference was then read from a pressure gauge and used to calculate the sample carbonate concentration using a regression equation. TriPLICATE carbonate measurements on several samples resulted in a precision of $\pm 0.10\%$. Tests of reagent grade (99%) carbonate and a mixed sample standard (95% feldspar 5% carbonate) gave accurate results to $\pm 1\%$.

3.6. Sedimentary Organic Matter

The quantity and type of sedimentary organic carbon offer major evidence for reconstructing paleoclimatic and paleoceanographic change. Organic matter in core sediments was measured using independent analyses. Total organic carbon (TOC) was measured on core sediments using a LECO Carbon analyzer at 20 cm sample intervals for both MV-51 and MV-54. Rock-Eval Pyrolysis (REP) was measured on sediments from MV-51, because this second core contained the longest time-stratigraphic record. TOC and REP measurements were made at Baseline Resolution Analytical Labs, Houston Texas, with analytical errors of $\pm 0.16$ wt.% and $\pm 0.02$ mgHC/gTOC respectively.
TOC data quantify weight percent of organic carbon [Jarvie, 1991] and may serve as an initial proxy for productivity [Berger and Herguera, 1992; Rühlmann et al., 1999]. Mass accumulation rates (MAR) for TOC were then calculated using the accumulation rates in both piston cores:

\[
TOC\ MAR = \frac{\text{TOC}}{\text{DBD} \times \text{LSR}}
\]

\[
TOC\ MAR = \frac{g \times g \times \text{cm}^{-3} \times \text{cm}^{-2} \times \text{kyr}^{-1}}{g \times \text{cm}^{-2} \times \text{kyr}^{-1}} = \frac{g}{\text{cm}^2 \times \text{kyr}}.
\]

where \(\text{TOC}(\text{wt.}%) = \frac{\text{mg} \text{C}_{\text{org}}}{\text{g} \text{sed}} \times 100\), \(\text{DBD} = \text{Dry Bulk Density} \) (g cm\(^{-3}\)), and \(\text{LSR} = \text{Linear Sedimentation Rate} \) (cm kyr\(^{-1}\)). REP helps to determine the principle sources of the TOC by examining the amount of free (S1) and pyrolytic (S2) hydrocarbons preserved in sediments. REP was developed as a screening method to determine the hydrocarbon potential of petroleum source rocks [Tissot and Welte, 1984; Rullkötter, 2000], but it is becoming widely used in paleoceanography [Lallier-Vergès et al., 1993; Meyers, 1997; Bouloubassi et al., 1999; Rullkötter, 2000; Wagner, 2000]. Two REP parameters are used in this study, the hydrogen index (HI) and the thermal maximum (Tmax). The hydrogen index is the mass of hydrocarbons per gram of TOC and is determined from heating kerogen in a sample until the maximum amounts of hydrocarbons are evolved [Espitalié and Bordenave, 1993]. Equation (2) was used to calculate the hydrogen index. Tmax is the temperature during pyrolysis at which the most hydrocarbons are evolved.

\[
HI = \frac{100 \times S2}{TOC} = \frac{\text{mgHC}}{\text{g}} \times \frac{100 \text{g}}{\text{g}} = \frac{\text{mgHC}}{\text{g} \text{TOC}},
\]

where \(HI = \text{Hydrogen Index}\) and \(S2 = \left(\frac{\text{mgHC}}{\text{g} \text{sediment}}\right)\).

[21] Hydrogen index values are generally low (≤150 mgHC gTOC\(^{-1}\)) for organic matter composed of vascular-land-plant carbon in recent deep-sea sediments, and are characterized as Type III and Type IV kerogen [Tissot and Welte, 1984; Meyers, 1997]. Algal-rich marine organic-matter values range from 200 to 400 mgHC gTOC\(^{-1}\) for recent marine sediments and are characterized as Type Ila kerogen [Rullkötter, 2000]. Tmax values of modern “immature” organic carbon are below 430°C for kerogen types III and IV, whereas “overmature” organic carbon from reworked fossil organic matter are >430°C [Delvaux et al., 1990].

[22] More precise determination of sedimentary organic-carbon types was achieved by analyzing the C/N and \(\delta^{13}C\) ratios on selected samples from MV-51. Isotopic analyses were performed on a Thermo Finnigan Delta Plus XP Isotope Ratio Mass Spectrometer with an online elemental analyzer at the Coastal Ecology Institute, LSU. Dry sediment samples were first acidified with 10% HCl until the reaction with calcium carbonate ceased, or approximately 10 min. Sediments were then acidified with 50 mL concentrated HF for 8 hours to remove minerogenic matter and isolate the kerogen. Residues were centrifuged and rinsed with distilled water several times until completely neutralized, then dried at 50°C and powdered for analysis. We also made slides from this isolated kerogen for later optical palynofacies analysis. Standard deviations of duplicate measurements show analytical precision of \(\delta^{13}C\) to be ±0.15‰ and C/N ratios to ±0.20 molar percent.

[23] The terrigenous component of TOC was determined from \(\delta^{13}C\) values using a two-source mixing model in which the marine (−20.5‰ ± 0.5‰) and terrigenous (−26.5‰ ± 0.5‰) isotope values are used as end-members [Bird et al., 1995; Aller and Blair, 2004]. Equations (3)–(5) are adapted from Amo and Minagawa [2003] and Bird et al. [1995].

\[
f_{\text{terrigenous}} = \left(\frac{\delta^{13}C_{\text{measured}} - \delta^{13}C_{\text{marine}}}{\delta^{13}C_{\text{terrestrial}} - \delta^{13}C_{\text{marine}}}\right)
\]

\[
\% \text{ Terrestrial TOC} = f_{\text{terrigenous}} \times \text{TOC}
\]

\[
= \left(\frac{\delta^{13}C_{\text{measured}} - \delta^{13}C_{\text{marine}}}{\delta^{13}C_{\text{terrestrial}} - \delta^{13}C_{\text{marine}}}\right) \times 100
\]

