STRATIGRAPHIC FORWARD MODELING OF BASIN-MARGIN CLINOFORM SYSTEMS: IMPLICATIONS FOR CONTROLS ON TOPSET AND SHELF WIDTH AND TIMING OF FORMATION OF SHELF-EDGE DELTAS

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ABSTRACT: Understanding the origins of shelf-edge deltas is important because they are a key component in the basin-scale sediment routing system. Previous analyses have shown that shelf-edge deltas may be either lowstand or highstand features, depending on the magnitude of fluvial sediment supply, rate of accommodation creation, and the operation of delta autoretreat. In this paper, a stratigraphic forward model is used to generate progradational sediment wedges of the type found on many basin or shelf margins. Numerical experiments are conducted to further investigate how delta autoretreat influences shelf-edge delta formation, and to investigate relationships between sediment supply, accommodation creation, and formation of shelf-edge deltas generally. Results from the stratigraphic forward modeling are limited by the assumptions inherent in the model, but they suggest that autoretreat is more likely in systems with relatively high rates of marine sediment transport. Autoretreat appears to be least effective at low rates of sea-level rise, high rates of sediment supply, and low rates of marine sediment transport. Model results also suggest that sediment-wedge topset width (shelf width plus width of coastal plain) is controlled by the initial bathymetry into which delta progradation occurs, and the balance between sediment supply and long-term rate of accommodation creation, with higher-frequency relative sea-level oscillations playing only a minor role once the topset width has been determined by initial progradation. Based on this finding, formation of highstand shelf-edge deltas may be a process of self-regulated equilibrium regression; new phases of progradation, either forced or unforced, can reach the shelf edge by default because the shelf-edge position is a consequence of previous phases of progradation. Thus well-supplied shelf-edge deltas fed by medium to large river systems with consistent sediment supply may commonly form somewhat independently of forcing by relative sea-level oscillations, implying that throughout much of the geological record, timing of delivery of sand to deep-marine systems is unlikely to depend on simple forcing by short-term relative sea-level oscillations.

KEY WORDS: clinoform systems, shelf width, topset width, delta autoretreat, shelf-edge delta, self-regulating shelf-edge deltas, self-regulated equilibrium regression, stratigraphic forward modeling

INTRODUCTION

Shelf-edge deltas are important because once a delta has reached the edge of the shelf, marked by the shelf–slope break geomorphic feature, sediment delivered to the delta mouth is effectively bypassing the terrestrial topset area and being delivered to the top of the submarine slope, from where it can be routed down into deep-marine depositional systems (Suter and Berryhill, 1985; Porębski and Steel, 2003; Carvajal and Steel, 2006). Understanding controls on the formation of shelf-edge deltas is therefore an important component in understanding the sediment routing system generally, and is important in developing predictive models of siliciclastic sediment partitioning into the various depositional environments, from coastal plain to deep marine. Understanding timing of the formation of shelf-edge deltas with respect to relative sea-level cycles, understanding the development of large-scale basin-margin clinoforms, and understanding what controls the width and bathymetry of the shelf specifically, from shoreline to shelf–slope break, are important because all three of these factors are critical in determining how and when shelf-edge deltas can reach the shelf edge. This paper sets out to investigate these factors using experimental stratigraphic forward modeling to better understand the controls on delta progradation, shelf width, and shallow to deep-marine sediment partitioning.

Features of Basin-Margin Sediment Wedges: Topsets, Foresets, and the Shelf

Understanding the timing and process of formation of shelf-edge deltas requires a clear definition of the component features of basin-margin sediment wedges and the processes and controls involved in generating these features. At the most basic level, large-scale sediment wedges formed at basin margins in various tectonic settings are generated by delivery of siliciclastic sediment from terrestrial source areas. These sediment wedges often have a pronounced topset–foreset clinoform geometry (Fig. 1), observed at various scales, e.g., from individual bedforms, to delta lobes, to passive-margin sediment wedges. Depending on details of history of accommodation and sediment supply, part of the sediment-wedge topset may be a subaerial coastal plain, and part a subaque-
ous marine shelf (Fig. 1B). In this case, the change in topographic gradient marking the change from topset to foreset occurs at the shelf–slope break. Conversely, during lowstands of relative sea level, or in systems with relatively high sediment supply, the whole topset may be occupied by a subaerial coastal plain (Fig. 1A). In this case, little or no marine shelf is present, and the break of slope from topset to foreset is coincident with the deltaic shoreline, at least along the main axis of the sediment-transport system.

