CLINOFORM PROGRADATION BY TURBIDITY CURRENTS: MODELING AND EXPERIMENTS

THOMAS P. GERBER,1 LINCOLN F. PRATSON,1 MATTHEW A. WOLINSKY,2 RON STEEL,1 JERÉ MOHR4*  
JOHN B. SWENSON,4 AND CHRIS PAOLA2

1Division of EOS, Nicholas School of the Environment and Earth Sciences, Old Chemistry Box 90227, Duke University, Durham, North Carolina 27708, U.S.A.  
2Department of Geological Sciences, The University of Texas at Austin, 1 University Station, C-1110, Austin, Texas 78712, U.S.A.  
3St. Anthony Falls Laboratory, University of Minnesota Twin Cities, 2 Third Avenue SE, Minneapolis, Minnesota 55414, U.S.A.  
4Department of Geological Sciences and Large Lakes Observatory, University of Minnesota-Duluth, 229 Heller Hall, 1114 Kirby Drive, Duluth, Minnesota 55812, U.S.A.  
e-mail: tpg@duke.edu

ABSTRACT: Clinoform morphologies and growth patterns are typically viewed as the product of a particular mode of sediment transport, but process-specific models for their generation from turbidity currents are few, despite observations of turbidity currents on modern clinoforms and turbidites in ancient clinoform deposits. We present a simple morphodynamic model, supported by laboratory experiments, which shows how net-depositional turbidity currents can build sedimentary clinoforms. Most conceptual models for clinoform evolution assume that under constant (sea level, sediment supply) forcing progradation occurs via continuous basinward migration of a depocenter localized near the clinoform rollover. Abrupt basinward shifts in depocenter location are therefore an indication of allogenic variability in forcing. In contrast, our results document a unique style of progradation driven by autogenic cycles of slope steepening, sediment bypass, and depositional backstepping on the foreset slope. In experiments designed to investigate this slope–flow feedback, deposition from a continuous turbidity current transporting fine sand and silt repeatedly steepened a clinoform foreset to a graded slope which bypassed sediment to the slope base, depositing a sediment lobe. Continued deposition then caused the lobe to backstep up the slope, building a lower-slope foreset and eventually reinitiating the cycle. Our model shows how this cyclic depositional pattern arises from a morphodynamic feedback between the foreset gradient and the rate of sediment resuspension by the overriding turbidity current. Using our model to scale experimental results to the field, we predict gradual foreset steepening in prograding turbidity-current-dominated clinoform strata, with slope grading and cyclic deposition favored where these clinoforms build over steep basin slopes under conditions of high sediment supply and/or lateral confinement. We demonstrate how our results can add new insight to process-based interpretation of clinoform strata by application to a modern and an ancient field example.

INTRODUCTION

A characteristic shape of strata along both deltas and continental margins is the sedimentary clinoform. These settings also share an important formative process in turbidity currents. Bypass and erosion of delta foresets and continental slopes by turbidity currents are widely viewed as processes that build delta bottomsets, excavate submarine canyons, and generate submarine fans (Walker and Mutti 1973; Piper 1978; Shepard 1981; Kostic and Parker 2003a, 2003b). But turbidites, the characteristic deposits of turbidity currents, are also the basic facies in prograding clinoform strata along shelf margins (Steckler et al. 1999; Posamentier et al. 1991; Johannessen and Steel 2005). In these settings, depositional turbidity currents play a fundamental role in determining clinoform shapes and internal stacking patterns, but how this occurs is largely unexplored (Fig. 1).

Two recently proposed process models offer a starting point for addressing the question. Observations along inner-shelf clinoforms that lie seaward of rivers with high sediment discharge show that wave-driven sediment resuspension can initiate turbidity currents, which then flow down clinoform foresets under the influence of gravity (Traykovskiy et al. 2000; Wright et al. 2001; Wright et al. 2002). These findings have led Friedrichs and Wright (2004) to propose an analytical model in which the shallow, wave-dominated portions of clinoforms are in morphodynamic equilibrium with turbidity currents. Wave-generated turbulence along this graded reach of the clinoform diminishes with increasing water depth, leading to purely density-driven flow over the foreset. Parker (2006) extended this view by proposing a model in which clinoform progradation is driven by deposition across the foreset from weak turbidity currents descending below wave base. Rather than being a static feature that turbidity currents bypass en route to deepwater, the clinoform evolves from them. Like Friedrichs and Wright, however, Parker (2006) assumes morphologic equilibrium for the clinoform by invoking an equilibrium shape that does not vary during progradation.

If shelf clinoforms are purely the product of turbidity currents, their stratigraphic architecture should reflect elements of both the Friedrichs and Wright (2004) and Parker (2006) models. The former model suggests a tendency for clinoform slopes to approach grade. (Note that in this paper we use “grade” to describe a condition where no net surface aggradation or degradation occurs over morphologically significant time scales, as opposed to vertical grain-size trends in event deposits.) The latter model suggests that turbidity-current deposition can dominate
sediment storage on clinoform foresets and drive progradation. A simple thought experiment hints at a possible connection. Turbidity currents continuously deposit and resuspend sediments as they move at rates that depend among other factors on bed slope (Bagnold 1962; Pantin 1979; Parker et al. 1986). If the foreset slope of a clinoform is steep enough to induce sediment resuspension, it gradually weakens and dies out because the loss of suspended sediment decreases the excess density that drives its motion. A waning turbidity current leaves behind a basinward-thinning deposit. The buildup of multiple such deposits gradually steepens the foreset, increasing the gravity force driving subsequent turbidity currents and reducing their net rate of deposition. Thus deposition from repeated turbidity currents progressively transforms a clinoform as suggested by Parker (2006), but with a morphology that should evolve through time. Deposition eventually steepens the foreset to a graded slope in equilibrium with the flow. This leads to significant bypassing of the clinoform foreset just as the wave-generated equilibrium envisioned by Friedrichs and Wright (2004) causes flows to bypass the shallow reaches of the clinoform.

In this paper, we develop a model that combines elements of the Friedrichs and Wright (2004) and Parker (2006) models and formalizes the feedback between turbidity-current dynamics and clinoform evolution. We use this model to explore the clinoform foreset dynamics and turbidity-current sediment discharges under which deposition and resuspension of sediments occur. We then present results from simple flume experiments that support basic predictions of the model. Finally, we discuss our results with reference to field observations of clinoforms in settings where turbidity currents play a significant role in strata formation.

ANALYSIS

Model Derivation

The 1-D, layer-averaged conservation equations for momentum (Equation 2.1), fluid (Equation 2.2), and suspended sediment (Equation 2.3) for a dilute turbidity current (neglecting pressure forces) carrying sediment of uniform size are given by (Pantin 1979; Parker 1982; Parker et al. 1986):

\[
\frac{\partial Q_u}{\partial t} + \frac{\partial U Q_u}{\partial x} = Rg CH \sin(\theta) - \frac{\tau_b}{\rho} \tag{2.1}
\]

\[
\frac{\partial H}{\partial t} + \frac{\partial Q_u}{\partial x} = E_w U \tag{2.2}
\]

\[
\frac{\partial CH}{\partial t} + \frac{\partial Q_s}{\partial x} = -\omega_b \rho_0 C + F_c \tag{2.3}
\]

In Equations 2.1–2.3, x is a streamwise coordinate tangent to the bed, and \( U, C, \) and \( H \) are respectively the layer-averaged streamwise velocity, layer-averaged volumetric sediment concentration, and flow thickness. The downstream fluxes of water and sediment are given by

\[
Q_u = U H, \quad Q_s = U C = C Q_c \tag{2.4}
\]

\( Rg \) denotes the submerged specific weight of the sediment (given by \( \rho_f / (\rho_f - 1) g \), where \( \rho_f \) is the grain density and \( \rho \) is the fluid density) and \( \theta \) is the bed slope angle. The turbulent bed shear stress is \( \tau_b \), and \( E_w \) is the coefficient of fluid entrainment. The grain settling velocity, \( \omega_b \), and near-bed concentration parameter \( \rho_0 > 1 \) define the deposition term in Equation 2.3, while \( F_c \) denotes the resuspension of sediment (turbulent Reynolds flux) at the bed.

