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Did the last sea level lowstand always lead to cross-shelf valley formation and source-to-sink sediment flux?

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[1] It is commonly believed that the efficiency of sediment flux from continents to oceans is maximized during relative sea level (RSL) lowstands by means of cross-shelf valleys that are directly connected to the continental slope and deep marine environment. While such conditions have been documented for the last RSL lowstand along several continental margins, there is increasing evidence that radically different conditions persisted during the Last Glacial Maximum (LGM) elsewhere, with lowstand shorelines that remained on the continental shelf. Here we analyze the relationship between the LGM (21 ka) shoreline and the shelf edge for the Gulf of Mexico off the United States and the Bay of Biscay off France. A geophysical model is used to compute shoreline positions corrected for isostatic movements, and the shelf edge position is quantified by means of curvature. The conditions in the two study areas differed markedly: throughout the Gulf of Mexico study area, LGM sea level dropped to a point commonly ~40 m below the shelf edge, consistent with conventional sequence stratigraphic models, while in the Bay of Biscay the modeled LGM shoreline remained well landward of the shelf edge, in places separated by hundreds of kilometers. These observations hint at potentially significant implications for (1) the source-to-sink sediment flux from continents to oceans and its variation in time and space, (2) sequence stratigraphic models that predict deep marine sedimentation as being particularly prominent during RSL lowstands, and (3) the occurrence of paleovalleys and related features on the continental shelf. In addition, our findings raise fundamental questions about the mechanics of shelf edge formation.

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1. Introduction

[2] There is an increasing desire to understand sediment dispersal systems holistically, from source to sink. As a prelude to current thinking, the past few decades have witnessed the proliferation of sequence stratigraphic models (reviewed, for example, by *Posamentier and Allen* [1999]), constituting a paradigm shift in sedimentary geology. Sequence stratigraphy seeks to understand large-scale sedimentation patterns in the broad zone along continental margins, largely as a function of relative sea level (RSL) change, and, increasingly, hinterland sediment supply [e.g., *Jervey*, 1988; *Posamentier et al.*, 1988; *Galloway*, 1989; *Miall*, 1991; *Schlager*, 1993].

[3] Building on pioneering early work such as that by *Fisk and McFarlan* [1955], a key feature of sequence stratigraphic models is the notion that deep marine sedimentation, notably by means of the growth of submarine fans, is amplified during RSL lowstands [e.g., *Vail et al.*, 1977; *Shanmugam and Moiola*, 1982; *Posamentier and Vail*, 1988; *Emery and Myers*, 1996; *Miall*, 1997; *Posamentier and Allen*, 1999], although it may sometimes continue well into the subsequent RSL rise [*Kolla and Perlmutter*, 1993]. The mechanism behind this amplification is thought to be straightforward: when the shoreline is located near the shelf edge, sediment released at river mouths can easily find its way into the deep ocean across the steep continental slope [e.g., *Mulder and Syvitski*, 1996]. On the other hand, examples have been presented where sand-prone deposition is limited to the continental slope, even with sea level below the shelf edge [*Plink-Björklund and Steel*, 2002]. Nevertheless, a genetic link between shelf edge deltas (reflecting RSL lowstand) and deepwater sedimentation is considered to be very likely according to a recent synthesis [*Porębski and Steel*,

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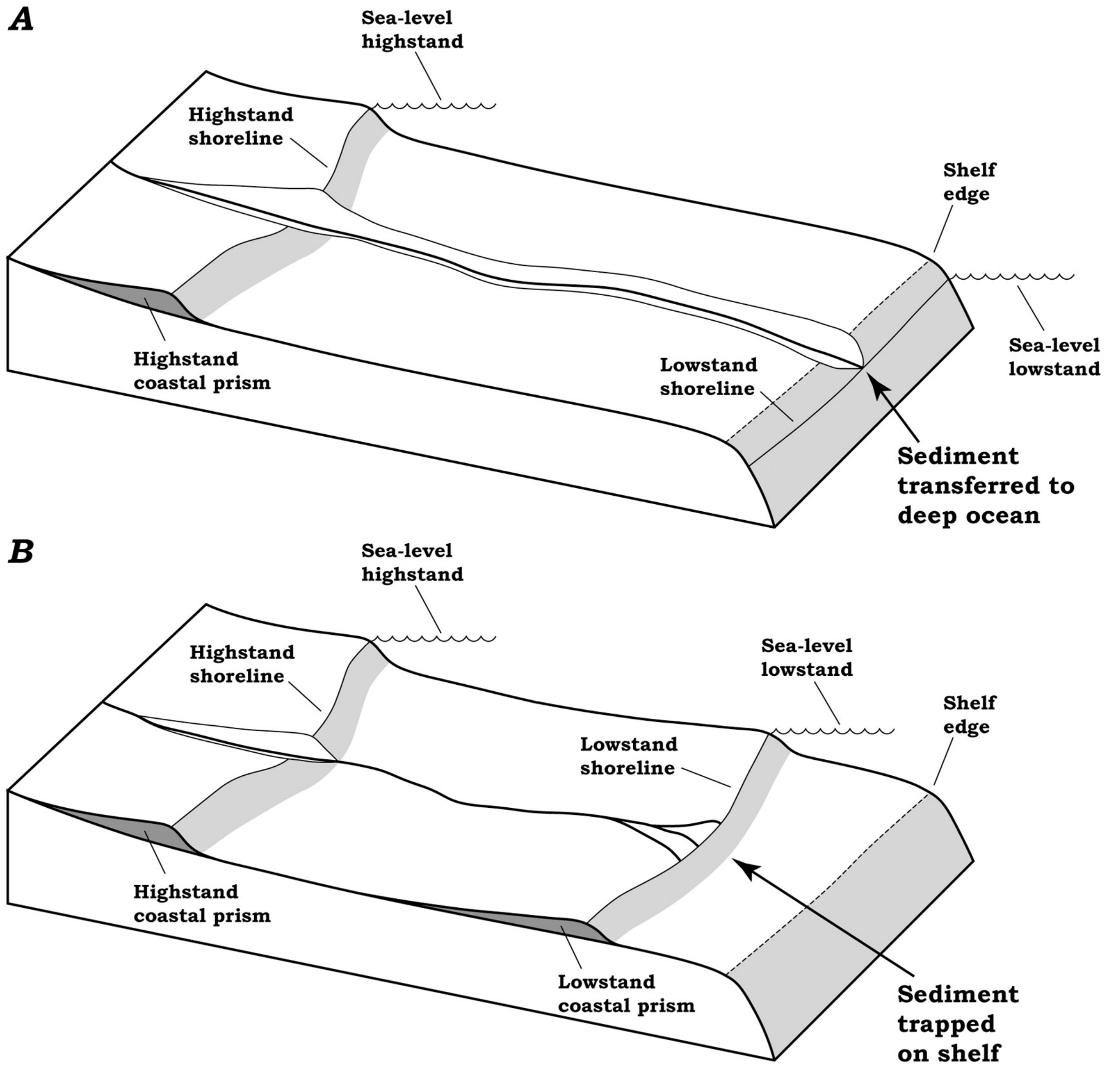

Figure 1. Contrasting scenarios of relative sea level fall and fluvial response [after *Posamentier and Allen*, 1999]. (a) Lowstand shoreline below the shelf edge triggers headward erosion and the development of a cross-shelf valley that provides a direct connection between the hinterland and the deep marine realm. (b) Lowstand shoreline that remains on the shelf limits fluvial incision and valley formation to the highstand coastal prism and traps the majority of terrigenous sediment updip of the shelf edge.