Deltas that cross the shelf are likely to be the key to shelf-margin accretion and clinoform topset width (Burgess and Hovius, 1998; Muto and Steel, 2002; Steel and Olsen, 2002; Steel et al., 2000). Repeated progradation of delta systems, fed by siliciclastic fluvial sediment input, gradually build coastal sediment prisms that eventually take on the classic clinoform geometry of siliciclastic shelf margins. If there is substantial water depth basinward of the shelf–slope break (hundreds to thousands of meters) the clinoforms may also support deep-water sands in their slope and basin-floor reaches.

When the deltaic sediment-delivery system is sited at the shelf edge the surface of the entire sediment-wedge topset is likely to be largely subaerial (Fig. 1A), regardless of highstand or lowstand timing, whereas an inner-shelf location for the delta implies a late-stage transgressive or early highstand drowned (and perhaps more muddy) shelf platform (Fig. 1B), as is the case on many modern shelves. Increments of sediment accumulated during the delta’s regressive and transgressive shelf transits, plus eventual increments on the clinoform slope and basin floor, are thus the basic building blocks of the siliciclastic shelf–slope system on continental margins (Swift and Thorne, 1991; Burgess and Hovius, 1998; Steckler, 1999; Steel and Olsen, 2002). Swift and Thorne

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Fig. 1.—Diagram illustrating two different configurations for the sediment-wedge topset. Which of the two configurations applies at any time depends, at the simplest level, on the interaction between relative sea level and sediment supply.
(1991) classified this type of active shelf depositional system as supply-dominated or allochthonous.

**Influence of Holocene Transgression**

Delta progradation is clearly not the only process influencing topset physiography. Over the past 10 ky rapid Holocene glacio-eustatic sea-level rise, combined with subsidence of the sediment–water interface due to tectonic subsidence, isostatic loading, and sediment compaction, created accommodation at a rate faster than it could be filled, even for the largest river systems. This resulted in marine transgression of many of the former deltaic shorelines and widespread flooding of the deltaic and coastal plains (commonly generating estuary or barrier coasts) that had formed, during the lowstand of the last glacial maximum, as topsets of a previous phase of delta progradation. The modern shallow-marine continental shelves were thus created (Morton and Suter, 1996; Andersen et al., 2004; Steckler et al., 1999) with water depths of up to 200 m at the shelf–slope break. Shelves with little or no additional sediment input allowed reworking of relict sediment, left over from a previous progradation and lowstand phase, and were termed autochthonous shelves by Swift and Throne (1991). Based on observation of modern flooded autochthonous shelf physiographies, they defined the term “shelf” as “shallow marine surfaces of regional extent”.

It is important to note that this may be a temporarily biased view of a prograding-wedge topset configuration. Later in a glacio-eustatic sea-level cycle, or during a supply-dominated greenhouse interval with much lower-amplitude sea-level oscillations, sediment-wedge topsets would perhaps consist predominantly of coastal and delta-plain surfaces of regional extent, with little or no marine shelf present.

**Unforced Regression and Delta Autoretreat**

It is clear that these shelf transits by delta complexes are not restricted to Pleistocene and Holocene times. Thick shelf aggradation and systematic progradation of the shelf margins by repeated regression and transgression of a deltaic sediment-delivery system are processes that have built sediment wedges throughout icehouse periods (e.g., Miocene–Pleistocene Ornoco shelf growth, Fig. 2A, from Sydow et al., 2003), and also throughout greenhouse times (e.g., Eocene Porcupine Basin shelf aggradation and progradation, Fig. 2B from Johannesen and Steel, 2005), despite reduced amplitude and perhaps different frequency of sea-level change (see also Galloway, 2001).