The fluid entrainment coefficient \( E_w \) depends on the properties of the flow, and is typically closed using the following relations (Parker et al. 1986):

\[
E_w = \frac{a}{b + Rg b} \tag{2.5}
\]

\[
Rb = \frac{Rg CH}{U^2} = \frac{Rg Q_c}{U^3} \tag{2.6}
\]

In this formulation, the rate of fluid entrainment is a function of the flow’s bulk Richardson number \( Rb \) and the empirical constants \( a \) and \( b \).

In this work we focus on large-scale evolution of the bed under sustained flows. The timescale for bed evolution is long compared to both the time for unsteady passage of the flow front and the residence time of transient suspended sediment storage in the water column (Paola and Voller 2005). Hence we approximate the flow as quasi-steady (i.e., time derivatives in Equations 2.1–2.3 are negligible). While the flow is steady on “hydrodynamic” timescales, on “morphologic” timescales the flow evolves due to gradual changes in the underlying bed morphology, which evolves according to the Exner equation:

\[
(1 - \lambda) \frac{\partial \eta}{\partial t} = -\frac{\partial Q_s}{\partial x} \tag{2.7}
\]

The Exner equation represents changes in bed elevation \( \eta \) as a function of downstream changes in sediment flux \( Q_s \) (moderated by the reference bed porosity \( \lambda \)). We do not include subsidence or compaction terms in Equation 2.7, so \( \eta \) changes only from the exchange of sediment between the flow and bed.

Following previous work (Middleton 1966a, 1966b; Stacey and Bowen 1988; Pirmeez and Imran 2003), we consider sustained currents with approximately uniform velocity (\( DU/Dt = 0 \)). This approach assumes that inertial accelerations are small relative to the gravity and resistance forces acting on the flow and (along with our neglect of pressure gradients in Equation 2.1) is equivalent to neglecting backwater effects in open-channel flow. Under this assumption, combination of Equations 2.1 and 2.2 gives a Chezy-type momentum balance:

\[
E_w U^2 + \frac{\tau_b}{\rho} = Rg CH \sin(\theta) \tag{2.8}
\]

Here we note that the effect of ambient water entrainment at the upper flow interface is a drag \( (E_w U^2) \) that adds to the resistance from the bed.
Substituting the (turbulent) quadratic drag law with friction coefficient $C_d$ for the bed shear stress ($\tau_b = \rho C_d U^2$) into Equation 2.8 and rearranging provides an expression for the velocity:

$$U^2 = \frac{RgCH \sin(\theta)}{(C_d + E_w)}$$

Rearranging Equation 2.9 and substituting Equation 2.6 shows that:

$$R_h = \frac{C_d + E_w}{\sin(\theta)} = Fr_d^{-2}$$

Here, $Fr_d$ is the densimetric Froude number, which governs whether flow is supercritical ($Fr_d > 1$) or subcritical ($Fr_d < 1$), and is a function of slope and flow resistance. Combining Equation 2.10 with Equation 2.5 yields a quadratic form for $E_w$ with solution

$$E_w = \frac{1}{2} \left[ -(C_d + b \sin(\theta)) + \sqrt{(C_d + b \sin(\theta))^2 + 4a \sin(\theta)} \right]$$

We expand Equation 2.3 with a steady suspended-sediment balance and use a McCave-Swift (1976) form for the resuspension term $F_r$:

$$\frac{dQ_s}{dx} = -\omega_r r_0 \left(1 - \frac{\tau_b}{\tau_r}\right)$$

where $r$ represents a critical bed shear stress for full suspension. Substituting the definition $Q_s = Q_s/Q_r$ into Equations 2.12 and 2.7, and the quasi-steady approximation into Equation 2.2 gives

$$\frac{dQ_s}{dx} = E_w U$$

$$\left(1 - \frac{\tau_b}{\tau_r}\right) = \frac{-dQ_s}{dx} = \omega_r r_0 \frac{Q_s}{Q_r} \left(1 - \frac{\tau_b}{\tau_r}\right)$$

Equations 2.13 and 2.14 explain the mass-balance relation for the system. The growth of the bed requires net deposition, which reduces the flux of suspended sediment in the flow. Deposition by settling is directly affected by the turbidity-current deposition that drives progradation. Deposition by settling is directly affected by the turbidity-current deposition that drives progradation. The feedback between interplay between deposition and bypassing is thus embodied in the relationship between bed shear stress $\tau_b$ and bed slope sin$(\theta)$.

To capture this relationship, we use the drag law (Equation 2.9) to eliminate $U$ from Equation 2.8 and express bed shear stress $\tau_b$ in terms of sediment flux, $Q_s$:

$$\tau_b = \rho \left[ Rg \sqrt{C_d \left( \frac{Q_s \sin(\theta)}{1 + E_w/C_d} \right)} \right]^{2/3}$$

Importantly, this shows that bed stress $\tau_b$ is primarily a function of the densimetric stream power, $Q_s \sin(\theta)$, in contrast to rivers, where $\tau_b$ is a function of the sediment-free stream power $Q_0 \sin(\theta)$. Equation 2.15 also shows that surface drag due to fluid entrainment (i.e., $E_w > 0$) decreases bed drag (reflected in $\tau_b$), so that overall flow resistance in the momentum balance is maintained.

The shear stress, $\tau_c$, for full suspension is assumed to have the following form:

$$\tau_c \approx \rho \omega_r^2$$

where $\omega_r$ is a critical grain settling velocity that serves as a proxy for the critical shear velocity (Bridge and Bennett 1992; Kneller and McCaffrey 1999; Leeder 1999). Combining Equations 2.16, 2.15, and 2.14 gives

$$\left(1 - \frac{\tau_b}{\tau_r}\right) = \frac{\omega_r r_0 Q_s}{Q_r} \left(1 - \frac{Rg \sqrt{C_d}}{\omega_r^2} \left( \frac{Q_s \sin(\theta)}{1 + E_w/C_d} \right) \right)^{2/3}$$

Together, Equations 2.17, 2.13, 2.11, and 2.9 can be used to investigate the growth of strata subject to steady net depositional flows, where $\tau_b \leq \tau_r$. In the limit of nondeposition ($\tau_b = \tau_r$), the last term in Equation 2.17 equals one, providing an expression for the slope angle ($\theta_r$) at which sediment bypassing should occur:

$$\sin(\theta_r) = \left(1 + \frac{E_w}{C_d} \right) \frac{\omega_r^3}{Rg \sqrt{C_r \omega_r}} = \left(1 + \frac{E_w}{C_d} \right) \sin(\theta_r)$$

where we have introduced $\theta_r$, the slope angle for bypassing in the absence of entrainment (i.e., $E_w = 0$). Equation 2.18 shows that the effect of fluid entrainment ($E_w > 0$) is to increase the bypass slope, i.e., entrainment decreases the bed shear stress (Equation 2.15), which must be compensated for by an increase in slope in order to maintain the critical stress for full suspension (Equation 2.16).

Because the entrainment coefficient is a function of bed slope (Equation 2.11), to quantify the effect of entrainment on the bypass slope we must eliminate $E_w$ from Equation 2.18. We use Equations 2.10 and 2.5–2.6, along with $\omega_r = \sqrt{C_d} U$, to eliminate $E_w$ algebraically, giving

$$\sin(\theta_r) = 1 + \frac{a}{C_d} \left( \frac{Rg \sqrt{C_d}^3 \omega_r}{Q_s + b} \right)^{-1}$$

In summary, then, our 1D morphodynamic model consists of the layer-averaged conservation equations for fluid, momentum, and suspended sediment for a turbidity current but with the following simplifications: (1) pressure forces are neglected; (2) the hydrodynamic model is quasi-steady; (3) backwater effects in the momentum equation are negligible, permitting a Chezy-type balance; (4) deposition and resuspension of sediment occur simultaneously, and are adequately described using a single representative grain settling velocity in combination with a McCave-Swift Reynolds flux closure; (5) resuspension never exceeds deposition, so the flows are net-depositional, and (6) no account is made of flow intermittency, so that the Exner expression (Equation 2.17) implies a bed subject to continuous flow (this last assumption is not restrictive and can be easily modified to consider realistic time scales for bed evolution). Note that we use the term resuspension when referring to the addition of sediment to the flow, whereas entrainment is used only to refer to the addition of ambient (sediment-free) fluid.