2003]. Finally, some studies have advocated that submarine fans may accrete during RSL highstands as well [e.g., *Weber et al.*, 1997; *Burgess and Hovius*, 1998], but others have argued that such conditions may be exceptions rather than the rule [e.g., *Gibbs*, 1981; *Muto and Steel*, 2002].

[4] While first-generation sequence stratigraphic models suggested that RSL fall at the “depositional shoreline break” would automatically lead to valley cutting [*Posamentier and Vail*, 1988], subsequent studies have stressed that this is only likely to happen if shelf gradients exceed coastal plain gradients [e.g., *Miall*, 1991; *Posamentier et al.*, 1992; *Schumm*, 1993]. It is now increasingly recognized that the common presence of a relatively steep shoreface makes it likely that even modest RSL fall can induce incision of highstand coastal prisms [cf. *Blum and Price*, 1998; *Talling*, 1998; *Törnqvist et al.*, 2000, 2003; *Posamentier*, 2001], unless sediment supply is sufficient to neutralize the exposure of the steep shoreface [*Leeder and Stewart*, 1996].

[5] A solid understanding of how, when, and where valleys are excavated in response to RSL fall remains a topic of considerable interest, for several reasons. The basal surface of paleovalley fills is commonly taken to define the sequence boundary [e.g., *Zaitlin et al.*, 1994], the most commonly used bounding surface for the identification of allostratigraphic units. Since sedimentary successions are often interpreted in terms of RSL cycles, it is critical to understand the relationship between RSL change and valley formation. Numerous recent studies [e.g., *Schumm*, 1993; *Blum*, 1994; *Koss et al.*, 1994; *Blum and Price*, 1998; *Talling*, 1998; *Blum and Törnqvist*, 2000; *Heller et al.*, 2001; *Posamentier*, 2001; *Van Heijst and Postma*, 2001; *Meijer*, 2002; *Törnqvist et al.*, 2003; *Wellner and Bartek*, 2003; *Fagherazzi et al.*, 2004; *Wallinga et al.*, 2004; *Swenson*, 2005] have documented the complexity of this relationship, both in a spatial and a temporal context. In particular, the question of whether RSL drops below the shelf edge is a critical one (Figure 1),

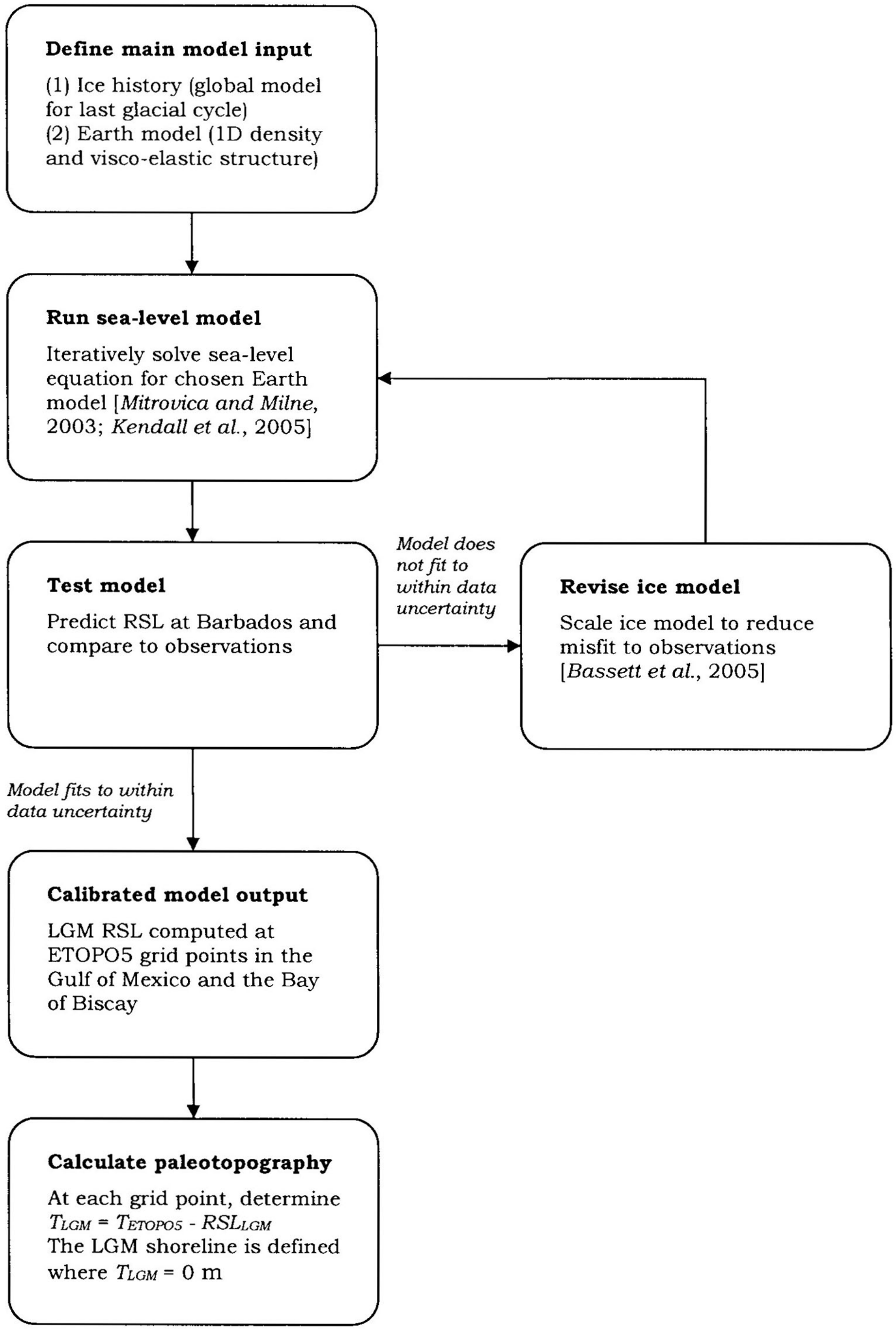

Figure 2. Primary steps in applying the geophysical model to obtain the paleotopography and paleobathymetry during the Last Glacial Maximum. T_{LGM} = predicted LGM topography/bathymetry, T_{ETOPO5} = present-day topography/bathymetry as defined by the ETOPO5 data set, RSL_{LGM} = predicted LGM relative sea level field. Further details on the modeling are provided in the text.

a problem we will explore in more detail in this paper. These issues also have a strong bearing on the ongoing discussion about the positioning and timing of sequence boundaries, with much of the debate centering on the question whether the sequence boundary should underlie or overlie the falling stage (or forced regressive) systems tract [e.g., *Hunt and Tucker*, 1992; *Kolla et al.*, 1995; *Plint and Nummedal*, 2000; *Posamentier and Morris*, 2000; *Törnqvist et al.*, 2003]. Finally, and perhaps most importantly, the establishment of a fully developed cross-shelf valley during RSL lowstand promotes the direct connection between the hinterland sediment delivery system and deeper marine depocenters.