Assuming that unforced regression (normal regression with some component of aggradation) driven by delta progradation is also a primary mechanism of formation of sediment-wedge topsets, it is reasonable to assume further that topset width is controlled by a combination of sediment-supply volume, rate of subsidence due to tectonic, isostatic, and compactional mechanisms, rate of eustatic sea-level change, the topography and gradient of the initial surface onto which sediment prograded, and the collective geometry of the progradational clinoforms that constitute the sediment wedge (e.g., Thorne and Swift, 1991). Burgess and Hovius (1998) showed that sediment-supply magnitudes of many modern rivers should allow progradation to the shelf edge well within an interval of 1 My, even accounting for accommodation creation by compaction and for relative-sea-level rise. Muto and Steel (1992, 1997), on the other hand, have suggested that highstand progradation distances may be limited by the process of autoretreat, whereby delta progradation rates decrease through time as the delta aggrades and progrades into deeper water. Progressively more sediment is required to feed the growing delta-plain and foreset areas. Eventually insufficient sediment is available to continue progradation, and the delta system switches to a retrogradational mode, a process termed autoretreat. Muto and Steel (2002) calculated that 9 out of 24 of the Earth’s large- and medium-size deltas would turn around to transgression before reaching their shelf edges if subject to a relative-sea-level rise at rates similar to those in the late Holocene (ca. 2 m ky⁻¹), but that 15 would not. Based on these analyses, it appears that shelf-edge deltas can be either lowstand or highstand features, depending on a number of factors other than just accommodation.

The importance of autoretreat has been shown by several experimental groups (Nagasaki University and University of Minnesota, Duluth) using both geometric numerical models and flume-tank experiments (Muto and Steel, 1997; Muto, 2001; Swenson et al., 2000). Conversely, times when autoretreat is likely to be least effective, namely when rates of relative-sea-level rise are low and/or sediment supply is high, were clearly defined by Muto and Steel (1997, their figs. 3 and 4). Even if autoretreat is a commonplace autogenic phenomenon, there may also be other important mechanisms at work tending to increase the probability of highstand shelf-edge deltas.

**Influence and Identification of Forced Regression**

When deltas reach the preexisting shelf margin (shelf-edge deltas) they are commonly interpreted as lowstand features produced by forced regression, but given the various processes and controls described above, it is reasonable to assume that shelf-edge deltas are always a product of forced regression (Burgess and Hovius, 1998; Muto and Steel, 2002). Falling-stage and lowstand timing of such deltas can be demonstrated clearly in the last glacial cycle and probably in much of the Plio-Pleistocene, where there is good dating and strata can be tied to oxygen isotope curves providing an independent proxy for sea-level history. This interpretation implies that width of sediment-wedge topsets is, in part at least, controlled by repeated oscillations of relative sea level. However, in older strata where independent evidence for the magnitude and timing of relative-sea-level history is not available, and glacio-eustasy may have been less common, or may have operated at reduced amplitudes, the timing of formation of shelf-edge deltas with respect to relative-sea-level history is much less certain.

**Topset Width and Self-Regulating Shelf-Edge Deltas**

Any new phase of delta progradation, either forced or unforced, inherits a shelf–slope bathymetry that is a consequence of previous phases of progradation. The preexisting topset and shelf width is determined by the previous interaction of sediment-supply magnitude, rate of relative-sea-level change, additional accommodation created by sediment compaction, effects of initial bathymetry (e.g., Ulicny et al., 2002), and the various sediment-transport processes that have operated. If these controls remain approximately constant through time, any new phase of delta progradation should be able to build a new shelf-edge delta system, because it is merely repeating the process that occurred during previous phases of progradation. In other words, if deltas could prograde a given distance in the previous accommodation cycle, what is to stop them prograding a similar distance in the next cycle? In this case, repeated cycles of delta progradation may create highstand shelf-edge deltas, even when autoretreat operates, because the process of autoretreat has already determined the shelf width in previous cycles. Consequently shelf and shelf-margin strata show a more or less complex stacking pattern, but are probably limited by the inherited
position of the shelf–slope break, because this transition to deep water may be an effective barrier limiting progradation distance. From the above, it thus seems clear that the factors controlling topset width are complex, and perhaps less well understood than sometimes assumed. Given this apparent complexity, experimental application of a stratigraphic forward model is a potentially useful technique to investigate potential controlling factors. Is it realistic to view topset width as a self-regulating factor in shelf-edge accretion, encouraging the generation of highstand shelf-edge deltas? This paper describes results from numerical experiments conducted with Dionisos, a three-dimensional diffusional stratigraphic forward model. The numerical experiments have been designed to investigate the controls on delta progradation, sediment stacking patterns, and the width of the topsets in the resulting sediment wedge.