\times \text{wt.} \% \text{TOC}.
\]

3.7. Benthic Foraminifera

[24] Benthic foraminiferal assemblages are very useful paleoproductivity proxies because their abundances and assemblage compositions are linked to the flux of organic carbon arriving at the seabed. Several studies have demonstrated that foraminiferal densities are positively correlated with changes in nutrient fluxes, surface water productivity, and the subsequent flux of that material to the seabed [e.g., Gooday, 1988; Jorissen et al., 1992; Sjoedsma and van der Zwaan, 1992; Loubere, 1996]. Certain assemblages of benthic foraminifera are adapted to particular types of organic-carbon flux, such as highly variable or “seasonal” flux that is characteristic of upwelling zones compared to relatively constant flux [Loubere and Farriudin, 1999; Smart and Gooday, 1997; Licari and Mackensen, 2005]. We use benthic foraminifera as an independent metric of organic-carbon flux, as well as the stability of flux through time.

[25] In this study, eight to ten grams of 36 freeze-dried samples from core MV-51 were gently washed with distilled water over a 63 μm sieve and dried at 50°C. The >63 μm fraction was then weighed and further split into 150–250 μm and >250 μm fractions. At least 300 benthic foraminifers were then picked from each of the fractions >150 μm, so that ~600 specimens were counted from each sample. Specimens were placed on assemblage slides, sorted, and then enumerated. Assemblage data are routinely collected from the >150 μm fraction in ecological and paleoceanographic studies [e.g., Pérez et al., 2001] because ecologically important taxa may be missed by examining only the >250 μm fraction [Sen Gupta et al., 1987]. Examination of these fractions also permits comparison with other studies. The >63 μm fraction was inspected for all sieved samples in MV-54 to determine presence or absence of foraminifera.

[26] Benthic foraminiferal densities (BF number = total shells or specimens g\(^{-1}\)) and benthic foraminiferal accumulation rates (BFAR; number of specimens per cm\(^{-2}\) ka\(^{-1}\)) were calculated from absolute foraminiferal densities [Herguera, 1992; Herguera and Berger, 1994]. Groupings
null
Figure 3. Linear sedimentation rate (LSR) and Mass Accumulation Rate (MAR) data. Downcore gamma-ray density (GRD) and dry bulk density (DBD) are plotted (second plot) with LSR and MAR (third plot) for MV-51. MAR data are calculated from DBD and LSR. Grain-size and calcium-carbonate data are shown in the fourth plot with total organic-carbon (TOC) and TOC MAR shown in the fifth plot. Magnetic susceptibility (MS) data are shown in the first plot for the core logger Bartington Loop Sensor (labeled Bartington) and the low-field laboratory measurements labeled MS.
Figure 4. Linear sedimentation rate (LSR) and Mass Accumulation Rate (MAR). Downcore gamma-ray density (GRD) and dry bulk density (DBD) are plotted (second plot) with LSR and MAR (third plot) for MV-54. MAR are calculated from DBD and LSR. Grain-size and calcium-carbonate data are shown in the fourth plot with total organic-carbon (TOC) and TOC MAR shown in the fifth plot. Magnetic susceptibility (MS) data are shown in the first plot for the core logger Bartington Loop Sensor (labeled Bartington) and the low-field laboratory measurements labeled MS.
Espitalié, Katz, Langford and Blanc-Valleron.

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4.4. Calcium Carbonate and Volcanic Ash Layers

Siliciclastic particles dominate sediments from cores MV-51 and MV-54. Samples in MV-51 gave very low calcium-carbonate percentages (mean 3.8%, n = 68) whereas values are slightly lower in MV-54 (mean 3.0%, n = 37). Calcium carbonate in MV-54 occurs only in the upper 3.6 m and is completely absent below 4.0 mbsf (Figure 4).

Several volcanic ash layers are present in MV-51. MS maxima at 2.4, 4.8, and 6.9 mbsf are a result of elevated levels of paramagnetic grains such as biotite, hornblende, and amphibole that were identified from sieved samples (>63 μm) using a binocular microscope. These grains are accompanied by abundant pumice and glass shards and result in sharp increases in the sand-sized fraction (>63 μm).

4.5. Sedimentary Organic Matter

Organic-carbon content in MV-51 and MV-54 is relatively low overall and at site MV-51, appears largely to have a source from land-plant debris. TOC in core MV-51 ranges from 0.41 to 0.97 wt.% (mean 0.75), generally lower than values in core MV-54, which range from 0.63 to 2.29 wt.% (mean 0.92). In MV-51, TOC is highest from 1.4 to 1.8 mbsf and >7.4 mbsf, or during 18,400–20,400 Cal. years B.P. and >28,000 14C years B.P., respectively. MAR of TOC is also highest during these time intervals (Figure 3).

MV-54 shows nearly uniform values in TOC over most of the core length. Four prominent subsurface maxima in excess of 1.3 wt.% occur at 1.8, 3.2, 6.2, and 7.0 mbsf and correspond to maxima in both MS and the sand-sized fraction (Figure 4). Sediments at 1.5–1.9 mbsf and at 6.2 mbsf are composed of a dark, sandy material, with abundant and well preserved pteropods, foraminifera, and woody material. Organic particles are common in all sieved samples from MV-54, many of which are unpyritized and can be identified as plant cuticle and woody cortex under binocular microscopy. One sample from 7.0 mbsf contained tree resin (amber) and a complete leaf (~2 mm long) with well preserved venation.