[6] The dramatic nature of Quaternary sea level change is driven by orbitally paced glacio-eustatic cyclicity. RSL low-

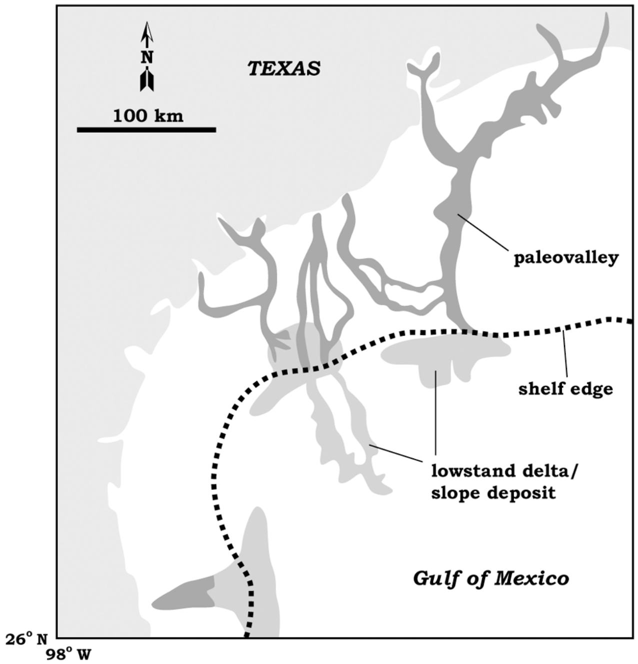

Figure 3. Major cross-shelf valleys and lowstand deltas in the northwestern Gulf of Mexico, 22–16 ka, as mapped by means of high-resolution seismic data [after *Anderson et al.*, 2004].

stands coincide with glacial maxima, like the Last Glacial Maximum (LGM) ~20 kyr ago. Records of RSL change from the Caribbean [*Fairbanks*, 1989], Southeast Asia [*Hanebuth et al.*, 2000], and Australia [*Yokoyama et al.*, 2000] estimate this lowstand at about 120 m below present sea level. The global average depth of the shelf edge is 130 m [*Shepard*, 1973], remarkably similar to the estimates of the last RSL lowstand. Indeed, there is a long-standing and widely held belief that the shelf edge approximately coincides with the lowstand shoreline [e.g., *Vanney and Stanley*, 1983, and references therein]. However, there is substantial variability as to the depth of the modern shelf edge [*Shepard*, 1973], ranging from less than 100 m to values of 200 m or more [*Burgess and Hovius*, 1998]. It is difficult to attribute the departure of modern shelf edge depths from LGM sea level to post-LGM tectonic subsidence or uplift, sedimentation/erosion, or sediment compaction. These lines of evidence alone suggest that the position of the lowstand shoreline relative to the shelf edge may have been highly variable, and some authors [e.g., *Trincardi and Field*, 1991] have pointed out the stratigraphic significance of variable depths of the shelf edge.

[7] Geophysical model studies that compute RSL changes during the last deglaciation demonstrate that during the LGM, the shoreline did not always find itself near the shelf edge, but in some regions attained a position halfway across what is now the continental shelf [*Lambeck*, 1995, 1997]. These findings, supported by recent geological evidence which we will discuss further below, would imply that in some cases the continental and deep marine segments of sediment dispersal systems remained physically disconnected even during RSL lowstand. This has considerable repercussions for the sediment transfer from source to sink and its variability in time and space. Under such conditions, the sources for deep marine sedimentation would be restricted to sediment gravity flows and related mass-wasting processes that originate on the continental slope, and, perhaps, long-run-out hyperpycnal flows that traverse the shelf under conditions of high riverine sediment discharge. Such a scenario is largely at odds with the commonly held view, as reflected by the early sequence stratigraphic models in *Special Publication of Society of Economic Paleontologists and Mineralogists*, volume 42 [*Wilgus et al.*, 1988] that continue to permeate current thinking, of a connection between source (hinterland) and sink (deep ocean) during RSL lowstand. The objective of our contribution is to show the striking contrast between two passive margins in terms of the position of the LGM shoreline relative to the shelf edge, to quantify this relationship, and to discuss its implications for source-to-sink sediment flux, sequence stratigraphic models, and shelf edge evolution.

2. Geophysical Modeling of Lowstand Shorelines

[8] Our analysis is underpinned by a geophysical model that calculates shoreline positions corrected for isostatic crustal movements. Mass transfer between ice sheets and the global oceans, and the associated deformation of the solid earth, constitute the main processes driving Quaternary RSL change. This is particularly true during the last deglaciation (approximately 20 to 7 ka) when ~70% of the global ice melted. The sheer magnitude of this mass transfer produced a dramatic isostatic response that continues today and remains an important contributor to present-day RSL change both close to and far from the centers of glaciation.

[9] Geophysical models of glaciation-induced RSL change were first introduced in the 1970s [e.g., *Farrell and Clark*, 1976; *Peltier and Andrews*, 1976; *Clark et al.*, 1978]. These models comprise three key components: (1) an ice history to drive the system, (2) an algorithm to solve the sea level equation [*Farrell and Clark*, 1976; *Milne et al.*, 1999] and thus ensure that the meltwater redistribution is computed accurately, and (3) a model of earth rheology to calculate the isostatic response to the ice and ocean load histories. These models have undergone considerable improvement (reviewed in more detail by *Milne* [2002]), especially in the past decade, to include such effects as time-dependent shoreline migration [e.g., *Johnston*, 1993], the influence of glaciation-induced perturbations of earth rotation [e.g., *Milne and Mitrovica*, 1998], and a consistent treatment of sea level change in regions characterized by marine-based ice [*Milne et al.*, 1999]. The sea level algorithm used in this study includes all of these advances and is based on the most recent version of the theory [*Mitrovica and Milne*, 2003; *Kendall et al.*, 2005].

[10] The large body of work that has been carried out to improve this type of geophysical models has dramatically increased their predictive power. Extensive data-model comparisons have been performed for various regions and have mostly validated these models with errors typically <10 m [e.g., *Lambeck et al.*, 1998; *Milne et al.*, 2002; *Shennan et al.*, 2002].