**MODEL DESCRIPTION**

**Formulation of the Stratigraphic Forward Model**

Dionisos is a three-dimensional numerical stratigraphic forward model developed by the Institut Français du Petrole. It is based on a generalized, modified diffusion formulation of sediment transport, where transport rate is split into a long-term component dependent on topographic slope, diffusion coefficient, and water discharge volume, and a short-term component also dependent on water velocity and inertia (Granjeon and Joseph, 1999; Granjeon et al., 2002). Long-term low-energy gravity-driven sediment flux is calculated per model cell from

\[ Q_s = \kappa S \]

where \( Q_s \) is the sediment flux, \( \kappa \) is the diffusion coefficient in \( \text{m}^2\text{yr}^{-1} \), and \( S \) is the gradient of the topographic surface at a point in the model grid. Similarly long-term, low-energy water-driven sediment flux is calculated per model cell from

\[ Q_s = \kappa Q_w S \]

where \( Q_s \) is a dimensionless number representing relative water discharge that is routed across the model grid using a simple steepest-descent algorithm (Granjeon and Joseph, 1999). As well as these two sediment-transport mechanisms, other important processes such as tectonic subsidence, flexural isostatic loading,
and eustasy, are also represented (Grajeon and Joseph, 1999), allowing construction of reasonably complex three-dimensional models.

**Parameter Values**

Parameter values for the model are given in Table 1. The model grid was 500 km by 500 km, configured with a simple ramp initial topography, with water depth increasing from 0 m on the left margin to 200 m on the right margin. Sediment was introduced at the midpoint (x = 250 km) along the left margin. Note that this means that the total width of the sediment wedge, from ultimate landward onlap pinchout to ultimate margin downlap, was not modeled. Modeling only the more distal part of the entire sediment wedge in this manner reduces computation time, and tests with model of the whole sediment-wedge system reduced at the midpoint (x = 250 km) along the left margin. Note that this means that the total width of the sediment wedge, from ultimate landward onlap pinchout to ultimate margin downlap, was not modeled. Modeling only the more distal part of the entire sediment wedge in this manner reduces computation time, and tests with model of the whole sediment-wedge system show that the selected boundary condition does not unduly influence the described behavior of the model. The sediment-supply volume of 33,750 km² My⁻¹ represents a median value from the 24 modern river-mouth suspended-load volumes given in Burgess and Hovius (1998) that span two orders of magnitude. This median value was selected to be representative of a mid-range river which, adopting an approximately uniformitarian approach, can perhaps be assumed likely to have been common throughout most of the Phanerozoic. Note however that obviously this depends on the rivers sampled in the dataset (Burgess and Hovius, 1998), which do not include, for example, rivers from the very low end of the size distribution.

Values of subsidence rate were constant across the model grid, and were kept constant throughout each model run. Rates were based on typical passive-margin values, increased to account for an element of additional subsidence due to sediment compaction. Mechanical sediment compaction can be calculated in Dionisos explicitly, but it is computationally expensive for large models, so a higher rate of tectonic subsidence has been included in the model run shown. Several of the parameters are varied in the model runs presented; the range of variation adopted represents an experimental attempt to understand the behavior of shelf–delta systems while remaining within the plausible range of parameter values.