Hydrogen index values of sedimentary organic matter from MV-51 are very low (7 to 152.6 mgHC gTOC⁻¹; average 74) and contain very low S2 pyrolytic yields (Figure 5a). These data reflect hydrogen-poor organic carbon and therefore may indicate either type III organic matter or a mixture of type III and oxidized type II marine organic matter. Figures 5a and 5b show the data plotted in two different diagrams, both generally plotting below the Type III terrestrial-organic-carbon thresholds. However, low S2 values, below 1 mgHC g sediment⁻¹, may be a consequence of mineral matrix dilution of hydrogen-poor hydrocarbons, which makes it difficult to accurately determine the organic-matter type [Espitalié et al., 1980; Katz, 1983; Langford and Blanc-Valleron, 1990]. No correlation exists between hydrogen index and TOC values, as expected given the very low hydrogen index values and narrow range in TOC values [Bouloubassi et al., 1999].

Tmax measurements generally indicate that MV-51 sediments contain immature organic carbon, as expected for recent marine sediments. However, six samples from 3.0, 3.2, 4.6, 5.2, 8.8, 9.8 mbsf (Figure 5) indicate overmature organic carbon, which probably had a source from reworked sedimentary rocks infield (e.g., coal). There does not appear to be any stratigraphic significance to the timing of reworked organic matter in MV-51.

The sand-sized fractions (>63 μm) comprise as much as 85% of the sediment at 1.5–1.9 mbsf and 25% at 6.20 mbsf (Figure 4). Sediments are generally coarser in MV-54 of the sediment at 1.5–1.9 mbsf and 25% at 6.20 mbsf. Additional, the pumice-rich layer at 7.4 mbsf, or during 28,000 14C years B.P. and >28,000 14C years B.P., respectively. MAR of TOC is also highest during these time intervals (Figure 3).

Organic particles are common in all sieved samples from MV-54, many of which are unpyritized and can be identified as plant cuticle and woody cortex under binocular microscopy. One sample from 7.0 mbsf contained tree resin (amber) and a complete leaf (~2 mm long) with well preserved venation.

TOC is also highest during these time intervals (Figure 3).
and deeper than 7.4 mbsf (Figure 6). Three terrestrial input maxima occur in MV-51, one at 1.80 mbsf (62%), another at 8.08 mbsf (55%), and the third at 10 mbsf (57%). A strong positive correlation exists between TOC concentrations and percent terrigenous TOC ($r^2 = 0.73$, $p < 0.001$; $n = 24$), indicating that increases in TOC concentrations result from enhanced input of continental sources and not from marine primary productivity.

[42] Elemental and stable-isotope data provide further constraint on organic-carbon sources. A plot of elemental (atomic C/N) and stable isotopic ($\delta^{13}$C, %) values from MV-51 illustrates two groupings of data (Figure 7). The first group (0.4–6.8 mbsf) plots closer to modern marine organic-matter values than does the second group (core top, and >7.4 mbsf). This second group (core top, and >7.4 mbsf) clusters nearest to the region of C3 land plants. Atomic C/N and stable isotopic data generally fall on a mixing line between the two sources, except for one point from an ash layer at 6.9 mbsf that rests well below the mixing line (Figure 7). It is unclear why the value from the ash layer forms an outlier, but it may be a result of significantly different organic-matter source transported in the ashfall debris.

4.6. Benthic Foraminifera

[43] Benthic foraminifera are abundant in MV-51 sediments. Absolute densities and BFAR (Figure 8) generally give the same trends but with some differences in the lower seven meters. Lowest densities of foraminifera (average $\sim$ 98 specimens g$^{-1}$, $n = 22$) occur shallower than 6.8 mbsf, whereas densities between 6.8 and 12 mbsf are nearly 2 times higher (average $\sim$ 174 specimens g$^{-1}$, $n = 14$; Figure 8). A sample at 6.8 mbsf contains the highest density of foraminifera in MV-51 ($\sim$283 specimens g$^{-1}$). In contrast, lower BFAR (average $\sim$ 5800 cm$^{-2}$ ka$^{-1}$, $n = 31$) characterize core depths less than 10 mbsf, except for the maximum at 6.8 mbsf (12,600 cm$^{-2}$ ka$^{-1}$). Deeper than 10 mbsf, benthic foraminiferal densities are 6 times higher than in samples less than 10 mbsf (average $\sim$ 34,800 cm$^{-2}$ ka$^{-1}$, $n = 5$; Figure 8). The significance of benthic foraminiferal densities, BFAR, and organic-carbon flux at site MV-51 (Figure 8) was examined with multiple linear regression analysis.

[44] A statistically significant association exists between total benthic foraminiferal densities >150 $\mu$m and TOC concentrations ($r^2 = 0.21$; $p = 0.003$; $n = 35$). However, the low correlation coefficient suggests that while TOC plays a role in the densities of foraminifera, other factors are important as well. If the mass accumulation rates are incorporated into the raw data so that TOC MAR and BFAR are calculated, a much stronger correlation of analytical data exists ($r^2 = 0.90$; $p < 0.001$; $n = 35$). These statistics emphasize the importance of incorporating sediment accumulation rates with raw foraminiferal data [Herguera, 1992]. Together, they show that the mass accumulation of organic carbon and foraminifera are intrinsically linked, and therefore that BFAR is a reflection of organic-carbon flux.

[45] Two periods of high BFAR and high TOC MAR occur in MV-51 (Figure 8). The first interval is 3.1–4.8 mbsf and corresponds to $\sim$18,400–20,400 Cal. years B.P., or approximately the late Stage 3 interstadial. The latter period contains the highest BFAR and TOC MAR in the entire core. In addition to total BFAR, important paleoenvironmental information is contained in the accumulation rates for groups of benthic foraminiferal species.

