indicating that the standard view of clinothems as two-dimensional prograding bodies is not applicable here. The polygenetic nature of this Holocene clinothem is also indicated by the mineralogy of its surficial sediments. Illite/smectite ratios in the clay fraction (Figure 14) and quartz/feldspar ratios in the 62–90 μm fraction (not shown) can be interpreted to reflect a mixing of subarkosic sands and illite-dominated muds delivered by the Fly River from the south with immature arkosic sands and smectite-dominated muds.

Figure 6. Seismic fence diagram of study area looking southwest along the Holocene clinoform face (see Figure 2 for location). Three stratal units are separated by two surfaces of erosion and lap. Lowest and most regionally extensive surface ($S_1$) separates an older, more acoustically laminated, obliquely prograding, and in places, folded unit (Yellow unit) from two younger units that are more acoustically transparent, unfolded, and up-building (Orange and Red). Surface $S_2$ separates a localized acoustically bright unit (Orange) from an acoustically transparent regional drape (Red). (See Johnstone et al., submitted manuscript, 2008, for a more detailed description).

Figure 7. Enlargement of Figure 6 showing the transition from oblique progradation near depocenter to sigmoidal progradation on southern edge of the study area. Also note folds in Yellow unit.
delivered by the steeper rivers tapping the volcanogenic terrain to the north (Table 1).

5. Interpretation

5.1. Geologic History of the Gulf of Papua Shelf

[24] The upper approximately 100 m of the GoP shelf consists of two stacked clinothems: an older, partially eroded clinothem forming the floor of the middle shelf, and a Holocene clinothem forming the inner shelf. On the basis of the stratal geometry of the older clinothem and its present elevation, we deduce that it prograded two-thirds of the way across the preexisting shelf in response to a relative sea level fall during Stage 4 (Figure 15), when the rate of sediment supply filled the available accommodation. We infer that these oblique clinoforms were deposited during Stage 4 (C24.5m m C01m) because we observe no aggradation during progradation (e.g., Figures 4a and 4c); this requires a rate of eustatic sea level fall that is faster than subsidence. This inference is consistent with a major siliciclastic pulse observed on the slope between 60,000–100,000 years BP (Mallarino et al., 2004).

[25] Above the oblique clinoforms lie parallel reflectors that represent topsets (e.g., Figure 4). Using sequence stratigraphic principles we infer that these topsets formed when the rate of sediment supply equaled the creation of new accommodation (e.g., Figure 1) during Isotope Stage 3 prior to the large sea level fall associated with the LGM (Figure 15). This age is consistent with data from several cores acquired by Harris et al. [1996] which recovered Pleistocene sediments from the top of the parallel sequence where exposed at or near the seafloor. Core 37PC11 from the top of an erosional “mesa” yielded a radiocarbon date of 33,850 years BP (Harris et al., 1996) and other cores from the “mesa” features that sampled the parallel, concordant unit (e.g., Fr-93–1PC1, 30PC8) recovered Pleistocene deltaic facies and cohesive reddish brown clays. On the basis of these observations, Harris et al. [1996] suggested that the pretransgressive deposits were deposited 50,000–20,000 years BP, consistent with our interpretation. If correct, such a scenario suggests that tectonic subsidence in the region was outpacing the slow eustatic fall during Stage 3 (C(24.5m m 1m) or 10 m/10,000 years) to create the available accommodation (Figure 15).

[26] This magnitude of differential tectonic subsidence (~1 mm a⁻¹ or 10 m/10,000 years) is problematic. Dated coral material acquired from fringing reefs along the Ashmore Trough at 125 m water depth (Core MV-73) suggests there has been little to no tectonic subsidence since the LGM in that region (Droxler et al., 2006). However, Core MV-73 was acquired south of the peripheral bulge for the GoP foreland basin (Figure 2) and thus might not reflect tectonic subsidence within the foreland. Furthermore, the tectonic subsidence rate demanded by the parallel, concordant topsets observed in the older clinoform is consistent with differential tectonic subsidence rates determined from exploratory wells in the region (e.g., Borabi-1 and Pasca C1; Davies et al., 1989). In addition, the transgressive surface formed as a consequence of the rapid
sea level rise following the LGM exhibits a regional dip toward the northwest. If the transgressive surface had minimal relief when it was formed by wave base erosion, then a dip of this magnitude requires a differential tectonic subsidence rate of ~1 mm a⁻¹. Finally, it is interesting to note that high accumulation rates on the slope between 60,000–100,000 years BP are succeeded by lower rates from ~30,000–55,000 years BP [Mallarino et al., 2004], consistent with an increase of accommodation on the shelf in response to a relative sea level rise produced by tectonic subsidence.