[11] The ice model adopted in this study is derived from the global deglaciation model ICE-3G [*Tushingham and Peltier*, 1991] and has been tuned to fit the Barbados RSL

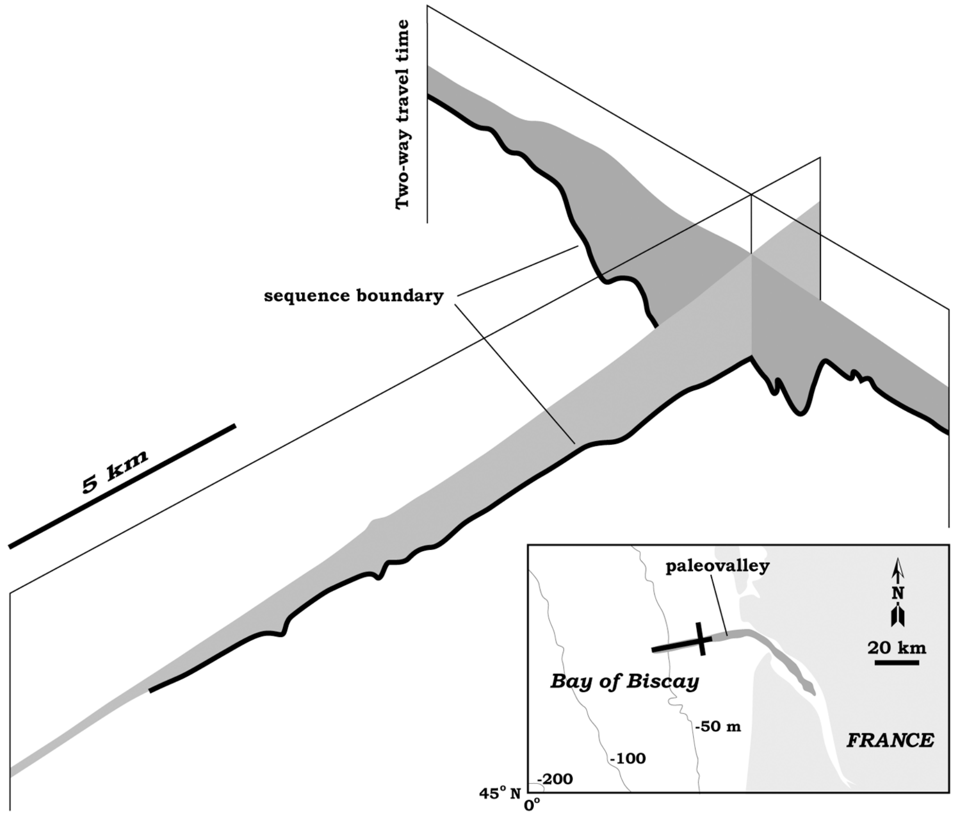

Figure 4. Paleovalley of the Gironde in the Bay of Biscay as revealed by high-resolution seismic data. The cross in the inset map shows the location of the strike and dip sections, with the latter showing that the valley pinches out at a depth of only ~50 m below present sea level [after *Lericolais et al.*, 2001].

record [cf. *Bassett et al.*, 2005]. The adopted earth viscosity model has a 96-km-thick elastic lithosphere, a viscosity in the upper mantle (base of lithosphere to 670 km seismic discontinuity) of 5×10^{20} Pa s, and a viscosity in the lower mantle (base of upper mantle to core-mantle boundary) of 10^{22} Pa s. This viscosity model is broadly compatible with a number of recent inferences [e.g., *Mitrovica*, 1996; *Lambeck et al.*, 1998; *Kaufmann and Lambeck*, 2002; *Mitrovica and Forte*, 2004]. Figure 2 synthesizes the main steps in the modeling procedure.

3. Model Data Analysis

[12] For this study, we selected two ~1000-km-long continental margins: the Gulf of Mexico offshore of the United States and the Bay of Biscay offshore of France. We focused on these two areas because previous work has suggested that they might constitute contrasting settings in terms of the position of the LGM shoreline relative to the shelf edge. A combination of cross-shelf river valleys and shelf margin deltas characterized the northern Gulf of Mexico during the last RSL lowstand [e.g., *Suter and Berryhill*, 1985; *Anderson et al.*, 2004], suggesting that the lowstand shoreline dropped to a point at or below the shelf edge (Figure 3). In contrast, model calculations by *Lambeck* [1995, 1997] indicate that offshore of the British Isles and France the LGM shoreline was located well landward of the shelf edge, in many cases by as much as hundreds of kilometers. This is consistent with high-resolution seismic data collected offshore of the Gironde estuary [*Lericolais et al.*, 2001], providing a striking example of a lowstand paleovalley that pinches out on the proximal part of the shelf (Figure 4). While paleovalleys have been mapped along the shelf edge off the English Channel, they are believed to be mostly pre-Quaternary in age [*Bourillet et al.*, 2003], and no such features have been reported from this area in association with the LGM.

[13] We calculated the paleotopography and paleobathymetry of our study areas (Figure 5) by subtracting modeled LGM (21 ka) RSL positions from the present-day digital elevation model (DEM) ETOPO5, available from the National Geophysical Data Center. This data set has a spatial resolution of 5 min (~10 km) and a vertical resolution of 1 m [*Smith and Sandwell*, 1997]. Thus the lowstand shoreline is defined as the intersection of LGM sea level with the modern sediment surface, suitably adjusted for isostasy. Since post-LGM transgressive erosion has resulted in extensive obliteration of shoreline features, the morphologic record of lowstand shorelines has usually been removed. To our knowledge, no independent and sufficiently accurate sea level indicators for the 21 ka lowstand are available from the two study areas.

[14] We imported the model output into the Geographic Information System ArcView to analyze the calculated DEMs of the two study areas, using an Albers projection. Specifically, this concerned the quantitative characterization of the shelf edge. In this study, the shelf edge is defined as the first significant downdip increase in slope, following the rationale that a relatively modest slope increase will com-