[46] R-mode cluster analysis divided benthic foraminiferal species into two distinct groups (Figure 9). The first cluster is characterized by low-abundance taxa including *Oridotarsalis tener,* and *Melonis barleeanus.* Species of this cluster rarely exceed 1–5 specimens per gram or 150 specimens cm$^{-2}$ ka$^{-1}$. The second cluster is composed of ten taxa (Figure 9) with three distinct subclusters being the most abundant and paleoceanographically significant. The first subcluster is composed of *Uvigerina peregrina/hispida* and *Cibicides pachyderma,* and *Bolivina robusta* as a companion taxon (Figure 9). The second cluster is composed of *Bulimina aculeata,* *Sphaeroidina bulloides,* and *Bolivinida quadrilatera* as a companion taxon. The more distant third cluster contains *Gavelinopsis translucens* and *Globocassidulina subglobosa.* Species distributions in Figure 9 are arranged in order of the clusters described above.

[47] Three distinct species BFAR intervals are present in MV-51 (Figure 8, shaded areas). The first interval (10–12 mbsf; >32,000–33,000 $^{14}$C years B.P.) is characterized by elevated abundances of nearly all foraminiferal species with the exception of *M. barleeanus* (Figure 8). The most abundant taxa are, in order of decreasing abundance, *B. robusta,* *U. peregrina/hispida,* *S. bulloides,* *Bulimina aculeata,* *C. pachyderma,* and *O. tener.* The second interval is composed of a single sample at 6.8 mbsf ($\sim$28,000 $^{14}$C years B.P.) and is composed chiefly of *B. robusta,* *U. peregrina/hispida,* and *C. pachyderma.* The third abundance interval occurs from 1.4 to 3.4 mbsf ($\sim$18,400–20,400 Cal. years B.P.) and is composed mostly of *B. robusta,* *G. translucens,* *G. subglobosa,* *O. tener,* and *M. barleeanus.* Overall, the most abundant species throughout MV-51 is *B. robusta.*

[48] Large sections of MV-51 contain fewer benthic foraminifera and stand in contrast to previously mentioned high BFAR and abundance intervals. In particular, two intervals 0.1–1.4 and 3.4–6.8 mbsf contain the lowest BFAR values (Figure 8). The interval 0–2.4 mbsf (<18,400 Cal. years B.P.) is characterized by *B. robusta,* *U. peregrina/hispida,* *Bulimina aculeata,* and *G. subglobosa,* in order of decreasing average BFAR values. In contrast, the second interval 3.4–6.8 mbsf (20,400 Cal. years B.P. to <28,000 $^{14}$C years B.P.) is composed mostly of *B. robusta,* *G. subglobosa,* *Bulimina aculeata,* and *G. translucens.* The interval 7–10 mbsf (>28,000 to <32,000 $^{14}$C years B.P.) shows intermediate BFAR values and is composed of the same dominant species as for 10–12 mbsf.

[49] The Q-mode dendrogram shows the clustering of samples based on the BFAR values for 12 species at each depth interval (Figure 10). In particular, the stratigraphic record is grouped according to BFAR and the species assemblages discussed above.

5. Discussion

5.1. Magnetic Susceptibility and Climate

[50] Variations in MS signals primarily reflect the processes governing terrestrial erosion and transport into the marine environment, largely the result of climate change...
Figure 6. Magnetic susceptibility and organic geochemical data are plotted with radiocarbon ages for MV-51. The data plotted in the percent terrestrial TOC profile are derived from stable-carbon isotopic data.
Ellwood et al.

Ellwood et al. = 0.002, \( p = 0.29, n = 70 \)

Ellwood et al. = 0.01, \( p = 0.548, n = 37 \).)

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Organic geochemical data from MV-51 are

Ellwood and Ledbetter = 0.02, \( p = 0.255, n = 68 \); MV-54: \( r = 0.025, p = 0.195, n = 68 \) or MV-54 (\( r = 0.029, p = 0.311, n = 37 \)). Additionally, no correlation exists between TOC and calcium carbonate content (MV-51: \( r^2 = 0.02, p = 0.255, n = 68 \); MV-54: \( r^2 = 0.01, p = 0.548, n = 37 \)).

Eroded detrital minerals dominate the MS signal, and their production, fate and transport are controlled by mineral provenance and climate. We suggest that changes in river source(s) and/or climatic may explain MS patterns in MV-51. The most notable pattern in the MS data for MV-51 is the major change at 10 mbsf (Figure 6). We propose some possibilities to explain the higher MS values on the basis of the surrounding PNG geology.

5.1.1. PNG Geology

Papua New Guinea is composed of a complex, heterogeneous geological terrain consisting of predominantly folded sedimentary rocks in western PNG and volcanic rocks in the central eastern PNG region [Rickwood, 1968; Davies and Smith, 1971]. The Fly Strickland River drainage basin in western PNG erodes predominantly limestone, siltstone, and sandstone lithologies [Rickwood, 1968, Brunskill, 2004] and as a result discharges relatively mature sediments with high quartz/feldspar ratios (Figure 11, fluvial source 1) [Slingerland et al., 2008]. In contrast, central and eastern PNG are dominated by metamorphic and volcanic rocks that result in immature sediment lithologies draining from the other PNG rivers, such as the Turama-Kikori-Purari Rivers (Figure 11, fluvial source 2) [Davies and Smith, 1971; Slingerland et al., 2008].

In eastern PNG, the Owen Stanley Range is a large metamorphic region that forms the prominent mountainous spine of the PNG highlands (Figure 11). Metasedimentary rocks from the Owen Stanley Range consist of greenschist, slate, and phyllite, and are interrupted by numerous igneous intrusions [Davies and Smith, 1971; Blake, 1976]. Basaltic, volcanic rocks intrude the Owen Stanley Range and compose a 1000-km belt of Cretaceous to Pliocene rocks that are potassium rich and silica undersaturated [Mackenzie, 1976; Smith, 1982]. Although the precise mineralogies of these metamorphic and igneous rocks are not well documented, a number of studies suggest that these rocks are mineralogically distinct from the dominantly sedimentary and intrusive andesitic volcanic rocks in western PNG [Rickwood, 1968; Mackenzie, 1976]. Therefore, we hypothesize that in the past, river sediments from central and eastern PNG (Figure 11, fluvial sources 2–3) may have been more important than the Fly Strickland as a source of iron-bearing clastics.