The form of our model reflects a desire to capture those processes essential for generating the feedback proposed above while avoiding unnecessary complexity. Our model differs from that of Friedrichs and Wright (2004) in that (1) we deal with turbidity currents where resuspension is due to turbulence generated by the flow itself (vs. wave-supported flows), and (2) successive flows evolve in response to changes in seabed morphology caused by previous turbidity-current deposition. Our model is simpler than that of Parker (2006) in terms of hydrodynamics but is more general in that it does not restrict clinoform progradation to traveling waves (i.e., constant in shape). Instead, the shape of the clinoform changes as it progrades according to how the clinoform slope affects the turbidity-current deposition that drives progradation.

**Nondimensional Form**

We now cast our model in a nondimensional form in order to explore its behavior without having to uniquely specify the turbidity current.
discharge and grain size. To do this we measure distances relative to a reference settling length $L_s$, and we measure slopes relative to a reference friction slope $S_f$.

In the absence of entrainment ($E_a = 0$) and resuspension ($\tau_r = 0$), Equation 2.14 reduces to

$$\frac{dQ_s}{dx} = -\frac{Q_s}{L_s} L_s = \frac{Q_s}{\omega_r \rho_0}$$

(2.20)

so that sediment settles out over an $e$-folding length $L_s$, which we refer to as the settling length. The settling length is a function of the (constant, i.e., $E_a = 0$) discharge $Q_s$, and the grain settling velocity $\omega_r$.

In our modeling we make the low-slope approximation such that $S = -\frac{\eta g}{C \rho x} \ll 1$ and $\sin(\theta) = \tan(\theta) = S$. In the absence of entrainment ($E_a = 0$), and for low slopes, Equation 2.18 reduces to

$$S_f = \tan(\theta_f) \approx \sin(\theta_f) = \frac{\omega_r}{R g \sqrt{C_s \frac{\rho_s}{\rho}}}$$

(2.21)

We nondimensionalize our model by scaling the variables in Equations 2.1–2.19 as follows:

$$Q_s = Q_{s0} \tilde{Q}_s; \quad x = L_s \tilde{x}; \quad S = S_s \tilde{S};$$

$$\eta = H_s \tilde{H}_s; \quad t = T_s \tilde{T}_s; \quad H_s = S_s L_s; \quad T_s = (1 - \lambda) L_s H_s / Q_{s0};$$

$$E_a = \tilde{E}_a \tilde{E}_{s0}$$

where tildes denote scaled dimensionless variables. Sediment and water fluxes are scaled by their input values ($Q_{s0}$ and $Q_r$, respectively). We scale $\tilde{S}$ using the bypass slope $S_f$ (Equation 2.21) corresponding to the input sediment flux in the absence of fluid entrainment, which is a convenient choice inasmuch as $S_f$ combines grain size and sediment discharge into a single parameter. Streamwise distance is scaled by the settling length $L_s$, in the absence of resuspension (Equation 2.20). Bed elevation is scaled by a reference height $H_s$, which represents the change in elevation of a slope with gradient $S_f$ over a distance $L_s$. Time is scaled by a reference time scale $T_s$, which represents the time needed for the input sediment flux to fill the reference area $H_s L_s$. Finally, we scale the entrainment (“surface drag”) coefficient $E_a$ by the bottom drag coefficient $C_d$.

Substituting the above dimensionless variables into Equations 2.17, 2.13, 2.11, and 2.9 gives the following dimensionless system:

$$\frac{\tilde{\eta}}{\tilde{Q}_s} = \tilde{Q}_s \left[ 1 - \left( \tilde{Q}_s \frac{\tilde{S}}{1 + E_a} \right)^{2/3} \right]$$

(2.22)

$$\tilde{E}_s = \frac{1}{2} \left[ -1 + \tilde{b} \tilde{S} \right] + \sqrt{\left(1 + \tilde{b} \tilde{S}\right)^2 + 4 \tilde{b} \tilde{S}}$$

(2.23a–c)

$$\tilde{b} = \frac{b S_s C_s}{C_d}; \quad \tilde{a} = \frac{4 a S_s}{C_d}$$

$$\frac{d \tilde{Q}_s}{d \tilde{S}} = \left( \frac{\sqrt{C_s \frac{\omega_r}{\rho_s}}}{\rho_0 \omega_r} \right) \tilde{E}_s \left( \frac{\tilde{S}}{1 + E_a} \right)^{1/3}$$

(2.24)

In this work we use standard values for auxiliary parameters $a$ (= 0.00153), $b$ (= 0.0204), and $\rho_0$ (= 1.5), and for the most part we consider a single grain size ($\omega_{s0} = \omega_s$). The bed friction coefficient $C_d$ typically ranges between $10^{-3}$ and $10^{-2}$ for field-scale flows (Parker 2006).

In this dimensionless framework, we note that the scaled bypass slope (Equation 2.19) corresponding to the sediment supply $Q_{s0}$ is given by

$$\tilde{S}_s = \frac{S_s}{S_f} = 1 + \frac{a}{C_d (b + S_f C_d)} = 1 + \frac{\tilde{a}}{4 (\tilde{b} + \tilde{S}_f)}$$

(2.25)

Sensitivity of the Graded Slope

According to Equation 2.14, deposition relative to resuspension in the presence of fluid entrainment governs the mass balance of the system. Any point beneath the flow where deposition is balanced by resuspension is graded such that sediment bypasses that point. By this definition, grade may have only local significance, in contrast with the concept of grade commonly applied to turbidite channels at scales greater than a reach length (Kneller 2003).

For grade to occur at any given slope, a unique combination of discharge, grain size, and flow resistance must be realized. Each effect is isolated in Figure 2, where the graded bed slope angle for bypass both with and without entrainment are plotted as a function of sediment discharge over several orders of magnitude for a range of grain sizes. The slope angles plotted in Figure 2 depend on the critical grain settling velocity ($\omega_{s0}$) introduced in Equation 2.16 as a proxy for the critical shear velocity. In the following we assume for simplicity that this settling velocity is equal to the settling velocity in Equation 2.3 that sets the length scale (Equation 2.20). However, for flows with nonuniform grain size, such as the bipartite experiments reported below and most flows in nature, these velocities probably diverge. We do not pursue a model that tracks multiple grain sizes here, but for the purpose of scaling the magnitude of the graded slopes predicted by our model we adopt the convention that $\omega_{s0}$ should represent the coarsest ($\sim D_{10}$) fraction of the load (Komar 1985; Kneller 2003). Assuming fine to medium-fine sand characterizes the coarse tails of many flows in nature, the estimates in Figure 2 suggest that field-scale slopes graded by the mechanism we present rarely exceed $10^5$.

The plots in Figure 2 have a straightforward interpretation. Higher discharges increase shear at the bed, while finer grains are more easily resuspended. Both effects lower the graded slope for bypass. Flows partition the resistance to any given flow between bed friction and fluid entrainment. In turn, the bed friction controls the sediment resuspension rate. If bed friction is reduced, or fluid entrainment increases relative to bed friction, sediment resuspension decreases and the graded slope increases. Overall, the importance of entrainment diminishes at low slopes, so that for high sediment discharges (or high values of bed friction) the graded slopes with and without entrainment converge.