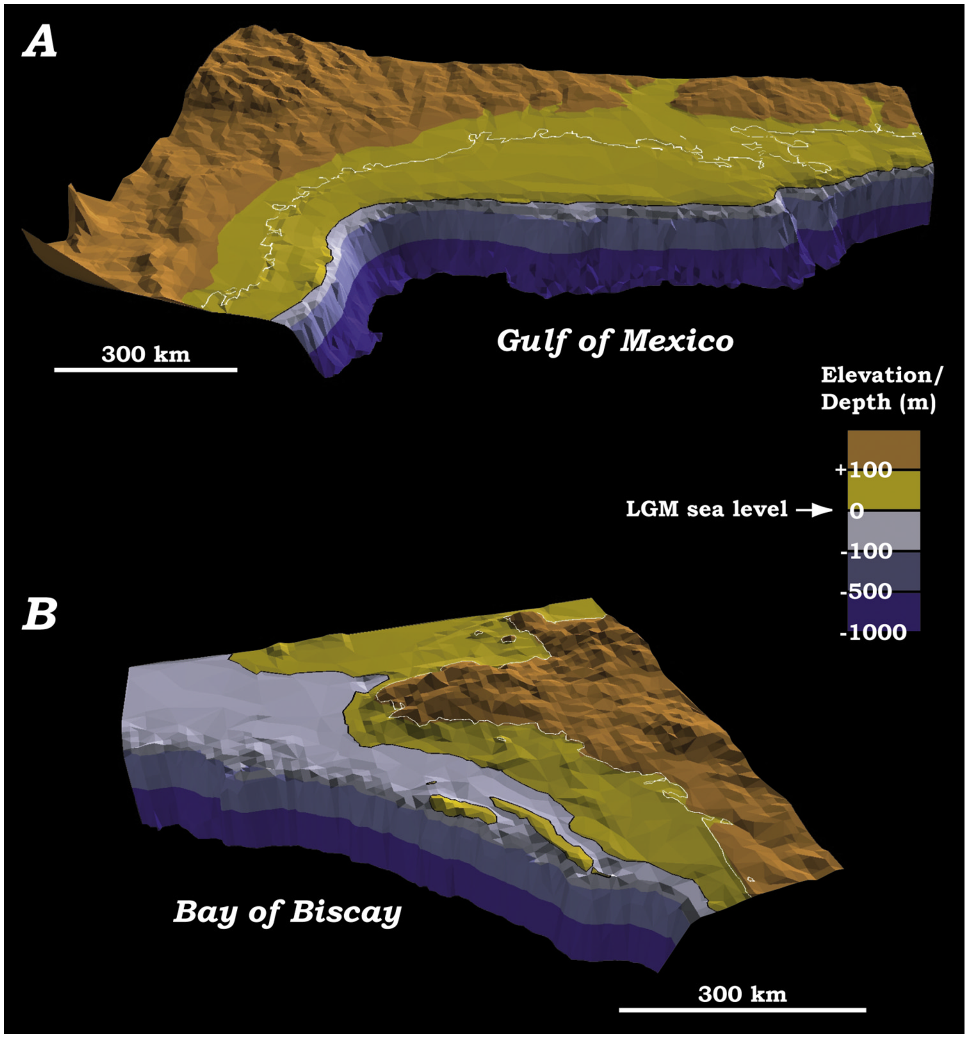

Figure 5. Three-dimensional view of the land/seascape in (a) the northern Gulf of Mexico and (b) the Bay of Biscay during the Last Glacial Maximum (21 ka), as calculated with the geophysical model discussed in the text. The modern shoreline is indicated in white, and the modeled LGM shoreline is in black. Vertical exaggeration is 100X; horizontal scale bars are approximate. The striking feature is the opposite spatial relationship between the lowstand shoreline and the shelf edge between the two areas. Additional 3-D imagery of the study areas, featuring overviews that include the basin floor as well as low-angle perspective views, is available in the auxiliary material.

monly lead to fluvial incision. The first step of our analysis consisted of filtering of the DEM using a 3×3 pixel moving window. Next, the ArcView curvature algorithm was applied and the output was once again smoothed using a 3×3 pixel moving window. The curvature calculation uses a polynomial surface that is fitted to the DEM and provides negative values for convex-up features, as illustrated in the auxiliary material.[1] The following step consisted of contouring the dip component of the curvature (known as "profile curvature" in ArcView). We subsequently selected, for each study area, the curvature contour that provided the best characterization of the shelf edge.

[15] The problem outlined here is highly three-dimensional in nature. Therefore the calculated shelf edge positions were verified using the GeoWall, a stereo projection system that aids the understanding of 3-D spatial relationships. Our strategy was to use the GeoWall to validate the results and to ensure that the selected curvature contours capture the morphologic features of interest. As will be discussed below, this approach was mostly successful, although some problems were encountered in a few instances. The curvature value that best characterizes the shelf edge is an order of magnitude higher in the Bay of Biscay (-7×10^{-5}) compared to the Gulf of Mexico (-6×10^{-6}). This is not surprising, given the wide morphometric variability of continental margins [*Pratson and Haxby*, 1996] and the fact that the Bay of Biscay has been proposed as a type locality of steep passive margins [*O'Grady et al.*, 2000].

4. Results

[16] The results of our analysis confirm the striking differences between the two study areas. Figures 5 and 6 show that

[1]Auxiliary materials are available in the HTML. doi.10.1029/2005JF000425.