During the Last Glacial Maximum (LGM), lowered sea level in the GoP enabled river dispersal systems to incise across the exposed shelf to the shelf edge [Harris et al., 1996]. Given the proximity of MV-51 to central and eastern PNG, we suggest that the Turama-Kikori-Purari river system may have incised a more direct pathway to MV-51 so that high MS values below 10 mbsf result from a more direct input of volcaniclastic material from central and eastern PNG (Figure 11, fluvial sources 2–3). The Turama-Kikori-Purari river system would have had far less distance to travel to the northeastern GoP than the Fly River. This hypothesis is testable in the future by collecting quantitative mineralogical data from MV-51 and comparing them to different PNG river sediments. Another test could include age dating amphiboles from above and below 10 m. For

Figure 7. Organic geochemical data from MV-51 are plotted in an elemental (atomic C/N) and stable-isotopic data cross plot. Data (circles) generally plot along a theoretical mixing line (dashed line) between two sources, modern marine organic matter and C3 land plants. OM refers to organic matter and SOM refers to soil organic matter. Plot after Meyers [1997] and sources mapped from values reported by Meyers [1994] and Goni et al. [2006].

and eustacy [Tite and Linington, 1975; Verosub et al., 1993; Banerjee, 1996; Ellwood et al., 1996, 2000] and biological productivity [Ellwood and Ledbetter, 1977; Mead and Tauxe, 1986]. Fluvial and airborne delivery mechanisms account for most detrital input and are themselves primarily functions of precipitation and aridity, respectively. Volcanic ashfall may also provide ferri- and para-magnetic compounds to marine sediments. Also affecting the MS, albeit to a lesser extent, are fluxes of diamagnetic calcium carbonate, biogenic silica, and organic matter to the seabed, which are functions of primary productivity in the surface water and therefore also of nutrient supply.

The effects of climate change and provenance on MS records are well documented for terrestrial and marine sediments [e.g., de Menocal et al., 1991; Heller and Evans, 1995; Ellwood et al., 1999; Weedon et al., 1999; Balsam et al., 2005]. Precipitation is one of the principal drivers of chemical and physical weathering, particularly in the monsoon-dominated tropics. Periods of increased precipitation are generally associated with enhanced rates of runoff and erosion of weathered detrital particles. Likewise, drier conditions are less favorable for erosion and transport of weathered material, particularly during glacial ice advances, because they remove water vapor available for erosion. During eustatic sea level lowstands, river mouths may prograde across exposed shelves to create coastlines that are much closer to shelf breaks, and therefore may deliver greater amounts of detrital sediment to marine basins.

MS is not strongly influenced by diamagnetic minerals and therefore reflects a dominantly detrital signal. Linear regression analysis was used to test MS values against other variables such as TOC and calcium carbonate in cores MV-51 and MV-54. No association exists between TOC and M$S$ in either MV-51 (\( r^2 = 0.002, p = 0.29, n = 70 \)) or MV-54 (\( r^2 = 0.004, p = 0.26, n = 62 \)). There is also no correlation between calcium carbonate and MS in either
Figure 8. Profiles of total benthic foraminiferal accumulation rates (BFAR) and for individual taxa in relation to TOC, TOC MAR, coarse fraction percentage, and calcium carbonate content. Benthic foraminiferal concentrations are denoted as the BF number, or the total number of individual specimens per gram sample. Three zones of increased BFAR are highlighted in gray and are indicative of enhanced organic-carbon flux.
example, Pleistocene age dates would indicate that amphiboles came from younger volcaniclastic sources of western PNG and older ages (e.g., Cretaceous-Pliocene) would indicate a source from the Papuan Peninsula.

5.1.2. Volcanic Ash Layers

The source(s) of MV-51 ash layers remains uncertain because few ash layers have been reported in and around PNG that date to this time interval. Some ash deposits (e.g., Tomba Tephra) have been studied around Mt. Hagen in western PNG and dated to ~28,000–50,000 years B.P. [Pain and Blong, 1976; Chartres and Pain, 1984]. Other ash deposits, such as the Sagamasi and Natanga Tephras, have been examined around Mt. Lamington in eastern PNG and dated as 15,000 ± 500 to 20,100 ± 600 Cal. years B.P. [Ruxton, 1966].

Our radiocarbon dates provide some constraint on the ash layers present in MV-51. The ash layer at 2.4 mbsf is well constrained between ~18,400 and ~20,400 Cal. years B.P. The ashes at 4.8 and 6.9 mbsf are not as well constrained but were deposited sometime between ~20,400 Cal. years B.P. and ~28,000 14C yrs B.P. Mt. Lamington may have been a source for the ash layer at 2.4 mbsf but there are not enough dated ash layers of the correct time frame on PNG to hypothesize the source for the older ash layers. Therefore, they may originate from any of the active late Quaternary volcanic peaks shown in Figure 11. One exception is Madilogo north of Port Moresby, because it is thought to have formed during the past ~1000 years [Blake, 1976]. These ash layers may serve as correlation points in future GoP studies over this time interval.

5.2. Organic Geochemical Data

Organic geochemical proxies suggest that there were significant variations over time in the paleoflux of organic carbon and the types of sedimentary organic matter buried in the northeastern Pandora Trough. These data illustrate two important points. First, TOC and hydrogen index values of sediments are consistently low. Secondly, stable isotopic and elemental data indicate a complex mixture of marine and C3 vascular plant matter in the organic carbon, with a potential contribution of other terrestrial plant matter.