In the central gulf midshelf region NE of the Fly River mouth, the older clinothem has been dissected by six major valleys and numerous channels (cf. Figure 4). We interpret this erosion to be fluvial in origin and to have occurred during the rapid eustatic fall commencing ~25 ka BP (Stage 2) that lowered sea level to ~125 m. An alternative hypothesis is that they are tidally incised shelf valleys in the manner of Harris et al. [2005], but we consider this unlikely given the dendritic nature of the valley systems and the deep channels observed within the valleys (Figure 4d). As eustatic sea level rose from ~125 to +3 m between 20 and 7 ka BP, the former midshelf river channels within the river valley were partially filled by fluvial and estuarine facies [Milliman et al., 2006]. The transgression was so rapid however, that the valleys themselves remained unfilled except for minimal input from eroding interfluvies. Farther landward near the Fly River

Figure 9. (a) Chirp dip line PNG4L10 near toe of modern clinothem showing location of JPC-43. (b) Core lithic log, attenuated gamma and magnetic susceptibility logs, and radiocarbon ages. Acoustically transparent, very fine silt/clay intervals intercalated with medium thick beds of more reflective very fine sands downlap onto older clinothem. Surface $S_I$ is the light gray horizon just beyond the bottom of the core.
mouth, more of the transgressive systems tract (TST) was preserved \cite{Harris et al., 1996}, but even there it remains thin and discontinuous. In most places the sequence boundary, the transgressive surface, and the maximum flooding surface coalesce (e.g., Figure 4a). When and where a Holocene GoP clinotherm began advancing seaward remains unknown; the turn around point lies landward of our data. The Holocene clinotherm had prograded to within 15 km of its present position by about 4.8 ka BP, although the Yellow unit may be older than 9–9.5 ka BP (i.e., MWP 1C). This early growth produced acoustically high-amplitude reflectors that exhibit oblique progradation at the top of the Yellow unit (Figures 6 and 7), indicating that sediment supply was greater than the creation rate of new accommodation and filled all the available accommodation. Stratal strike profiles show undulations in the clinoform surface caused by the mesas and valleys over which the Holocene clinotherm was prograding (Figure 12). The folds in the Yellow unit in the NE corner of the study area appear to record continued downslope soft sediment deformation during clinotherm growth. The Orange and Red units downlap onto the $S_1$ surface and this requires the generation of new accommodation after the deposition of the Yellow unit. Given the uncertainty in the age of $S_1$, two possible scenarios might explain how the new accommodation was formed. In scenario 1, the oblique clinoforms of the Yellow unit were formed during a eustatic stillstand approximately 9–9.5 ka BP. A subsequent rapid rise in sea level during MWP 1C (dated in the Great Barrier Reef and in southern Asia at 9.1–9.6 ka BP \cite{Larcombe and Carter, 1998; Liu et al., 2004b}) produced the $S_1$ surface and the accommodation in which the finer-grained Orange and Red units came to downlap the Yellow unit (Figures 5 and 6). Alternatively, if the $^{14}$C dates from the flow-in of core JPC-40 (Figure 8) come from both above and below $S_1$, then the age of $S_1$ lies between 2.41 and 5.2 ka BP. Because there are no known rapid rises of sea level during this interval, $S_1$ must be due to a rapid decline in the rate of sediment supply at a constant subsidence rate. In scenario 2, the oblique clinoforms are formed during a period of high sediment supply and then the sediment supply is shut off and long-term subsidence accounts for the generation of the new accommodation and formation of the downlap surface. We can find no evidence from other studies to decide between the two scenarios. Longer cores are required to test these two alternatives. 

Figure 10. Shaded structure contour map of $S_1$ surface draped on modern smoothed bathymetric contours. See text for details.