Model Setup

In a clinoform evolving from repeated flows, determining if and where the graded slope is reached, as well as the fate of bypassed sediment, requires us to model long-term strata formation. We use Equations 2.22–2.24 to numerically model deposition and bed evolution beneath sustained turbidity currents flowing into a basin over a ramp with fixed initial slope, $S_0$. The elevation at the top of the ramp sets the minimum water depth for deposition from the turbidity current. Above this level, we assume a uniform topset with slope $S_t (\ll S_0)$ is maintained during progradation which completely bypasses the sediment that sources the turbidity current. This zone can be generally thought of as a fluvial or wave-graded clinoform topset, with the dimensionless input sediment and water fluxes for the turbidity current set to unity where the topset meets the elevation of the ramp top. Implicit in this approach is a sediment supply adequate to aggrade the topset while maintaining a fixed initial turbidity-current discharge. Though we maintain this assumption for our modeling studies, we consider the effect of a temporally changing turbidity current discharge in the experiments discussed below.

The above setup, combined with fixed values for $a$, $b$, and $\rho_0$, leaves three independent parameters that must be specified: the friction slope $S_t$, the scaled initial ramp slope, $S_{t0}$, and the coefficient of bed friction $C_d$. The friction slope $S_t$ folds the initial discharge and characteristic grain size of flows into a single parameter (Equation 2.21). We are primarily interested in the depositional case, where the initial ramp slope $S_{t0}$ is less...
than the graded slope $S_g$ (Equation 2.25) corresponding to the unit sediment influx $Q_0 = 1$. Because the graded slope $S_g$ depends on the friction slope $S_f$ and the drag coefficient $C_d$ (Equation 2.25), in our simulations we specify the ramp slope $S_0$ as some fraction of the graded slope.

To illustrate, Figure 3 shows $S_s = S_f / S_f$ plotted as a function of $S_f$ for three values of $C_d$. If we specify a drag coefficient $C_d = 0.005$ and a friction slope $S_f = 1^\circ$, the relation plotted in Figure 3 gives a value for $S_s$ of 2. In other words, with entrainment the input sediment flux bypasses a slope of $2^\circ$, rather than $1^\circ$. So if we choose an initial ramp slope $S_0$ that is less than the value of $S_s$ (i.e., $< 2$), then deposition commences and the model flows build strata. Conversely, if we choose an initial ramp slope equal to or greater than 2, the flow simply bypasses the ramp and deposition does not occur.

Adopting this approach, we first present and explain four end-member cases of model-generated strata in the following section to illustrate general aspects of the model behavior. We then briefly consider the influence of the initial slope ($S_0$), bed resistance coefficient ($C_d$), and friction slope ($S_f$) in a series of sensitivity experiments to explore the range of model behavior.

**General Model Results**

Figure 4A–D shows strata modeled using the example values from the previous section, with depositional surfaces recorded at equal intervals of model run time. For simplicity, we set the topset slope ($S_u$) to zero. In this comparison, we consider the effect of resuspension and entrainment processes on stratal geometry by plotting the following four cases: (1) flows that are purely depositional and do not entrain ambient fluid (Fig. 4A); (2) flows that are purely depositional but still entrain ambient fluid (Fig. 4B); (3) flows that resuspend as they deposit but do not entrain ambient fluid (Fig. 4C); and (4) flows that both resuspend and entrain ambient fluid (Fig. 4D). The initial slope $S_0$ for the latter two runs was set at 25% below $S_u$, the slope required to bypass the sediment influx.

The first two cases recover the expected exponential bed profile documented by Pirmez et al. (1998) because deposition rate is directly proportional to flow concentration (i.e., Equation 2.20). The e-folding length (or settling length) is proportional to the ratio of fluid discharge to grain settling velocity (as in Equation 2.20), but note that in the second case $Q_g$ increases downstream due to entrainment. Clearly, the inclusion of entrainment does little to change the oblique clinoform geometry beyond a slight extension to the clinoform toe caused by an increase in the fluid flux.

**Figure 2**—Log-log plot of predicted graded slope that will bypass sediment according to Equation 2.19 both with ($S_g$—solid lines) and without ($S_f$—dashed lines) fluid entrainment. The slopes are predicted over a range of width-averaged sediment flux, $Q_s$, spanning four orders of magnitude. Each line corresponds to a unique Dietrich (1982) settling velocity setting the critical shear stress: the grain diameter interval between lines is 20 μm. The plots (A and B) differ only in the value of $C_d$, which is kept fixed for each over the range of discharges and grain sizes. The plots show four key predictions of the model: (1) for a given value of $C_d$, the graded slope both with and without entrainment decreases as $Q_s$ increases and $S_f$ decreases, (2) both decreasing $C_d$ and including fluid entrainment increases the graded slope for any combination of $Q_s$ and $S_f$, (3) increasing $Q_s$ causes the bypass slope to decrease more significantly for all $S_f$ when fluid entrainment is included, and (4) the difference in bypass slopes between flows with and without fluid entrainment diminishes as $C_d$ increases and $Q_s$ increases.

**Figure 3**—Plot showing the ratio of the graded slope, $S_g$, to the friction slope, $S_f$, as a function of $S_f$ from Figure 2 for three values of the drag coefficient $C_d$. The importance of fluid entrainment is seen to increase with increasing friction slope and decreasing drag coefficient.
length of settling. Overall entrainment is weak, because pure deposition quickly attenuates the flow downstream, so even increasing the initial slope causes only a modest extension of the clinoform toe without otherwise affecting the oblique geometry of the concave surfaces.

Because the clinoforms prograde onto a slope, the foresets gradually steepen (Pirmez et al. 1998). However, no mechanism in the purely depositional model prevents oversteepening, so in this case the foresets can grow to maximum slopes well beyond mechanical stability. In reality, for progradation to continue in Figure 4A and 4B, the foreset would have to be periodically cut back and cleared of sediment such that its slope never exceeded a failure threshold.

Figure 4C and 4D shows clinoforms with distinctly different morphologies generated from model runs with sediment resuspension included. Initial depositional profiles now show an overall sigmoidal morphology characterized by an upper, convex foreset and a lower, concave foreset (e.g., time \( t_1 \) in Figure 4D). In both cases, the flows begin strongly depositional but with enough resuspension near the inlet to produce a slightly convex upper foreset (as in Parker 2006). Continued deposition builds prograding clinoforms that gradually steepen with this geometry as they build out across the ramp (from time \( t_1 \) to time \( t_2 \)). Once the graded slope is reached on the clinoform foresets (time \( t_2 \)), sediment is swept basinward to the clinoform toes (time \( t_3 \)). The sharp reduction in slope there (\( \sim 25\% \)) triggers rapid deposition from the flows, generating lobes that begin to backstep on the foreset (i.e., onlap). Continued onlap eventually creates a new foreset, upon which the clinoform continues to prograde and steepen until the cycle is repeated. Hence, for the cases shown in Figure 4C and 4D, once a clinoform has prograded to the threshold steepness, progradation occurs only via cyclic steepening, bypass, and backfilling by turbidity currents.

The effect of fluid entrainment on the overall behavior can be seen between Figure 4C and 4D. Flow dilution adds to the effect of resuspension to reduce the spatial gradient in sediment fallout rate, resulting in more elongate foresets and extended lobes of bypassed sediment at the base-of-slope foreset toe (Fig. 4D). Still, in both cases onlapping of the foreset following an episode of sediment bypassing sustains progradation.

A notable observation from the differing cases seen in Figure 4 is the progradation rate of the upper clinoform. Pirmez et al. (1998) show that progradation rate should decrease as the water depth increases. Additionally, slope grading by turbidity currents in our model effectively redistributes clinoform strata basinward during episodes of bypassing. If the graded slope is less than other slope-setting processes, the basinward migration of the topset-foreset break is reduced relative to a purely depositional clinoform (e.g., Fig. 4B vs. 4D).