Terrestrial plant matter is common in the modern GoP and therefore it is not surprising that hydrogen index values are indicative of refractory types III and IV organic matter. Small, mountainous rivers are very important to the GoP [Milliman, 1995] and on average, 65% of river-borne terrestrial organic carbon is refractory [Ittekkot, 1988]. We frequently observed floating megadetritus (branches to whole tree trunks) in the northeastern GoP during the cruise. In addition, Robertson and Alongi [1995] estimated that as much as 9 Mt C a⁻¹ of floating macro and megadetritus occur in the surface water of the Fly delta, and that significant quantities are also found on the shelf and in deep-sea troughs of the GoP. In addition, sediments with hydrogen-poor organic matter (i.e., terrestrial plant matter) are susceptible to much lower hydrogen indices because the hydrocarbons are retained on mineral matrix during pyrolysis [Katz, 1983; Langford and Blanc-Valleron, 1990].

Low hydrogen index values in MV-51 are consistent with recent studies of such fluvial-marine dispersal systems [Aller and Blair, 2004; Aller et al., 2008; Goni et al., 2008]. These systems can promote efficient remineralization of labile organic compounds, particularly during sediment reworking on the inner shelf, and leave only refractory organic matter. Therefore, it is possible that a greater amount of labile organic carbon was originally delivered...
to MV-51, but that it was later removed from the burial record by oxidation.

We note that trends in pyrolytic hydrogen index data do not correlate consistently with TOC and stable-isotope-derived percentage of terrestrial TOC (Figure 7). One explanation is that the hydrogen-index data have been biased by hydrogen-rich (algal) labile organic matter in sediments and their oxidation to lower hydrogen index values to reflect type III organic matter [Meyers, 1997].

Palynomorph observations indicate common phytoclasts such as pollen, spores, leaf and grass cuticles, but rare marine palynomorphs such as resistant dinoflagellate cysts. However, observations also indicated an abundance of amorphous organic matter, which is derived from degraded marine organic matter and supports the presence of oxidized labile compounds [Levan, 1986].

Stable-carbon isotopic data indicate that TOC contains >50% marine organic carbon throughout much of MV-51.
51. However, the mixing model only takes into consideration two sources, marine and C3 plants. There is also the possibility of an influence from vascular C4 plant matter, soil organic matter upon which C3 and C4 plants grew, or also freshwater algae from rivers and estuaries (Figure 7). In addition, it is also possible that the values for the marine end-member used in the mixing equation (−20.5%±0.5%) changed during the recent geologic past to a more enriched value (e.g., −19.5%), which would cause the percentage to change toward more (>50%) terrestrial fractions. At this time, we are unsure about a dominance of marine organic matter as suggested by the stable isotope data and recognize that the estimates of terrestrial sedimentary organic carbon may be minimums. [s4] While keeping in mind the assumptions made in the simple, two-end-member mixing model, the stable-isotopic and elemental data illustrate a mixture of marine and predominantly C3 vascular plant matter (Figure 7). C3 plants are arborescent/herbaceous angiosperms and gymnosperms that occur in a wide range of settings where precipitation is abundant, such as warm and wet tropical rain forests (e.g., mangroves [Tyson, 1995]). C4 plants are fast growing, herbaceous angiosperms (e.g., grasses) that prefer warm and arid tropical environments such as savannah grasslands [Tyson, 1995; Wagner, 2000; Wagner et al., 2004]. C3 plants are geochemically distinct from C4 plants in that they are ~10–15% depleted in 13C (−27‰) compared to C4 plants (−14‰) [Tyson, 1995; Meyers, 1997; Wagner et al., 2004]. Because of these geochemical distinctions, we suggest that a large proportion, though not all, of the vascular plant matter is from C3 plants.

[s5] Many global climate records indicate that LGM was both a cooler and drier time than present; even in the warm tropics [Thompson et al., 1995; Broecker, 1996; Lea et al., 2000]. Therefore, we had hypothesized that during LGM, a grassland savannah may have covered much of the PNG coastal plain as it did in northeastern Queensland [Kershaw, 1978, 1988]. However, none of the samples measured from MV-51 exhibit an obvious C4 contribution during LGM. Instead, values show a C3 signature that seems to indicate that wet conditions and C3 vegetation persisted in PNG for the past ~33,000 14C years B.P.

[s6] Terrestrial climate records from the GoP are mainly from the Papuan Highlands [Bowler et al., 1976; Hope, 1976; Hope and Tulip, 1994], though some records exist from the coastal plains [Garrett-Jones, 1979]. Pollen records from the PNG highlands indicate an atmospheric cooling of 6o–10C from 40,000 to 30,000 14C years B.P. and during LGM from 25,000 to 15,000 Cal. years B.P. [Bowler et al., 1976; Webster and Streten, 1978]. During the past ~10,000 Cal. years B.P. the coastal plain in northern PNG, adjacent to the Solomon Sea, was similar to today in that it was covered by a lower Montane Rain forest with some savannah grasses [Garrett-Jones, 1979]. Very little is known about the coastal plain surrounding the GoP during the past ~10,000 Cal. years or preceding time intervals. Pollen records from Lynch’s Crater, northeastern Queensland (~1000 km SW) indicate dry conditions and Eucalyptus woodlands dominated during 26,000–8,000 Cal. years B.P. [Kershaw, 1978, 1988].

[s7] We tentatively suggest that the C3 signal preserved in MV-51 sediments is indicative of wet tropical vegetation covering the PNG coastal plain and mangroves rimming the coast similar to today [Harrison and Dodson, 1993; Robertson and Alongi, 1995]. Nevertheless, we are unable to rule out the possibility of C4 grasslands in the vicinity of the GoP without pollen data. Therefore, we recommend a future examination of pollen spectra from GoP sediment cores to further test this interpretation.