**Applying the Model**

One significant advantage of this kind of numerical stratigraphic forward model, as compared to qualitative conceptual models, is the ability to adopt an experimental approach and to systematically investigate parameters such as sediment supply and amplitude of sea-level oscillation. This experimental method of stratigraphic forward modeling is of course limited by the quality of the model. Model quality can be gauged by considering the assumptions that underpin the model formulation (e.g., Perlmutter et al., 1999; Paola, 2000). In the case of Dionisos in this analysis, the most significant assumptions are in the use of diffusion to represent sediment transport, but because the diffusion relationship is derived from physical first principles (Grajeon and Joseph, 1999) the assumptions tend to be low-level, reducing the possibility of circular reasoning in application of the model results. However, it is still important to state unequivocally that, like all models, results should not be overinterpreted, and interpretations that are made should be checked back against the original assumptions inherent in the model to avoid circular reasoning.

**MODEL EXPERIMENTS—AUTORETREAT AND SEDIMENT-TRANSPORT RATES**

Dionisos has been used to construct a simple model of delta progradation occurring under steadily rising relative sea level. The cross section and chronostratigraphic diagram in Figure 3 show the final stratal geometry produced by the model after 2 My of simulated time. Initial delta progradation produces a sediment wedge with a distinct topset and foreset geometry (Fig. 3A). Rate of progradation decreases with time, and the stacking pattern becomes dominantly aggradational (Fig. 3B), even though the rate of relative sea-level rise is constant. The topset surface (mainly coastal-plain deposits) approaches a width of ~ 200 km, and the clinoform slopes dip into a sediment-starved basin. Progradation occurred during a time of initial significant disequilibrium between sediment supply, initial ramp bathymetry, and accommodation creation (see also Ulicny et al., 2002). The aggradational phase represents a situation whereby sediment supply and sediment-transport rates were sufficient, compared to the constant rate of relative-sea-level rise, to establish a state of quasi-equilibrium and maintain the sediment-wedge topset–foreset profile in one position on a long time scale. The quasi-equilibrium state, with an aggradational shoreline trajectory, appears to indicate that delta autoretreat was ineffective and the autoretreat trajectory was incomplete. Longer-duration model runs show that this aggradational pattern can persist with these particular model parameters for more than 10 My of elapsed model time.

This apparent absence of autoretreat in the modeled system highlights the conditions under which autoretreat is least effective (i.e., high sediment–supply rates and/or low rates of rise, as outlined by Muto and Steel, 1997, their figs. 3 and 4). However, it should also be noted that autoretreat as investigated so far operates on fourth-order time scales in essentially icehouse systems (Muto and Steel, 2002), and maybe not on much longer time scales (third-order) of the shelf-edge trajectory demonstrated in the above experiment. It is also important to note that additional processes that could contribute to delta autoretreat, such as slope failure and related sediment-transport processes, have not been included in the model run shown. Additional sediment-transport processes, or just higher rates of diffusional transport, may lead to autoretreat behavior because they tend to remove material from the delta front into deeper water in more distal parts of the basin and perhaps shift the deposition–accommodation balance enough to precipitate transgression.

Figure 4 shows a plot of shoreline trajectory for six models with varying values of marine sediment-transport rate represented by varying the marine diffusion coefficients for sand- and mud-grade sediment (Grajeon and Joseph, 1999). Increasing

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**Table 1.—Values of model parameters.**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Source reference</th>
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<tr>
<td>Grid length x direction</td>
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<td></td>
</tr>
<tr>
<td>Grid length y direction</td>
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<tr>
<td>Grid cell length and width</td>
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<td></td>
</tr>
<tr>
<td>Model duration</td>
<td>2 My</td>
<td></td>
</tr>
<tr>
<td>Model time step</td>
<td>0.001 My*</td>
<td></td>
</tr>
<tr>
<td>Tectonic subsidence rate</td>
<td>200 m My⁻¹</td>
<td></td>
</tr>
<tr>
<td>Elastic thickness</td>
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<td></td>
</tr>
<tr>
<td>Sediment supply</td>
<td>33,750 km² My⁻¹</td>
<td></td>
</tr>
<tr>
<td>Sand%</td>
<td>21.4</td>
<td>Burgess and Hovius (1996)</td>
</tr>
<tr>
<td>Mud%</td>
<td>78.6</td>
<td></td>
</tr>
<tr>
<td>Terrestrial x sand</td>
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<td></td>
</tr>
<tr>
<td>Marine x sand</td>
<td>0.1 m³ yr⁻¹</td>
<td></td>
</tr>
<tr>
<td>Terrestrial x mud</td>
<td>500 m³ yr⁻¹</td>
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</tr>
<tr>
<td>Marine x mud</td>
<td>1 m³ yr⁻¹</td>
<td></td>
</tr>
</tbody>
</table>