**General Model Behavior**

Key to understanding the model stratigraphy in Figure 4D (hereafter referred to as the general case) is the effect of resuspension, which introduces a slight sigmoidal shape to the clinoform as it progrades. Figure 5A and 5B show the downstream change in dimensionless sediment discharge (sediment lost from the flow) and bed slope, respectively, starting at the topset–foreset break for each of the first three time steps shown as depositional surfaces in Figure 4D (i.e., time \( t_1 \)–
At time $t_1$, deposition rates are high on the upper foreset and increase basinward to a maximum that is reflected in the convex growth of the foreset. As the uppermost foreset steepens, resuspension becomes important, revealed at time $t_2$ by reduced deposition rates overall on the upper foreset and an initial basinward decrease in the sediment fallout rate. Here, even though the slope increases basinward the overall gradient is insufficient for resuspension to fully offset deposition as the flow is diluted, so that the deposition rate again increases to a maximum just basinward of the maximum slope. At time $t_3$, however, the uppermost foreset slope is graded to the sediment influx. Because the slope increases basinward from this point, bypassing is sustained and deposition does not commence until the flow encounters much lower slopes—lower even than the initial ramp slope—at the point of onlap (Figs. 4D, 5B) that has migrated up from the base of the clinoform.

Backstepping of the base-of-slope sediment lobe can be explained by recognizing that flows bypassing the upper foreset first encounter a concave lower foreset region (time $t_2$ in Figures. 4D, 5B) with a slope that decreases sharply to the initial ramp slope. Here, the sediment fallout rate quickly increases to a maximum before decreasing again as the flow dies over the distal ramp. The resulting depositional lobe develops a convex upslope region (time $t_1$ in Figs. 4D, 5B) with a slope that is less than the initial ramp slope. Rapid deposition occurs at the slope break, which separates this growing lobe (time $t_3$ in Fig. 5A) from the graded zone upstream, causing the convex upslope boundary of the lobe to onlap the foreset. Importantly, onlap begins at the base or toe of the clinoform.

Our model does not explicitly treat the flow field associated with the turbidity current, and so the emergence of a slope-break effect is due solely to our representation of slope-dependent resuspension. More sophisticated modeling studies (Kostic and Parker 2006), experimental work (Garcia and Parker 1989), and field evidence (Sinclair and Tomasso 2002; Russell and Arnott 2003) have all argued for the potential role of hydraulic jumps in forcing a depositional signature at slope breaks under certain conditions. However, our model does give a first-order estimate of the magnitude of the slope reduction required to cause a hydraulic jump (Equation 2.10), and we present visual evidence below for a weak hydraulic jump at a break in slope observed in our experiments.

**Model Sensitivity Runs**

Altogether, the runs in Figure 4 show that by including resuspension our simple morphodynamic model can generate sigmoidal clinoform geometries that in turn modify the flow in such a way as to cause forest grading, sediment bypass, and lobe backstepping. In this section we assess model sensitivity by varying the initial slope, resistance coefficient, and friction slope for the general case (Fig. 4D).

The results for the general case suggest that if other input variables are left unchanged, changes in the ramp slope $S_0$ (specifically, changes in the ratio $S_0/S_1$) affect only the proportion of model run time during which the clinoform foreset is below grade. Figure 6A–D shows the general case for different initial slopes $S_0$. Increasing or decreasing the initial slope increases or decreases, respectively, the number of bypass cycles recorded in the strata over a given run time. In the case of an extremely low initial slope (Fig. 6D), the clinoform never reaches grade.

In addition to controlling the proportion of model strata being deposited below grade, the initial slope modifies the runout length for bypassed sediment at the base of the foreset. Higher initial slopes lead to a higher resuspension rate maintained beyond the clinoform toe, since increasing the initial slope reduces the magnitude of the slope break at the clinoform toe. For the same reason, if the initial slope is reduced then bypassed sediment is largely retained at the clinoform toe. These effects are more easily seen by plotting the downstream change in sediment discharge (again starting at the topset–foreset break) at times corresponding to sediment bypass in Figure 6A–C as shown in Figure 6E. We can see that the falloff in sediment discharge, starting from the point where deposition resumes downstream of the graded bypass zone, is reduced for higher initial slopes. Note also that in the high initial slope case a zone of bypass occurs well below the mid-foreset. This further emphasizes the importance of slope convexity; even with the flow losing

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**Fig. 5.**—A) Plot showing the downstream loss of dimensionless sediment flux from deposition at time slices corresponding to those shown in Figure 4D. The curves are plotted starting at the topset–foreset break (i.e., $S > 0$). The zone of bypass for surface $t_2$ is labeled. B) Plot of bed slope for the same three surfaces, $t_1$–$t_3$, from Figure 5A. The graded slope for the sediment influx is marked. The curves are plotted starting at the topset–foreset break (i.e., $S > 0$). C) Plot of three successive bed-slope profiles for the model run in Figure 4B showing the prograding and steepening concave settling profile. The curves are plotted starting at the topset–foreset break (i.e., $S > 0$). Unlike the profiles in Part B, the purely depositional flows maintain profile concavity as the foresets progressively steepen so that the slope always decreases downstream.
sediment in the uppermost foreset, the downslope increase in gradient leads to a graded zone that bypasses sediment basinward. Of importance here is that the initial slope not only determines the amount of progradational steepening required to reach grade but also determines the clinoform morphology and stacking pattern.

As seen in Equations 2.23 and 2.24, the values of the drag coefficient $C_d$ and friction slope $S_f$ not only set the graded slope but also control fluid entrainment. Our model predicts that reducing the bed drag and increasing the friction slope increases the influence of entrainment. However, recall that the two are not independent: $S_f$ depends on $C_d$ in
addition to the sediment flux and grain settling velocity. To isolate the role of the drag coefficient, we must adjust $S_T$ for changes in $C_d$ such that the remaining factors in Equation 2.21 remain constant (i.e., sediment flux and grain size), equivalent to varying $S_T$ as $1/\sqrt{C_d}$. Allowing for this, in Figure 7A and 7B we plot strata modeled with $C_d$ decreased and increased relative to the general case. Decreasing the drag coefficient generates thinner units that taper more slowly downslope and complete only one episode of bypassing. Increasing the drag coefficient generates multiple bypass cycles that build a more compact clinoform with highly localized deposition. A similar but less pronounced difference is observed when $S_T$ is varied with $C_T$ held fixed (Fig. 7C, D). As seen in Equation 2.21, high values of $S_T$ result from coarse grain sizes and small sediment discharges, whereas low values result from fine grain sizes and large sediment discharges.

These effects can be understood from Figure 7E, which shows the downstream evolution of dimensionless fluid discharge, starting at the topset–foreset break, for the time slice corresponding to the first depositional surface in all four cases (below grade). Fluid entrainment is highest in Figure 7A and 7D. The effect emerges directly from our estimate of $E_f$ (Equation 2.11), which increases with decreasing bed drag, higher friction slopes, and, consequently, a higher graded slope. A basinward decrease in the concentration of the flow resulting from higher entrainment increases the effective settling length; diminished entrainment tends to reduce the settling length.

Taken together, the sensitivity runs described in this section highlight several aspects of model behavior not seen in the general case. The value of the initial slope ($S_0$) relative to the graded bypass slope ($S$) controls the time required for the clinoform to reach grade, and determines the stacking and backstepping of bypassed sediment by controlling the drop in slope that occurs where the graded clinoform toe meets the lower basin slope. Our simplified momentum balance estimates fluid entrainment as a function of the bed resistance coefficient ($C_f$) and the friction slope ($S_f$). When varied, these change not only the graded slope for sediment bypassing but also the sediment fallout rate, modifying the extent of deposition basinward and thus strongly influencing clinoform morphology and the frequency of depositional cycles.

Overall, we recognize three key findings from the modeling experiments: (1) our simplified model of depositional turbidity currents can generate sedimentary clinoforms whose progradation rates and surface morphologies depend on initial basin slopes and the size and sediment load of the inflows; (2) our model provides a mechanism by which net-depositional turbidity currents can grade the slopes they build; and (3) the feedback between morphology and resuspension by the flow introduces a mechanism by which turbidity-current-graded clinoforms can prograde via repeated cycles of initial growth, sediment bypass, and depositional backstepping. In the sections that follow, we report findings from simple flume experiments that demonstrate general aspects of the model behavior. Because the experiments were deliberately kept simple to isolate the feedback between bed slopes and flume flows, detailed simulation using the numerical model is neither feasible nor desired. Instead, we use qualitative observations and basic measurements to support the main findings of our modeling study.