[s8] TOC values are slightly higher in MV-54 than they are in MV-51. One explanation for higher TOC values may be that the high accumulation rates found in MV-54 would have limited exposure time to organic-carbon oxidation more than at site MV-51 [Harnett et al., 1998; Hedges et al., 1999]. However, wood-rich organic matter is abundant throughout MV-54 in the >63 μm grain-size fraction and that probably best explains the overall higher TOC values compared to MV-51. MV-54 also contains a higher sand-sized fraction of terrigenous siliciclastic particles (mean 15%) compared to the sand-poor sediments of MV-51 (mean 2%) (Figures 3 and 4). Even without carbon isotope data, it is apparent from microscopic observations of sand-sized particles, and the limited biogenic calcium carbonate, that MV-54 is dominated by vascular plant matter.

5.3. Benthic Foraminifera and Organic-Carbon Fluxes

[s9] Accumulations of fossil benthic foraminifers are byproducts of multiple competing biological and taphonomic factors, but overall their abundance and taxonomic compositions can indicate relative abundance of labile organic carbon at the time of deposition. In general, benthic foraminifera live in the upper few cm of sediment, and their microhabitat selection and maintenance primarily are controlled by food and oxygen, and secondarily by competition for those resources [Jorissen et al., 1995; van der Zwaan et al., 1999]. Shell accumulation results from growth and reproduction within their microhabitat and increases as nutrient supplies increase [Nees and Struck, 1999; van der Zwaan et al., 1999]. Postmortem shell abandonment is followed by mixing via bioturbation, transport, and possible removal by dissolution. Shells that eventually pass into the historical record are a reflection of all of those processes [Loubere, 1989; Loubere et al., 1993].

[s10] In core MV-51, BFAR and assemblages during >32,000 14C years B.P. and 18,400–20,400 Cal. years B.P. indicate substantially different oceanographic conditions compared to modern. The older interval >32,000 14C years B.P. contains the highest BFAR of B. robusta, U. peregrina/hispida, S. bulboides, Bul. aculeata, C. pachyderma, and O. tener. Species such as U. peregrina/hispida, C. pachyderma, and S. bulboides are characteristic of high-productivity surface waters and higher flux rates of organic matter under lower seasonality [Altenbach and Sarnthein, 1989; Minagawa and Minagawa, 1997; Loubere and Farradudin, 1999]. C. pachyderma is an epifaunal taxon (0–1 cm) and adapted to uptake of labile organic matter [Rathburn et al., 1996; Jorissen et al., 1998]. Collectively, these taxa are epifauna to shallow infauna (upper 2 cm), meaning they are adapted to abundant food and oxygen [Jorissen et al., 1995, 1998].

[s11] It is surprising that the lower part of MV-51, with highest MAR and terrestrial organic-matter content, also contains the highest densities of benthic foraminifers. Normally, high sediment accumulation rates dilute benthic foraminiferal concentrations, and productivity taxa are not
known to capitalize on refractory organic carbon. However, the higher accumulation rates and burial of refractory, terrigenous dominated TOC suggests that there was a much greater river influence at the time. We account for these discrepancies by suggesting that enhanced river discharge may have supplied nutrients to increase productivity. This provided labile organic carbon for foraminifera, but the labile carbon was not preserved. Highest BFAR before ~32,000 \(^{14}\)C years B.P. suggests that this was the time of highest marine productivity compared to any other time interval in MV-51.

[72] The benthic foraminiferal assemblage during ~18,400–20,400 Cal. years B.P. indicates a much higher seasonally variable flux of organic carbon to the seabed compared with assemblages from >32,000 \(^{14}\)C years B.P. \textit{B. robusta}, \textit{G. translucens}, \textit{G. subglobosa}, \textit{O. tener}, and \textit{M. barleeanus} comprise the assemblage during this time interval. \textit{G. Subglobosa} is a shallow infaunal (0–2 cm) and opportunistic species that is able to respond rapidly to phytodetrital pulses from the photic zone [Linke and Latze, 1993; Rathburn et al., 1996; Smart and Gooday, 2001]. \textit{G. translucens} is typically a shallow infaunal taxon and is found associated with fresh, labile organic carbon [Jorissen et al., 1998]. In the eastern South Atlantic, an association between \textit{G. translucens} and \textit{Epistominella exigua} is correlated to a higher seasonality of export production resulting from seasons of maximal upwelling [Licari and Mackensen, 2005]. \textit{Melonis barleeanus} is an intermediate infaunal taxon (1–4 cm) adapted to degraded organic matter in areas of high upwelling seasonality [Corliss, 1991; De Stigter et al., 1998; Jorissen et al., 1998; Loubere and Farrududin, 1999].

[73] Oceanographic settings with high seasonality in the flux of organic carbon to the seafloor can be found in many locations from the high latitudes to the tropics. In the northeastern Atlantic Rockall Trough, primary productivity and the flux of phytodetrital material to the seafloor is restricted to warmer, ice-free summer months, and as a result, specific benthic foraminifera rapidly increase in numbers with the arrival of labile material at the seafloor [Gooday and Hughes, 2002]. In the Gulf of Guinea, coastal upwelling is perennial, but it also has strong seasonal upwelling events during May–September and December, which are in part also related to maximal discharge from the Congo River [Licari and Mackensen, 2005]. Unlike the Gulf of Guinea, the GoP lacks a western boundary current to promote upwelling, therefore we cannot invoke an upwelling seasonality in the past. However, differences between monsoon and trade wind conditions are seasonal and have significant effects on the ocean.

[74] Enhanced seasonality in the GoP may have occurred in the past by increasing the contrast between the present-day monsoonal climate so that the southeast trades were stronger and the northwest trades were weaker. Marine productivity, which is dependent on nutrient availability in the surface waters, would have increased when ocean currents were more invigorated from stronger southeast blowing trades, and then reduced during calmer surface waters from lower energy northwest trades. This type of seasonal contrast has been demonstrated in the present-day Sulu Sea where primary productivity is highest when wind speeds and upper ocean nutrient mixing are maximal during the East Asian winter monsoon (January–March) [de Garidel-Thoron et al., 2001].