* Note that the time step used in certain algorithms may be shorter for reasons of numerical stability.
values of marine diffusion coefficient create a delta system with higher rates of down-slope sediment transport operating on the delta foresets, leading to less deposition in the proximal area around the delta mouth and at the top of the foresets, and more deposition lower down the foresets in a more distal marine setting. Thus these model results show a transition from aggradational stacking to retrogradational stacking with increasing rates of marine transport. Retrogradation in these cases is driven by delta autoretreat, suggesting that rate of sediment transport on the delta front may well be a key control on occurrence of delta autoretreat. When sediment can be efficiently removed from the river mouth and transported basinwards, autoretreat may tend to occur. In less efficient transport systems, where fluvially delivered sediment remains in a more proximal position, supporting construction of delta-front clinoforms, autoretreat appears to be ineffective, at least on the time scales represented in this modeling. This raises an outstanding question of how efficient delta-front sediment-transport systems are; which of the low to the high transport-rate values used here are the most realistic?

CONTROLS ON TOPSET WIDTH

Amplitude of Eustatic Sea-Level Oscillations

Dionisos can be used in investigate the extent to which topset width is controlled by amplitude of relative-sea-level oscillations and the ratio or interaction between longer-term accommodation creation and sediment supply. A series of model runs were executed with 400-ky-period eustatic sea-level oscillations of various amplitudes superimposed on a 200 m Myr\(^{-1}\) rate of long-term relative-sea-level rise. Sediment supplies of 20,000 km\(^3\) Myr\(^{-1}\), 30,000 km\(^3\) Myr\(^{-1}\), 40,000 km\(^3\) Myr\(^{-1}\), and 50,000 km\(^3\) Myr\(^{-1}\) were used for each amplitude of eustatic sea-level oscillation. Shoreline trajectories on chronostratigraphic diagrams were then analyzed for each sequence produced by the 400-ky-period eustatic oscillations, and topset progradation distance was recorded. Mean and maximum progradation distances from this analysis are plotted against sediment supply for each of the various eustatic oscillation amplitudes (Fig. 5).
oscillation does exert some control on progradation distance and hence on topset width. For example, for a sediment supply of 30,000 km³ My⁻¹, increasing the amplitude of relative-sea-level oscillation from 25 m to 100 m increases the progradation distance and hence the topset width from 84 km to 130 km. However, it is also important to note that even in these cases with higher sediment-transport rates, there is still a basic linear relationship between sediment supply and progradation distance; a four-fold increase in amplitude of sea-level oscillation increases progradation distance and shelf width by a factor of approximately 1.4, but increasing sediment supply by a factor of 2.5 increases shelf width by a factor of approximately 1.6. So, in cases of higher marine transport rate, sea-level oscillations play a role in determining shelf width, but sediment supply is still the more dominant control.

**Self-Regulating Sediment-Wedge Topset Widths**

Patterns of shoreline trajectory and shelf–slope break progradation shown above suggest that, for a given sediment supply and rate of accommodation creation, deltas prograde across a ramp bathymetry until they reach a point of quasi-balance between supply, accommodation creation, and delta-front bathymetry, at which point they switch to an aggradational phase or a retrogradation autoretreat phase (e.g., Fig. 7). Note that this occurs under conditions of steady accommodation creation. No increase in rate of accommodation creation is required to generate the switch from progradation to aggradation (see also, e.g., Muto and Steel, 1992; Muto and Steel, 1997; Muto, 2001). This observation of model behavior raises an interesting possibility, namely that this progradation distance, as a component of the whole sediment-wedge topset width, represents a quasi-equilibrium state for the sedimentary system under given parameter values. If this is the case, the system should be able to return to this state even after significant perturbations. In other words, once this progradation distance has been established, the delta system may be capable, by default, of repeatedly returning to the shelf edge during relative-sea-level highstands, simply because the topset width is defined by the distance of previous progradation episodes. Thus formation of highstand shelf-edge deltas may be a process of unforced, self-regulated equilibrium regression.