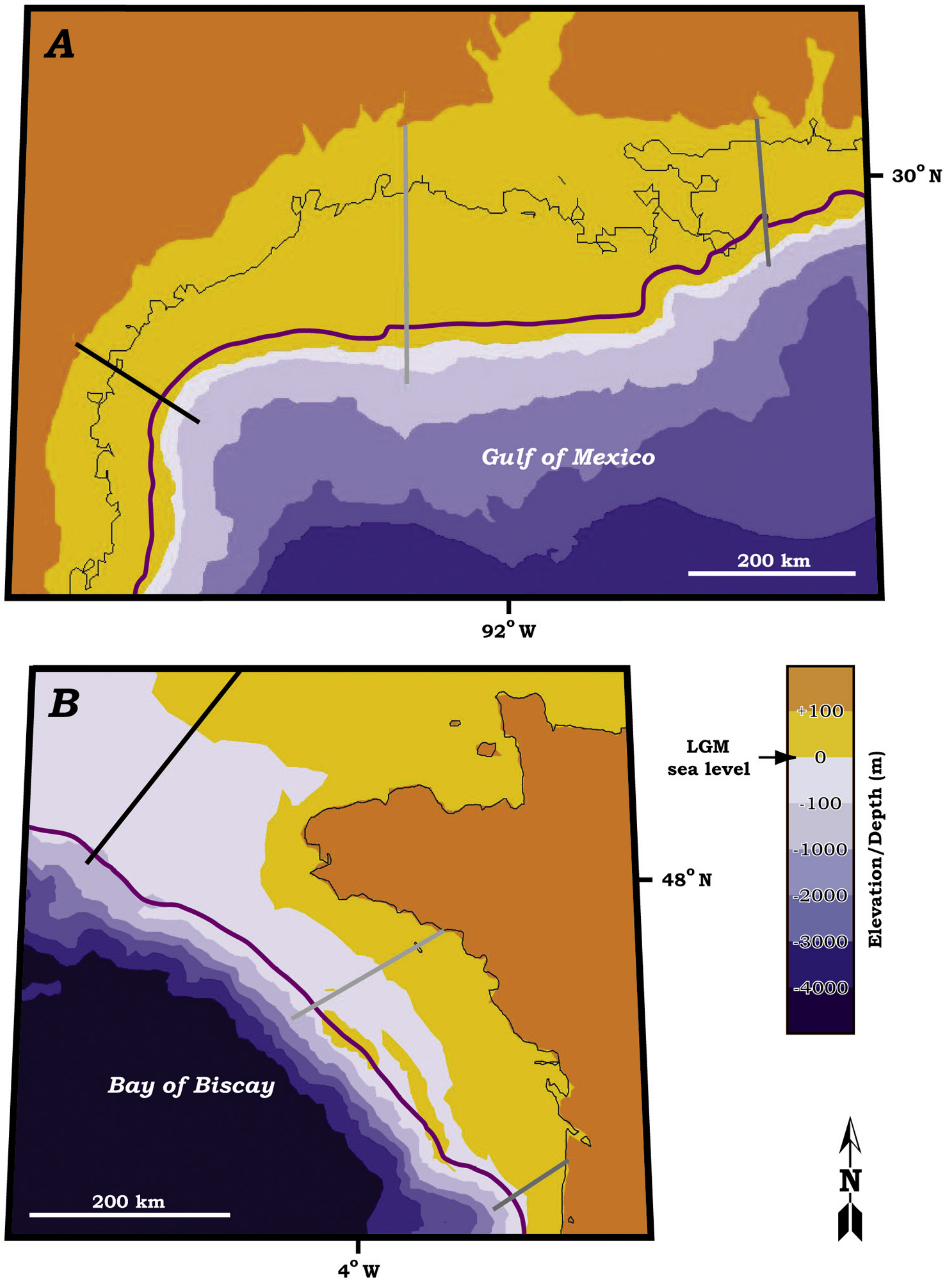

Figure 6. Modeled shoreline position during the Last Glacial Maximum in (a) the northern Gulf of Mexico and (b) the Bay of Biscay. The calculated shelf edge is indicated in purple. Note the large horizontal separation between the modeled lowstand shoreline and the shelf edge in the Bay of Biscay and the coincidence of the calculated shelf edge with two large offshore islands that existed during the LGM. Also note the intersection of the shelf edge with the tip of the bird foot of the Mississippi Delta, which is a result of slight tilting of this area due to the isostatic correction of the geophysical model. This shows that data should be interpreted cautiously when modern deltas are located at the shelf edge, a relatively rare phenomenon. Also indicated are the lines of section of Figure 7.

the modeled lowstand shoreline was located seaward of the shelf edge throughout the northern Gulf of Mexico. In the Bay of Biscay the opposite was the case: here, the modeled LGM shoreline was located well updip of the shelf edge and in the northwestern Bay of Biscay the horizontal separation between the LGM shoreline and the shelf edge increased abruptly to ~200 km.

[17] Representative cross sections (Figure 7) show the morphology of the continental shelf and uppermost continental slope in the two study areas. In the Gulf of Mexico, the shelf edge was fully exposed during the LGM, thus producing a slope increase that triggered widespread fluvial incision. In contrast, in the Bay of Biscay RSL fall did not lead to exposure of the continental slope and the lower continental

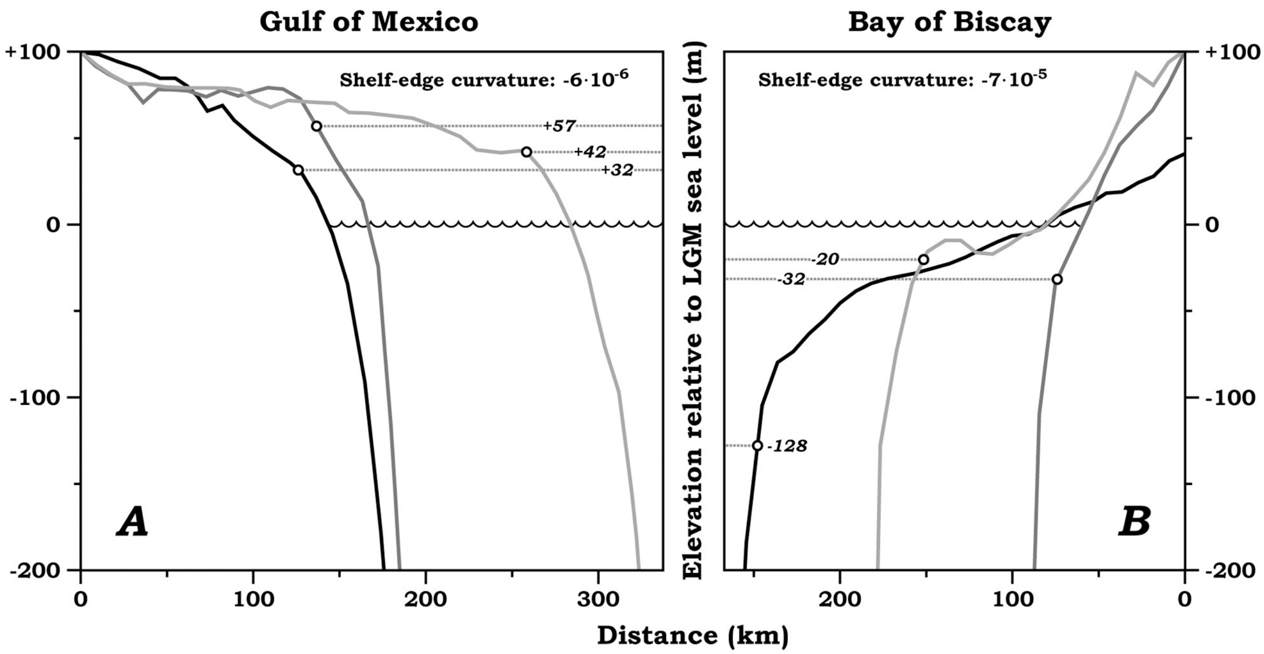

Figure 7. Representative topographic profiles from (a) the northern Gulf of Mexico and (b) the Bay of Biscay, exhibiting shelf morphology perpendicular to the shelf edge. The location of the profiles is shown in Figure 6. The open dots in the profiles indicate the calculated position of the shelf edge, with its elevation with respect to modeled Last Glacial Maximum sea level in meters. Note the distinctly different relationship between the shelf edge and the LGM shoreline for the two study areas.

shelf remained submerged. As a result, fluvial incision was limited to the highstand coastal prism as documented for the Rhine-Meuse [*Törnqvist et al.*, 2000, 2003; *Wallinga et al.*, 2004] and the Gironde [*Lericolais et al.*, 2001]. In most cases the calculated position of the shelf edge coincides with an abrupt increase of slope as shown by the topographic profiles, thus demonstrating the validity of the approach followed. The main exception is the northernmost cross section in the Bay of Biscay where the calculated shelf edge position plots lower than what would be inferred visually. The three profiles from the Bay of Biscay highlight the considerable morphometric variability of this passive margin, with average slopes increasing strongly toward the southeast. Hence it is not surprising that one curvature value cannot provide optimal results everywhere. However, these types of offsets do not invalidate the fact that the shelf edge in the northwestern Bay of Biscay occurs at the greatest depths, close to 100 m below the modeled LGM shoreline.