[28] Clinothem growth resumed after the $S_1$ event as localized sedimentation of the Orange unit in the southern half of the study area. It fills local accommodation, offlapping both to the north and south. By approximately 1.6 ka BP ago, sediments of the Red unit began to blanket the whole study area. These latest sediments consist of an acoustically transparent facies that forms an on- and downlapping toe of the wedge. The thickness of the Red unit

Figure 11. Isopach map of the interval between the $S_1$ surface and the seafloor, draped over shaded bathymetry. Note that on the upper clinoform face the Orange and Red units are thickest over promontories in underlying landscape, but at the clinoform toe they are thickest in reentrants. See text for details.
exhibits less variability than the Orange (Figures 6, 7, and 12). This transparent facies appears to be sediment winnowed from upslope strata, suggesting that at present, clinothem progradation may be moribund. The thickness variability observed in the isopachs as well as the high-amplitude reflectors observed in the valleys suggest that the dominant transport is strongly oblique to the isobaths and both northeast and southwest directed.

Mineralogy of the uppermost beds of the Holocene clinothem also indicate significant bidirectional alongshore sediment transport (Figure 14). The Fly River sediment consists of mature illite clay and quartz-rich sands, consistent with its source rock composition and long distance of transport. Sediments of the northern rivers such as the Kikori and Purari consist of immature smectite clays and lithic sands with a strong volcanic component. Sediments of the modern shelf represent a mixture of these two types. This mixed assemblage can only be explained by at least some component of southerly transport along much of the upper tread with a convergence zone just north of the Fly River where sediment is shunted seaward off the shelf.

5.2. Modes of Clinothem Growth

The Holocene GoP clinothem has not formed by simple parallel progradation of topsets over irregular bathymetry. Rather, the clinothem is composed of discrete lobes that formed at separate times. Cross-cutting relationships among the shingled lobes indicate that overall development of the clinothem complex has progressed from the center of the study area outward. These lobes are the fundamental architectural elements of the modern clinothem’s stratigraphy. They consist of both onlapping and downlapping reflectors, suggesting that depocenters originated and accreted on the clinothem face, expanding updip and downdip as well as laterally. The pattern of paired onlapping and downlapping is particularly evident in the thickest (central and southwestern) parts of each clinothem segment. This architecture requires a sediment transport process such as sedimentation from suspension or failed

Figure 12. Chirp seismic profile across central lobe illustrating the differential thickness of the upper two units (Orange and Red) being thickest on the promontories and thinnest in the valleys. Note high-amplitude reflectors within the valleys that correlate with sands.

Figure 13. Reflector dips of strata above and below $S_1$, showing that dip directions are roughly similar indicating self-similar progradation.
gravity flows that produces initial vertical accretion on the clinoform face and then extension of the locus of deposition both upslope and downslope. The volume of sediment per unit distance along strike deposited since $S_1$ time can be estimated from the isopach map (Figure 11) to be 420,000 m$^2$. For a mean vertical thickness above $S_1$ of 16.2 m and a mean modern clinoform slope of 0.14$^\circ$, the mean distance of recent clinoform progradation equals 6.6 km, and the mean clinothem vertical sedimentation rate is between 1 and 5 mm a$^{-1}$. Because the analysis area does not extend completely to the updip end of the clinothem, these values should be considered minimal. This vertical sedimentation rate is smaller by an order of magnitude than short-term sedimentation rates of 1–4 cm a$^{-1}$ reported by Harris et al. [1996] and Walsh et al. [2004], based on $^{210}$Pb and short-core $^{14}$C dates, but are consistent with rates of 3 to 4 mm a$^{-1}$ computed using about twenty AMS $^{14}$C dates from our cores, which range from a few hundred years to a few thousand years old (see the Marine Geoscience Data System at http://www.marine-geo.org/link/entry.php?id = VANC23MV for core descriptions and locations). Although it is well known that sedimentation rates calculated using dates spanning greater stratigraphic intervals are always lower [Gardner et al., 1987], this effect cannot explain the magnitude of the discrepancy here over such a relatively young time range of hundreds to a few thousands of years.

[31] Our data indicate that during the formation of the Holocene clinothem, the northern rivers in the Gulf of Papua were supplying much sediment to the midshelf region and filling the available accommodation. Near the Fly River in the southern gulf, sediment input to the region was less than the available accommodation, resulting in the formation of sigmoidal clinoforms.