To investigate how this process might operate, Dionisos was run with the standard model parameters (Table 1) except for a time-varying sediment supply operating in on-off mode, with four pulses of 33,750 km³ My⁻¹ supply, the first 1.5 My in duration and the latter three 0.5 My in duration. (Fig. 7A, B). At the start of the model run, initial progradation is rapid, but rates decrease as the delta front progrades into deeper water, and after 1 My elapsed model time, the stacking pattern is aggradational. Shutdown of sediment supply at 1.5 My elapsed model time leads to rapid transgression and flooding of the delta topset. Interestingly, this creates a significant erosional subaerial hiatus on the coastal plain. When sediment supply recommences, progradation is again rapid, but most importantly, the shoreline does not prograde beyond the position of the shelf–slope break established by the previous phase of progradation. This demonstrates that with repeated same-magnitude pulses of sediment supply, new phases of delta progradation during steadily rising relative sea level can create shelf-edge deltas, merely by replicating the previous phase of delta progradation.

A further test of this self-regulating process of delta progradation is presented in Figures 7C and 7D. In this case, sediment supply is pulsed, but the magnitude of the pulses varies, ranging between 16,000 and 50,000 km³ My⁻¹. The first 33,750 km³ My⁻¹
Pulse of sediment supply causes rapid initial progradation, slowing to aggradation as equilibrium between supply and delta-front bathymetry is established. A second pulse of 16,000 km³ My⁻¹ sediment supply causes the delta system to prograde 180 km across a shelf 240 km wide. The third pulse of sediment supply of 33,750 km³ My⁻¹ generates 200 km of progradation. Delta foresets from this progradation phase extend over the shelf–slope break.

A final pulse of sediment supply of 50,000 km³ My⁻¹ is almost twice the magnitude of the previous pulse and 1.48 times the volume of the initial pulse, yet the progradation does not extend beyond the position of the shelf–slope break established by the first phase of progradation because the slope bathymetry and slope gradient are too high to allow significant further progradation.

These model results demonstrate the possibility that once topset width is established by an initial phase of delta progradation, subsequent phases of progradation, even with relatively reduced sediment supply, generate deltas that cover a significant portion of the shelf width. Conversely, progradation driven by a pulse of sediment supply greater than that responsible for the initial phase may not extend significantly beyond the originally established shelf–slope break because of the limiting effects of the bathymetry beyond the delta front.

**FIG. 5.**—A plot of sediment supply rate against mean and maximum shoreline progradation distances, which is an indicator of topset width, for model runs with standard parameter values (Table 1) but various sediment-supply rates. This plot shows that there is a simple linear relationship between topset width and magnitude of sediment supply, and that amplitude of sea-level oscillation has little or no impact on topset width.

**FIG. 6.**—A plot of sediment supply rate against mean and maximum shoreline progradation distances, which is an indicator of topset width, for model runs with the same parameter values as in Figure 4 apart from higher values for sediment transport rates ($\kappa = 5$ m² yr⁻¹ for marine sand, $\kappa = 50$ m² yr⁻¹ for marine mud). This plot shows that there is an approximately linear relationship between topset width and magnitude of sediment supply, but in these high-transport-efficiency cases amplitude of sea-level oscillation does exert some control on progradation distance and hence on topset width.