[18] Topographic/bathymetric profiles such as those illustrated in Figure 7 enable us to calculate both the horizontal distance and the elevation difference between the modeled LGM shoreline and the shelf edge. We performed this analysis for every 5 km along both continental margins (Figure 8). In the northern Gulf of Mexico, the mean horizontal separation between LGM shoreline and shelf edge is typically 15–30 km, yet the elevation difference is considerable (mean value of ~40 m). In stark contrast, the vertical separation between the LGM shoreline and the shelf edge in the Bay of Biscay is the opposite (disregarding the lowstand islands) and is usually at least 30 m (Figure 8). This cannot be attributed to RSL lowstands coinciding with the shelf edge during glacial maxima prior to the LGM. Recent studies [*Rohling et al.*, 1998; *Waelbroeck et al.*, 2002] show that lowstand shorelines during those time intervals were, at the very most, 20 m lower than during the LGM.

[19] It is important to note that some processes affecting the time evolution of topography and bathymetry are not accounted for in this analysis. Most prominently, the geophysical model output of 21-kyr-old land/seascapes was obtained using present-day DEM data as model input. Ideally, the input data should be corrected for morphological changes due to sedimentation and erosion that may have occurred since the LGM. For example, major depocenters like the Mississippi Delta have formed during the past ~8.5 kyr [*Stanley and Warne*, 1994], occasionally with sediment thicknesses exceeding 100 m. However, significant depositional features are fairly localized and commonly constitute topographic anomalies that can be accounted for in the data interpretation. Indeed, the bird foot of the Mississippi Delta can clearly be recognized in Figure 8, where it constitutes a relatively narrow zone. Furthermore, the Mississippi Delta is one of the few highstand deltas that have prograded to a position near the shelf edge [*Burgess and Hovius*, 1998]. Horizontal and vertical shifts of the shelf edge since the LGM may also have occurred due to growth-faulting, slumping, or local tectonic motions, none of which is accounted for in our model. However, since we focus primarily on the tectonically relatively quiescent passive margins, these effects are likely fairly small, and growth faults are most commonly associated with spatially constricted areas of rapid sedimentation around major depocenters.

[20] For all the reasons discussed above, it is of critical importance that we merely consider elevation differences

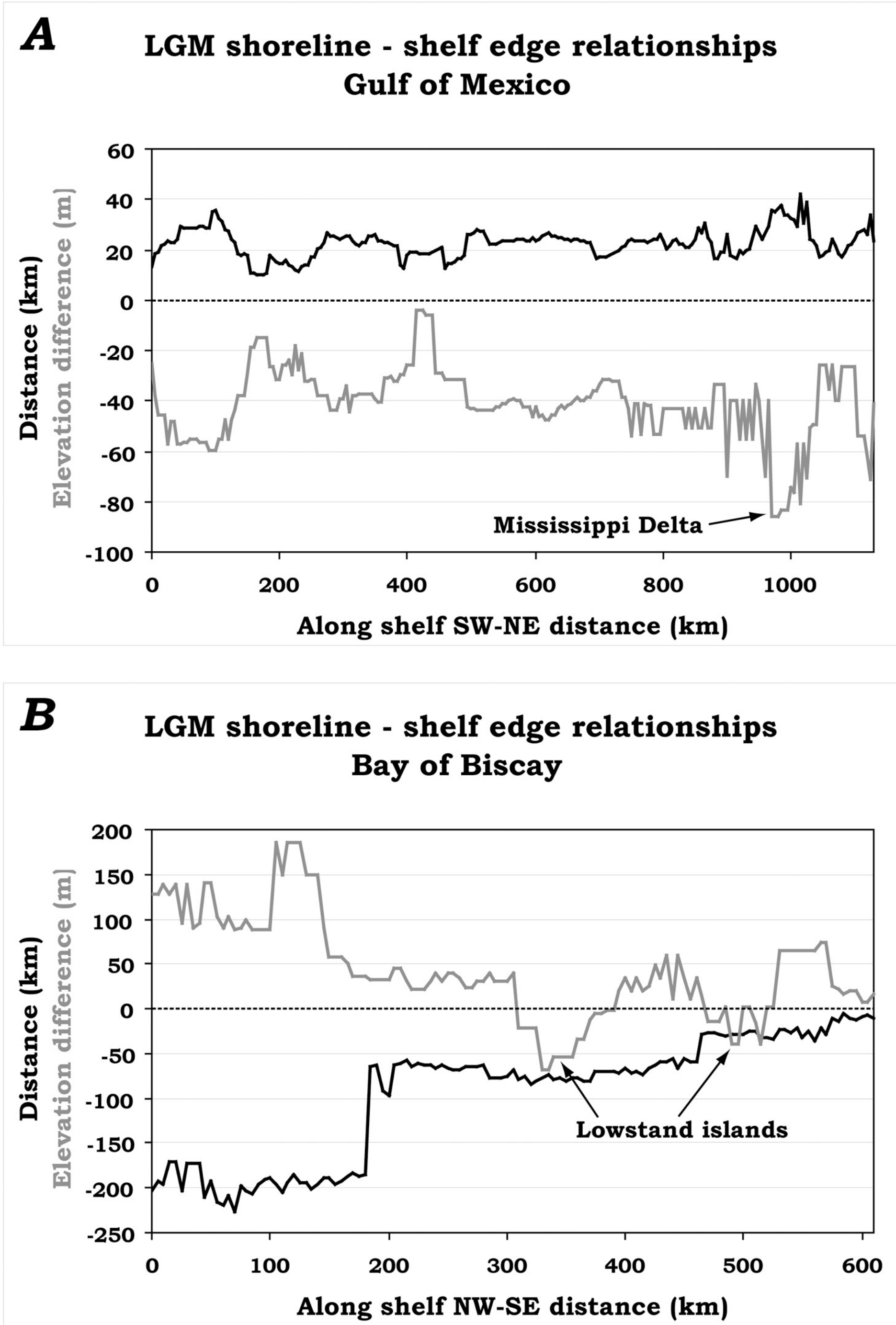

Figure 8. Plot of the modeled Last Glacial Maximum shoreline position relative to the shelf edge, both horizontally (black: distance in km) and vertically (gray: elevation difference in m), as a function of continental margin length along (a) the northern Gulf of Mexico and (b) the Bay of Biscay. The reference frame for these measurements is a smooth baseline that follows the general outline of the shelf edge, constructed with the ArcView buffer option. For every 5 km a line was projected perpendicular to this baseline using the Digital Shoreline Analysis System extension of ArcView developed by the U.S. Geological Survey [*Thieler et al.*, 2003]. Each line intersects the modeled LGM shoreline and shelf edge and enables calculation of both distance and elevation difference. LGM shoreline positions seaward and landward of the shelf edge are indicated by positive and negative values, respectively. LGM shoreline positions above and below the shelf edge are indicated by positive and negative values, respectively. Note the high elevation differences at ~1000 km in the Gulf of Mexico, coinciding with the bird foot of the Mississippi Delta. Also note the negative elevation differences at ~350 and ~500 km in the Bay of Biscay between LGM shoreline and shelf edge, coinciding with emergent lowstand islands.

between the modeled LGM shoreline and the shelf edge that are on the order of tens of meters (Figures 7 and 8), which is likely more than the effects of post-LGM sedimentation, erosion, slumping, or faulting in the vast majority of cases. For example, an analysis of high-resolution seismic data from the Gulf of Mexico offshore of Texas, an area with a comparatively thick blanket of transgressive muds, showed that the thickness of post-LGM strata near the shelf edge is typically 20 m or less [*Eckles et al.*, 2004]. If we disregard all the cases for the Gulf of Mexico where the elevation difference is less than 20 m, we still find that nearly 95% of the modeled LGM shoreline is located below the shelf edge.