[32] Can changes in sediment transport patterns explain the $S_1$ and $S_2$ surfaces and the shifting loci of deposition? Computed annual circulation of the GoP in response to

![Figure 14](image-url)  
**Figure 14.** Illite/smectite ratios of the clay size fraction in surface sediments of the Gulf of Papua. Ratio increases systematically from northeast to southwest, indicating smectite-dominated rivers to the northeast contribute a large fraction relative to the illite-dominated Fly River to southwest.

![Figure 15](image-url)  
**Figure 15.** Inferred formation of the Gulf of Papua clinothems. Older, partially eroded clinothem consists of approximately 30–40 m of generally subparallel reflectors mantling a set of obliquely prograding reflectors. Deflection of the transgressive surface from horizontal yields a post-LGM differential subsidence of about 1 mm a$^{-1}$ from the peripheral bulge in the southwest to near the basin depocenter in the northeast. On the basis of the Lambeck and Chappell [2001] and Chappell and Shackleton [1986] eustatic sea level curve, eustatic sea level fell from $-40$ to $-65$ m between 45 and 25 ka BP. Older clinothem topsets are thought to have accumulated at the end of Stage 3 about 30 ka BP as accommodation space was created by a relative slowing of sea level fall.
trade wind and monsoon conditions shows that the flow fields are significantly different [Slingerland et al., 2008]. Possibly the surfaces and lobes were created by century to millennial-scale changes in the wind fields. This can be a sufficient explanation only if the change in circulation changed the amount of accommodation, because that is the only way to explain oblique offlap changing to aggradation. The most obvious way to do this is to change wave base significantly.

6. Conclusions

The upper 100 m of the Gulf of Papua shelf are composed of two stacked clinotomes: an older deeply eroded clinosome forming the middle and outer shelf, and a superjacent younger clinotome extending from the coast offshore forming the inner shelf. The older, partially eroded clinotome was built during Stages 4 and 3, eroded into a series of northwest-southeast trending paleovalleys and mesas during Stage 2, and partially covered by a younger clinotome during Stage 1. The younger clinotome downlaps onto the erosional surface etched into topsets of the older clinotome. It consists of three major stratigraphic units that are separated by two surfaces of erosion or bypass or correlative surfaces of down/up/on/off-lap. The ages and origins of these surfaces are not well known, however the formation of S1 and its mantle of downlapping strata requires the formation of new accommodation. This stratal geometry cannot be formed by simple lobe switching. We present two scenarios to explain the formation of the new accommodation: changes in the rate of eustatic sea level rise, or changes in the rate of sediment supply with long-term tectonic subsidence.

The upper 100 m of the Gulf of Papua shelf are composed of two stacked clinotomes: an older deeply eroded clinotome forming the middle and outer shelf, and a superjacent younger clinotome extending from the coast offshore forming the inner shelf. The older, partially eroded clinotome was built during Stages 4 and 3, eroded into a series of northwest-southeast trending paleovalleys and mesas during Stage 2, and partially covered by a younger clinotome during Stage 1. The younger clinotome downlaps onto the erosional surface etched into topsets of the older clinotome. It consists of three major stratigraphic units that are separated by two surfaces of erosion or bypass or correlative surfaces of down/up/on/off-lap. The ages and origins of these surfaces are not well known, however the formation of S1 and its mantle of downlapping strata requires the formation of new accommodation. This stratal geometry cannot be formed by simple lobe switching. We present two scenarios to explain the formation of the new accommodation: changes in the rate of eustatic sea level rise, or changes in the rate of sediment supply with long-term tectonic subsidence.

References


Berne, S., M. Rabineau, J. A. Flores, and F. J. Sierra (2004), The impact of Quaternary global changes on strata formation, Oceanography, 17(4), 92–103.


Harris, P. T. (1990), Sedimentation at the junction of the Fly River and the northern Great Barrier Reef, in Torres Strait Baseline Study Conference, edited by D. Lawrence and T. Cansfield-Smith, pp. 59–85, Queensland, Australia.


Harris, P. T., A. Heap, V. Passlow, M. Hughes, J. Daniell, M. Hemer, and O. Anderson (2005), Tidally incised valleys on tropical carbonate