**DISCUSSION**

The results presented above show that for the particular stratigraphic forward model formulation in Dionisos, and for certain sets of model processes and parameters, the process of delta autoretreat is dependent on sediment-transport rates on the delta foresets and in the marine environment generally. Autoretreat appears to be least effective at low rates of sea-level rise and high rates of sediment supply (cf. Muto and Steel, 1997). Model results also suggest that sediment-wedge topset width is controlled predominantly by the initial bathymetry in which the sediment wedge forms, and the relationship between sediment supply and long-term rate of accommodation creation, with higher-frequency relative-sea-level oscillations playing only a minor role once the topset width is determined. Formation of hightstand shelf-edge deltas may be a process of self-regulated equilibrium regression; new phases of progradation, either forced or unforced, can reach the shelf edge by default because the shelf-edge position is a consequence of previous phases of progradation. In other words, if the deltas could do it before, they can do it again.
The model results presented here have some other interesting implications. If the modeled process of self-regulating formation of shelf-edge delta operates, shelf-edge deltas are unlikely to be only lowstand forced-regressive features, and bypass of sand-grade sediment into deep-marine systems may not be limited to times of relative-sea-level lowstand. This is particularly true during greenhouse times, when high-frequency sea-level oscillations did not occur, or occurred with lower amplitude. This tends to support the previous analyses of Burgess and Hovius (1998), Muto and Steel (2002), and Carvajal and Steel (2006). However, highstand deltas tend to have significant aggradational topsets, unlike lowstand delta systems fed by up-dip incised fluvial systems, which are typically considered less likely to experience aggradational deposition. The fraction of sand-grade sediment
that is trapped in such aggradational highstand delta topsets is uncertain, being dependent on various factors such as fluvial discharge, the bedload transport capacity of the fluvial system, the morphology of the floodplain, and the frequency and magnitude of overbank flood events (e.g., Schumm, 1993). Better understanding of fluvial response to base-level change is required to address this.

As always with application of models, results should be interpreted carefully considering the limitations imposed by the assumptions that underpin the model, both in the choice of modeled processes and in the realism of accuracy of their representation (e.g., Perlmutter et al., 1999; Paola, 2000; Burgess et al., 2006. The safest way to view these results is as a series of numerical experiments that pose a testable hypothesis of delta progradation and formation of basin-margin sediment wedges.

The main assumptions in Dionisos are in the use of diffusion to represent sediment transport, but because the diffusion relationship is derived from physical first principles (Granjeon and Joseph, 1999) the assumptions tend to be low-level. Based on this, Dionisos is probably a reasonably reliable model for representing long-term large-scale sediment-transport processes and their stratigraphic products. Specifically significant to the results presented here is the absence in the modeling of slope failure and related mass-transport processes that could act on the slope system, especially as it steepens, and enhance progradation by transferring material into deep water. This has the capacity to complicate the observed relationship between topset width and eustatic oscillations, and requires further investigation. Running Dionisos and other stratigraphic forward models with additional sediment-transport processes that could contribute to slope progradation (e.g., hemipelagic deposition from a suspended plume, turbidity-current transport from hyperviscous flows) would therefore be useful.

Other modeling studies have previously represented the shelf–slope break as a physiographic feature that can possibly prograde independently of deltaic processes (e.g., Burgess and Allen, 1996; Steckler, 1999; Carey et al., 1999). Hence further study of modern slope systems, to understand better how the shelf–slope break may prograde in a submarine realm, without direct input from a shelf-edge delta, would also be valuable. Further analysis of topset strata and progradation architectures in outcrop (e.g., Johannessen and Steel, 2005) and via subsurface datasets is also required, to determine, for example, if instances of shelf-edge deltas formed by repeated episodes of unforced regression exist.

CONCLUSIONS

1. For the particular formulation of a stratigraphic forward model in Dionisos, and for certain sets of model parameters, the process of delta autotreat is dependent on sediment-transport rate on the delta foresets and in the marine environment generally. Autotreat appears to be least effective at low rates of relative-sea-level rise, high rates of sediment supply and low rates of delta-front sediment transport. This supports previous analyses of the autotreat process, but further testing with different formulations of stratigraphic forward model is probably necessary.

2. Model results suggest that clinoform topset width is controlled predominantly by the initial bathymetry into which delta progradation occurs, and the balance between sediment supply and long-term rate of accommodation creation, with higher-frequency relative-sea-level oscillations playing only a minor role once topset width is determined by initial progradation.

3. Formation of highstand shelf-edge deltas may be a process of self-regulated equilibrium regression; new phases of progradation, either forced or unforced, can reach the shelf edge by default because the shelf-edge position is a consequence of previous phases of progradation. Thus shelf-edge deltas may commonly form largely independently of forcing by relative-sea-level oscillations, implying that throughout much of the geological record, timing of delivery of sand to deep-marine systems is unlikely to reflect simple forcing by high-frequency relative-sea-level oscillations.

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