5. Discussion and Conclusions

[21] The findings reported here raise some fundamental questions about the evolution of continental margins. As discussed above, the situation in the Gulf of Mexico is compatible with conventional models that predict a direct connection between the subaerial (continental) and subaqueous (deep marine) components of sediment dispersal systems during RSL lowstands. This may not be surprising given the profound role this particular area has played in the development of first-generation sequence stratigraphic models [e.g., *Boyd et al.*, 1989; *Galloway*, 1989; also see *Posamentier and Allen*, 1999]. However, the Bay of Biscay shows a fundamentally different scenario, one where the continental and deep marine segments remained physically disconnected even during the last lowstand. This finding is particularly significant when it is considered that conditions with sea level close to lowstand values lasted no more than some 10 kyr, i.e., <10% of the last glacial-interglacial cycle [*Lambeck et al.*, 2002]. One possible result of such conditions is that sediments transported from the hinterland are mostly trapped on the continental shelf and that the deep marine environment receives considerably less sediment than would be the case otherwise. Clearly, such a scenario is at odds with conventional sequence stratigraphic models and challenges our ideas about the timing of sedimentation on continental margins and the ocean floor. This, in turn, could impact the interpretation of the ancient rock record in terms of eustatic cycles [e.g., *Haq et al.*, 1987] because the relationship between deep marine sedimentation and RSL change might be much more complex than commonly assumed.

[22] In this context, one might question to what extent our findings concerning the LGM are relevant for earlier intervals of geologic history. The latter part of the Quaternary (notably the past ~800 kyr) stands out in terms of the high amplitude (100 m or more) of eccentricity-dominated glacio-eustatic cycles, while much of the Cenozoic was characterized by eustatic sea level changes with amplitudes of 20–80 m [*Miller et al.*, 2005], commonly dictated by the 40-kyr obliquity cycle [*Zachos et al.*, 2001]. The implication might be that during considerable parts of the Cenozoic sea level was less likely to drop below the shelf edge. On the other hand, periods dominated by the 10^6 yr third-order cycles would provide much longer time windows for coastal/deltaic progradation to the shelf edge. On balance, we anticipate a rich variety of possible stratigraphic responses to eustatic cyclicity that has yet to be explored to its full extent.

[23] The contrasting scenarios for our two study areas also have implications for the timing and positioning of sequence boundaries. The development of fully connected cross-shelf valleys such as in the Gulf of Mexico provides the most likely conditions for a sequence boundary to be close to LGM in age. This is supported by currently available data [*Anderson et al.*, 2004]. It appears likely that the final drop of sea level below the shelf edge during the time interval preceding the LGM enabled rapid, headward erosion and rejuvenation of the drainage system on the shelf (Figure 3). A situation with a lowstand shoreline remaining on the shelf such as in the Bay of Biscay would limit the paleovalley associated with the last glacial to an area farther updip [*Lericolais et al.*, 2001], primarily near the highstand coastal prism (Figure 4). The implication is that RSL-induced valley cutting under such conditions is completed at a much earlier stage of the glacio-eustatic cycle and the sequence boundary at the base of the valley is substantially older. For example, in the case of the Rhine-Meuse system, the last sequence boundary dates to nearly 80 ka [*Törnqvist et al.*, 2003]. These opposing conditions also show that a case can be made for both sides in the debate about the positioning of the sequence boundary [e.g., *Plint and Nummedal*, 2000; *Posamentier and Morris*, 2000]. The Gulf of Mexico scenario would favor a sequence boundary that is positioned above the falling stage systems tract while the Bay of Biscay scenario is more compatible with a sequence boundary below the falling stage systems tract [*Törnqvist et al.*, 2003]. Overall, we conclude that recent findings about the critical roles of highstand coastal prism formation, shoreface exposure during RSL fall, and the degree of exposure of the shelf edge during RSL lowstand need to be fully implemented in sequence stratigraphic models.

[24] Apart from the implications for the functioning of sediment dispersal systems and sequence stratigraphic issues, our results highlight the fundamental question of what the shelf edge represents and how it forms. Our analysis suggests that the standard model of shelf edge progradation associated with RSL lowstand does not always apply. Thus an important future research direction is to resolve the conditions under which the standard model fails. One possible avenue of exploration centers on the development of compound clinoforms [*Nittrouer et al.*, 1996] which can be observed in several large deltas. Rivers discharging in settings with high wave and tide energy develop compound clinoform geometries that consist of a subaerial delta (the delta plain including the shoreface) superimposed on a large, subaqueous delta. The lateral and vertical separation of the shoreline and the subaqueous delta rollover (the transition from subaqueous topset to foreset) is on the order of several tens of km and 30–50 m, respectively. Examples include the Amazon, Ganges-Brahmaputra, and Gulf of Papua systems [*Nittrouer et al.*, 1996; *Kuehl et al.*, 1997; *Pirmez et al.*, 1998; *Walsh et al.*, 2004]. In contrast, deltas in low-energy settings (e.g., the Mississippi River in the microtidal Gulf of Mexico) show shorelines and clinoform rollovers with negligible spatial separation [*Wright and Nittrouer*, 1995].

[25] Recent morphodynamic modeling by *Swenson et al.* [2005] highlights the relative importance of fluvial input of water and sediment vs. wave energy associated with large coastal storms in controlling clinoform morphologies under highstand conditions. In the latter, one might associate the subaqueous clinoform rollover with the shelf edge. Knowl-

edge of the response of compound clinoforms to repeated RSL cycling might shed light on the spatial relationship of the shoreline and what we typically term the shelf edge. Preliminary modeling of the stratigraphic response to RSL cycling on high-energy margins [*Pratson et al.*, 2004] suggests that the separation between lowstand shoreline and shelf edge can be on the order of 100 km.

[26] The passive margin of the Bay of Biscay is a highly energetic marine setting. However, unlike the examples mentioned above it is relatively sediment-starved, with only a few hundred meters of post-Cretaceous strata [*Bourillet et al.*, 2003]. We therefore tentatively suggest that this setting is dominated by slow thermal subsidence in the absence of sufficient sediment supply from northwest European rivers. How such sediment-starved, highly energetic systems respond to RSL change is an important area of future analysis. Our bottom line conclusion at this stage, given the data presented here, is that no universal model can be proposed that explains shelf edge evolution.

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