**FLUME EXPERIMENTS**

**Experimental Method**

We designed simple flume experiments to determine if weak turbidity currents at laboratory scale could build clinoforms to a graded slope and explore the sensitivity of that slope to sediment discharge and grain size. Experiments were conducted in a narrow glass-walled flume 0.15 m wide, 0.4 m high, and 12 m long (Fig. 8A). Turbidity-current discharges were mixed above the constant water level of the flume in a PVC channel with two industrial grades of sediment of mineral density 2.65 g cm$^{-3}$ (quartz), the sand fraction had a median grain diameter ($D_{50}$) of 104 μm, and the silt fraction one of 24 μm (Fig. 8B). The flume and the head tank used to generate the flows were both filled with 6–7 °C freshwater. Depending on sediment feed rates, the experiments were run for 40–180 min. In each experiment, a continuous turbidity current was introduced onto a ramp set at an angle (~ 18°) well below the expected range of bypass angles for laboratory-scale flows as suggested from our analysis (see Fig. 2) and measured in previous flume experiments where turbidity currents graded a delta foreset (Kostic et al. 2002). The use of a fine and coarse fraction allowed relatively rapid clinoform steepening from deposition of coarser sediment while flows were sustained over the length of the ramp by the suspended fines. Of the experiments reported here, half were performed with the distal end of the ramp suspended 5 cm above the flume. The remaining experiments were done with the ramp connected to the flume floor. Water and sediment feed were calibrated and either fixed or varied in set intervals over the course of an experiment, so that input sediment and water discharges were known. Either the initial concentration or the grain-size mix was varied between the experiments with fixed sediment and water inputs.

All experiments were video recorded. The final deposit was imaged in segments using a digital camera and mosaicked to produce dip sections of the strata. The strata were mapped by delineating depositional surfaces on the mosaics from the contrast between lighter silt and darker sand layers. Only those deposits laid down after a fluvial topset was in place were considered. The surfaces were correlated with observed events using video records. Slopes for each surface were measured from the mosaics.

**Results: Fixed-Discharge Experiments**

Six experiments performed with input sediment and water discharge held fixed produced the same general result (Fig. 9). Initially, the turbidity currents were mostly depositional, transporting only the finer fraction of the silt off the ramp (Fig. 9A). Sedimentation of sand and coarse silt did not extend beyond the upper part of the ramp and decreased down slope. This led to the development of an aggrading (indicated by thin arrow in Figure 9A) and then purely prograding clinoform with a foreset that steepened to grade (Fig. 9B). At this stage, the turbidity current completely bypassed the foreset (indicated by solid arrow in Figure 9B) and began depositing sand at the junction of the clinoform with the ramp. Continued deposition there led to the growth of a sandy convex lobe at the foreset base (Fig. 9C). The decrease in bed slope at the upstream boundary of the lobe slowed the flow such that it began depositing its coarser sediment load. In some cases, visual evidence suggested that the break in slope triggered a weak hydraulic jump (solid arrow in Figure 9C, inset). The gradient drop at the foreset–lobe boundary continued to force the turbidity current to deposit its coarse sediment load there (Fig. 9C, inset). This in turn moved the lobe upidip (indicated by dotted arrow in Figure 9C), shrinking the region of turbidity current bypassing while expanding the region of deposition from the bottom up. Upslope backstepping of the growing lobe eventually extended to the top of the old foreset, forming a new lower-angle foreset (Fig. 9D). Deposition of the coarse fraction in a basinward-thinning foreset then resumed, the cycle beginning again as the new foreset began to steepen towards the graded slope. Similar to the predictions of our model, the clinoform prograded across the ramp via these repeated episodes of foreset oversteepening and regrading.

Measurements of the graded slopes from the flume experiments are shown in Figure 10, differentiated by the weight % of sand ($D_{50}$ 104 μm) and plotted against the width-averaged sediment discharge into the flume. The graded slope that emerges from our model predicts that both grain size and sediment discharge should modify the slope at which bypassing is observed. A few comparisons from Figure 10
confirm this relationship. The plot shows that for nearly the same inflow sediment discharge, coarser mixtures produced higher bypass slopes (e.g., 20% sand = 25°, 24% sand = 27°, 44% sand = 27.5°). Conversely, for nearly the same grain-size mixture a flow with higher discharge produced a lower bypass slope (e.g., 20% sand = 25°, 21% sand = 29.7°). Note that for both low and high sand percentages, the range of bypass slopes is nearly 5°. We are not able to directly quantify how varying the sand percentage modifies the representative single grain size in our model—an issue we discuss further below—but our experiments show that raising the sand percentage does increase the graded slope as predicted for a higher
settling velocity (Fig. 2). The sediment discharge into the flume, however, is easier to measure, and also supports our expectation that higher discharges produce lower graded slopes.

**Results: Variable-Discharge Experiments**

The modeling and experimental results to this point have focused on autogenic behavior, i.e., variability generated only by feedbacks between the flow and the bed. However, in nature clinoforms are most certainly subject to allogenic variability, i.e., temporal changes in initial flow properties. Here we briefly consider how the slope-flow feedback governs clinoform morphology under the influence of allogenic variability in forcing.

We do this with an additional experiment in which sediment discharge was varied in three increments. For this run, the distal end of the initial ramp was suspended above the flume floor (Fig. 11D; see also Fig. 8A). Like the above example, the experiment began with initial water and sediment feed fixed so that cyclic oversteepening and regrading led to progradation of the clinoform toe to the suspended edge of the ramp (Fig. 11A). Sediment bypassing the graded slope at this stage (indicated by solid arrow in Figure 11A) was removed from the ramp entirely, halting progradation. At this point, the inlet sediment discharge was reduced by 50%. As predicted, this raised the graded slope, causing renewed steepening from storage of a sandy wedge in the upper foreset (indicated by thin arrow in Fig. 11B). Once the wedge had prograded and steepened to the increased graded slope, bypassing commenced again and the sediment inflow was swept beyond the ramp (as in Fig. 11A). Finally, the sediment discharge was raised to 75% of the original inflow. This caused erosion back to an intermediate graded slope (indicated by thin arrow in Fig. 11C), adding mobilized sand from the previously steepened foreset wedge to the bypassing inflow (indicated by solid arrow in Fig. 11C). An expanded view of the final graded slope that shows the distal edge of the ramp highlights the accumulating pile of sand sourced from bypassed and eroding flows (Fig. 11D).

The variable-discharge experiment demonstrates that our results are consistent with recent work showing that (1) a foreset slope above that required to bypass a turbidity current erodes back to that slope (Kostic et al. 2002), and (2) clearing of sediment from the toe of a clinoform can halt progradation (Paola et al. 2003). As Kostic et al. (2002) point out, the erosion of sand we observe on slopes elevated above that required for bypass is probably accomplished by a grain flow driven by the turbidity current. We do not attempt to capture this effect in our net-depositional model, but we acknowledge that it could be significant in environments where sandy slopes are reduced from erosion by turbidity currents.