[75] A third interval at 6.8 mbsf is inhabited by \textit{B. robusta}, \textit{U. peregrina/hispida}, and \textit{C. pachyderma} (Figure 8). This assemblage is indicative of high productivity, but it is a very short-lived event and interrupts an interval of otherwise low BFAR. We suggest that this sample may be a recovery fauna after deposition of ash at 6.9 mbsf. The presence of epifaunal species such as \textit{C. pachyderma} is consistent with post-ashfall recovery faunas [Hess et al., 2001].

[76] Intervening lower BFAR intervals experienced low organic-carbon flux. \textit{B. robusta} is consistently abundant in MV-51 and probably is a generalist species, though little is known about its ecology (Figure 8, unshaded areas). Dissolution may be one cause for the low numbers of foraminifera in MV-51; however, the pattern is similar to TOC MAR. Additionally, there is no correlation between the percent calcium-carbonate content and benthic foraminiferal densities (\(r^2 = −0.02; p = 0.59; n = 32\)) as one would expect if dissolution controlled the carbonate content of sediments.

5.4. Mass Fluxes, Sea Level, and Sediment Sources

[77] Two periods of high sediment and TOC accumulation are recorded in MV-54 and MV-51 sediments. The first period (32,000–33,000 \(^{14}\)C years B.P., late regression) corresponds to a time when sea level fall slowed and an extensive midshelf clinoform was aggrading on the midshelf [Slingerland et al., 2008]. High accumulation rates are also observed at other locations in the GoP during this time and may have contributed to mass transport deposits in the Pandora and Moresby Troughs [Francis et al., 2008]. For example, sediment cores from the northern Ashmore Trough and southern Pandora Trough are characterized by turbidites during late regression [Jorry et al., 2008] and a similar “regressive package,” characterized by high terrigenous sediment flux, but deposited during MIS stage 4, is also observed on the slopes of the Ashmore Trough [Francis et al., 2006; Dickens et al., 2006]. The source for the Ashmore Trough regressive unit may have been sediment resuspended by waves and tides from the prograding midshelf clinoform [Dickens et al., 2006] when sea level was ~60 m lower [Chappell and Shackleton, 1986; Lambeck and Chappell, 2001]. However, MV-51 rests on a bathymetric high (not strongly influenced by turbidity currents), and so sediment must have been delivered through the water column via nepheloid-layer advection or similar means.

[78] The MV-51 MS signature and chronology suggest a source shift from sediments with high concentrations (early) to lower concentrations (later) of iron-rich paramagnetic minerals around 32,000 \(^{14}\)C years B.P. Additionally, benthic foraminiferal species indicate oxygenated bottom water and greater organic-carbon flux before that time than after. Organic geochemical proxies suggest greater terrestrial input before 32,000 \(^{14}\)C years B.P. than after. These observations imply a fluvial source for these sediments before 32,000 \(^{14}\)C years B.P. possibly originating from the Papuan Peninsula (Figure 11, fluvial source 3).

[79] Sediment delivery was very high during ~15,000–18,000 years B.P. at MV-54 and likely corresponds to a stillstand time after the first major transgression (10–15 m) associated with a meltwater pulse at 19,000 years B.P. [Clark et al., 2004]. During this time, sea level was ~110 m lower and likely permitted the Fly, Turama, Kikori, and Purari rivers to empty much closer to the shelf edge.
6. Conclusions

[S1] Sediment cores examined in this study span the time interval from ~15,000 to 33,000 years B.P. and provide a unique window into depositional sites at MV-54 and MV-51 in the northeastern GoP. Geochemical and paleontological evidence indicates different oceanographic conditions in the GoP > 32,000 14C years B.P. First, higher TOC MAR and total BFAR prior to 32,000 14C years B.P. suggest that greater amounts of organic-carbon were being delivered to the seabed. Increases in the densities of productivity taxa (U. peregrina/hispida-C. pachyderma-S. bulloides) suggest a significant fraction of this organic carbon was in labile form. However, low hydrogen index values may suggest that if any labile compounds were originally present, they were mostly oxidized and not buried. During this time, increased percentages of terrigenous organic carbon over marine organic carbon suggest an abundant input of C3 and possibly other vascular plant matter. A major clastic sediment source change at ~32,000 years B.P. is suggested by MS. Because properties of sediments in these cores older than 32,000 14C years B.P. are much different from modern sediments, we suggest they were derived from a different source, possibly a closer river draining the Papuan Peninsula.

[S2] The time from 15,000 to 20,400 years B.P. was a period of high sediment accumulation rates. MV-54 experienced the greatest accumulation rates during late transgression (~15,000–18,000 years B.P.) when rivers were situated on the exposed shelf and delivered sediments directly to the slope, although we do not know what accumulation rates were prior to 18,000 years B.P. at this site. Stable-carbon isotopes from MV-51 point to enhanced delivery of terrestrial organic carbon, while the G. translucens, C. globulosa, and M. barleeanus benthic foraminiferal assemblage indicates greater seasonality of organic-carbon flux during this time. The significantly lower accumulation rates at MV-54 and MV-51 during late transgression to Holocene time are evidence of inland and coastal sediment storage [Harris et al., 1996] and possible isolation of sediment supply from the GoP slope by a shelf-edge coral reef complex upslope from MV-51 [Droxler et al., 2006].

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References
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Milliman, J. D., and J. P. M. Syvitski (1992), Geomorphic/ectoctic control of sediment discharge to the ocean: The importance of small mountainous rivers, J. Geol., 100, 525–544.


Milliman, J. D., N. W. Driscoll, R. Slingerland, J. Babcock, and J. P. Walsh (2004), Isotopic stage 3 deposition and stage 2 erosion of a clinoform in northeastern Papua, Mar. Geol., 212, 47–60.


Wagner, T., M. Zabel, L. Dupont, J. Holtvoeth, and C. J. Schubert (2004), Terrigenous signals in sediments of the low latitude Atlantic - Indications to environmental variations during the late Quaternary, Part 1: Organic-


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