**DISCUSSION**

**Scaling**

The experimental slopes measured above (25–30°) are much higher than most clinoform slopes in nature (typically < 10°). The discussion above outlines the controls on the graded slope, expressed in our model as a function of sediment discharge, grain settling velocity, and drag coefficient. We qualitatively expect that if the grain sizes used in the
experiments are reasonable for field-scale flows, but the sediment discharges are significantly reduced compared to field scale, then much higher graded slopes should be realized in the experiments. To interpret the experimental slopes quantitatively, we construct a plot similar to that shown in Figure 2 using Equation 2.19 and overlay the laboratory measurements presented in Figure 10 (Fig. 12). Following the suggestion of Parker et al. (1987), who show that higher bed drag coefficients result from increased effective roughness acting on thin flume flows, $C_d$ is increased by a factor of 10 from the typical field scale case. All of the experimental measurements fall within a range of graded slopes predicted by our model using grain settling velocities corresponding to sand ranging in size from 200 to 260 μm (calculated using the Dietrich (1982) equations). Referring to Figure 8B, this range of grain sizes is seen to agree reasonably well with the coarse tail ($D_{95}$-$D_{80}$) of the sand used in the experiments. While acknowledging the sensitivity of the predicted slope to grain settling velocity (i.e., $S_f \times \omega_2^2$), and the uncertainty surrounding the value of $C_d$, we conclude that even on laboratory slopes near their limit of mechanical stability the process of turbidity-current resuspension we propose can accomplish the observed grading.

We also expect that for a sediment load with similar grain sizes, a much lower fluid discharge in the laboratory results in a significantly shorter runout length for the bulk of the sediment in the flow. A lower bound on this distance could be quantified using the settling length $L_s$ (Equation 2.20), but this estimate omits the effects of resuspension and entrainment and does not account for multiple grain sizes. Given this uncertainty, and the aims of this study, we do not pursue a full dimensional scale-up for the experiments. However, two morphologic aspects of the experimental strata are worth consideration.

The experimental foresets showed little convexity in their long profiles except across the base-of-slope sediment lobes that onlapped a nearly linear graded slope (Figs. 9, 11). Additionally, sand bypassed to the base of the slope was deposited quickly, rarely extending more than 10 cm beyond the clinoform toe. Our sensitivity analysis highlights the independent effects of initial basin slope relative to the graded slope,
bed resistance coefficient, and the friction slope. In Figure 13A we present a model run using probable experimental values for each of these: a high bed resistance coefficient \((C_d = 0.05)\), a high friction slope \((S_f = 20')\), and an initial slope 60% below grade. Shown below this plot is an example photomosaic of laboratory strata following an experiment. With the elevation scaled by the slope, so that the model plot can be presented without vertical exaggeration, foreset morphology is seen to show only subtle convexity, in contrast to the backstepping sediment lobes that develop just beyond the clinoform toe. The high experimental slopes thus emphasize that even a modest curvature in the foreset slope can drive cyclic progradation.

The 1D setup of our model and the lateral confinement of the flume flow are an obvious limitation to extending our results to field scales. Though our width-averaged analysis cannot capture lateral lobe switching, compensational stacking, or incipient channelization—all likely to influence clinoforms built by turbidity currents—we argue that along-strike effects are not incompatible with the phenomenon in question here. Presumably, lateral divergence of flows slows the rate of progradational steepening and complicates the response of the flow to a break in slope. The backstepping turbidites documented here may be artificially enhanced due to our approach; indeed, the most compelling field evidence we know of for upslope-climbing turbidites is found in confined channel fills above a major break in slope (Fig. 14) (Pickering et al. 2001). While we maintain that the closest field analogues for our study might be found in fjords (Prior et al. 1987) or narrow reservoirs (Twichell et al. 2005) subject to frequent river-sourced turbidity currents (i.e., hyperpycnal inflows), we focus in the following section on the wider application of our results to unconfined settings along continental margins. Additionally, we acknowledge that our model does not include other slope-setting depositional or erosional processes (e.g., slope failure, internal waves, etc.) that can modify clinoform geometry. Because the goal of this study is to isolate a single mechanism and explore the degree

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**Fig. 10.**—Average graded slopes measured from the experimental strata plotted against the width-averaged volumetric sediment discharge \((Q_s = 2.65 \ g \ cm^{-3})\) into the flume. The experiments are differentiated according to the weight \% of sand in the sand–silt mixture.

**Fig. 11.**—Video frames of an experiment run with variable discharge: sand \% = 40, \(Q_{s0} = 18.7 \ mL \ s^{-1}\). Variation noted below is stated in terms of changing input sediment concentration by volume, \(C_0\). Grid cells on flume wall are 10 cm \(\times\) 10 cm. Note that camera view is slightly oblique to the flume wall. Refer to text for a more detailed description of the panel sequence. 

A) Sediment bypassing (solid arrow) of clinoform slope (dashed line) with \(C_0 = 6.6\%\). Note that the clinoform toe is at the ramp edge, so no base of slope lobe develops. Hence, progradation is halted. B) With \(C_0\) reduced to 3.2\%, storage of sand (thin arrow) commences on the upper foreset (dark wedge between dashed lines) until a new, higher bypass slope is reached (upper dashed line). C) \(C_0\) is increased to 5\%. Erosion commences (solid arrow) as the clinoform readjusts to a lower bypass slope (wedge shrinks between dashed lines). D) New bypass slope at \(C_0 = 5\%\) arrests progradation as all inflow sediment is swept off the suspended edge of the ramp (solid arrows).
to which it can explain experimental and field observations, it would be counterproductive to present a more complicated model with many interacting processes.

Morphologic Implications

Our focus in this study has been the sedimentary clinoform, reviewed in the introduction and defined here as a sedimentary body consisting of a region of maximum accumulation (foreset) bounded by a landward zone of minimal deposition (topset) and a basinward zone deprived of sediment (bottomset) (Fig. 1). Oblique clinoforms typically show nearly flat, low-sloping topsets that pass abruptly downdip to higher-sloping foresets with positive long-profile curvature (concave). Sigmoidal clinoforms, on the other hand, commonly show a low-sloping topset with negative long-profile curvature (convex) that gradually ”rolls over” across the foreset to an inflection point marking the transition to positive long-profile curvature (Sangree and Widmier 1977). In the most general sense, the topset is understood to be a region of elevated average bed shear stress and, consequently, low rates of net deposition (Pirmez et al. 1998). High bed shear stresses can be maintained either by wave-current stirring or residual fluvial momentum, both of which tend to decrease with increasing water depth. Consistent with this idea, but often overlooked, is the suggestion that subaqueous topsets may in fact be zones dominated by sediment gravity flows: in the former case, wave-current stresses sustain a turbid boundary layer that flows downslope under its own submerged weight (Wright et al. 2001), while in the latter case hyperpycnal flows sourced directly from rivers with high sediment loads keep sediment in suspension (Mulder and Syvitski 1995; Mulder et al. 2003).

The aforementioned model of Friedrichs and Wright (2004) demonstrates how the negative topset curvature of many subaqueous clinoforms develops from wave-current interaction that largely bypasses muddy river sediment basinward. A key component of their analysis is a critical topset slope (~ 0.5°) above which (or, equivalently, at basin depths below which) wave-supported turbidity flows no longer play an important role in determining clinoform morphology. Absent from this model is a view of how turbidity currents, no longer supported by wave agitation, can deliver the bulk of sediment to the clinoform foreset and control its long-profile curvature. By assuming a landward source and no wave influence, our model makes basic predictions for clinoform morphology that are not restricted by water depth. The sharp rollover of more oblique clinoforms may mark a rapid transition from wave-supported bypassing flows dominating the topset to purely depositional flows across the foreset (Fig. 4B). Alternatively, sigmoidal clinoforms showing a zone of increasing slope above an inflection point well below depths where wave stirring is important may be evidence for net-depositional turbidity currents sustaining a degree of resuspension, a point discussed at length in this paper and reported recently by Parker (2006). The ability of turbidity currents to resuspend a significant fraction of their sediment load as they deposit suggests that a clinoform dominated by hyperpycnal flows could also generate a topset zone that largely bypasses sediment basinward and exhibits a nearly flat, convex morphology. In this case, we expect the clinoform to develop a sigmoidal geometry basinward, since no abrupt process transition from wave-influenced transport would occur that could cause rapid deposition and a sharp rollover to positive profile concavity. A detailed discussion of how the interplay of wave-supported gravity flows and hyperpycnal flows generates distinctive morphologies is beyond the scope of this paper. Nevertheless, our model offers a more complete picture for how turbid underflows might control seabed morphology across the full reach of a subaqueous clinoform.

Stratigraphic Implications

Perhaps the most striking implication of our study is a distinct stratigraphic signature that originates from autogenic slope grading, bypass, and backfilling by turbidity currents. It is of considerable interest, then, to put some constraints on depositional environments where such cycles may occur and how their identification could influence an interpretation that might otherwise invoke allogenic causes.

As pointed out above, clinoform slopes must steepen to the graded slope set by the characteristic inflow and grain size before the progradation triggered by backstepping becomes active. Modern subaqueous delta-scale clinoforms strongly influenced by muddy sediment gravity flows that are actively prograding over low-gradient transgressive shelf deposits do show a tendency for gradual foreset steepening (Liu et al. 2002; Kuehl et al. 1997). Yet, overall the maximum slopes are still quite low (~ 10^{-3} degrees) (Fig. 15). Recall that when the initial slope was set to only 10% of that required for grading (Fig. 6D), the clinoform built by turbidity currents never reached the graded state. When slopes are well below grade so that resuspension is negligible, the rate of steepening depends only on the settling length (~ constant) and the basin depth. In this case, a low-gradient basin ramp—as in our configuration—can be thought of as a shallow shelf platform. For the sake of argument, we can estimate a length scale for the amount of progradation in Figure 6D using a reasonable settling length (Equation 2.20). We choose 500 m, corresponding to a flow 5 m in thickness moving at a nominal speed of 15 cm s^{-1} and carrying sediment with a fall velocity of 1 mm s^{-1}. The amount of progradation in Figure 6D thus equates to about 40 km without any sediment bypassing. Estimates of a time scale for this progradation would require specifying a typical concentration and intermittency for the flows (in order to compute an average bulk sediment supply), and we do not attempt this here. Instead, we simply suggest that in basins with low slopes and a modest sediment supply the time required for turbidity currents to build a graded slope may exceed intervals over which allogenic variables (e.g., sediment supply and eustasy) or other autogenic mechanisms (e.g., lateral lobe switching) modify clinoform evolution.

For this reason, we speculate that clinoforms prograding onto preexisting slopes steeper than most continental shelves are more likely to show accretional signatures of cyclic progradation by turbidity.
currents. By definition, shelf-edge clinoforms are progradational wedges that accrete out into deepwater basins as a growth of the shelf margin, typically during lowstands in relative sea level (Suter and Berryhill 1985; Posamentier et al. 1992), but also during sea-level highstands when sediment supply is very high (Carvajal and Steel 2006). Shelf-edge clinoforms commonly show relatively steep (3–5°) foresets and, where studied in outcrop or core, contain turbidite-prone facies assemblages that are commonly sandy due to a direct connection with a fluvial feeder system (Porębski and Steel 2003). Hyperpycnal river flows that generate turbidity currents are commonly assumed to dominate strata formation in these settings (Petter and Steel 2006). If turbidity-current deposits are the primary building blocks of clinoforms in these settings, and the mechanism

![Diagram](image-url)

**Fig. 13.**—A) Model run meant to approximate experimental scale strata. $S_f = 20°$, $C_J = 0.05$, and the initial slope is set 60% below that required to bypass the influx. Here, the friction slope scales nondimensional bed elevation so the plot can be viewed at true scale (no vertical exaggeration). B) Example photomosaic of flume strata following an experiment with key surfaces denoted that are outlined in the sequence of Figure 9 and discussed in the text.
we document is active due to higher basin slopes, then distinct stratigraphic patterns should be recognizable in the fine-scale stratigraphic architecture.

A pertinent example occurs in exposures of Eocene shelf-edge-to-deepwater clinoforms in Spitsbergen (Plink-Björklund et al. 2001; Mellere et al. 2002). Here, seismic-scale exposure of a stack of decameter-scale regressive clinoforms 60–80 m thick has allowed reconstruction of their undeformed sigmoidal geometry in combination with detailed facies descriptions (Fig. 16A–C). The clinoforms developed and prograded via a repeating pattern of downlapping foreset lobes that are capped by zones containing coarser-grained slumped and collapsed horizons, usually associated with shallow and narrow sandy channels (chutes) (Fig. 16C). These progradational lobe packages, some 10 m thick on average, appear to reflect a range of styles of slope construction that include (1) lobe progradation and downlap (characterized by upward-coarsening trends within heterolithic turbidite facies in Figure 16C), (2) channeling associated with slope instability, slumping, and sediment bypass, and (3) clinoform zones characterized by upward fining and thinning of turbidites (packages R5-R7 in profiles 5–8, Figure 16C), possibly reflecting oblique or even upslope shifting of slope lobes. The overall progradational pattern seen in Figure 16C is clearly cyclic or punctuated, analogous to the turbidite slopes produced experimentally. At some locations depositional lobes can be seen to wedge into each other in a complex manner (Fig. 16D), possibly one of the signatures of lobe backstepping or sidestepping on the slope. Examples like this appear to support our model for prograding turbiditic slopes, albeit modified to account for shallow incision and channelization. This is not surprising, since turbidity currents fed by variable river discharges moving across a slope at or near grade should at times be expected to develop some degree of confinement. In these settings, a more complete picture of slope growth, bypassing, and backfilling might emerge from a model that produces incision of graded slopes, consequent flow localization, and a more variable base-of-slope signature. Still, even in the absence of demonstrable evidence from the stratigraphic record for the full cyclic sequence predicted by our model, a slope graded by turbidity currents is still likely to be realized and, importantly, marks a significant shift in the depositional pattern across the clinoform.
CONCLUSIONS

In this paper we have proposed a model for turbidity-current deposition from which a fundamental feedback emerges between the balance of deposition and resuspension in the flow and evolution of the bed profile. Our 1D model predicts a sequence in which net-depositional turbidity currents build sigmoidal clinoforms to threshold slopes the flows bypass, delivering sediment to reduced slopes downstream. In the 1D case, rapid deposition in the lower sloping region forces slope backfilling, allowing the system to prograde. The sigmoidal geometry depends on the ability of the flow to resuspend sediment, so that in the limit of vanishing resuspension an oblique clinoform with a purely concave profile is predicted. The model provides a direct predictor for the graded slope that

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**Fig. 16.**—A) Photographic panel of exposed Eocene shelf-edge clinoforms in Spitsbergen taken from Figure 4 of Mellere et al. (2002). B) Wide exposure of sigmoidal clinoforms allows approximate long profiles to be measured. C) Prograding turbidite lobe complexes within the lowermost thick slope clinoform of Figure 16B. Note the cyclic arrangement in the repeated lobes of thin-bedded turbidites, sometimes capped by shallow channels or chutes (from Mellere et al. 2002). D) Complex pattern of lobe wedging, possibly implying sidestepping or backstepping on the slope at times. All beds are turbidites. Located on the middle–lower slope of clinoform complex shown in Part C.
depends on the flow discharge and grain size. Flume experiments demonstrate the feedback and provide reasonable agreement with the model. In considering appropriate field analogues, we emphasize the implications of our study for both clinoform morphology and stratigraphic architecture. We argue that our model provides a necessary extension to models of equilibrium clinoforms resulting from wave-supported gravity flows by showing how their morphology can evolve basinward from turbidity-current deposition that drives clinoform progradation. Stratigraphic evidence for steepening to grade and autogenic bypassing may be restricted to shelf-edge clinoforms prograding over steep basin slopes with relatively high sediment supply. Whether the full autogenic sequence presented here occurs in these settings is an open question, but repeating packages of forest accretion and channelization are suggestive. Overall, we hope that our study broadens conventional understanding for how net-depositional turbidity currents generate distinctive clinoform morphologies and subsequence-scale stratigraphic heterogeneity.

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REFERENCES


SYMBOLS AND NOTATION

\( a \) empirical constant, fluid entrainment relation
\( b \) empirical constant, fluid entrainment relation
\( C \) layer-averaged volumetric sediment concentration
\( C_0 \) layer-averaged volumetric sediment concentration at inflow boundary
\( C_d \) bottom drag coefficient
\( E_w \) fluid entrainment coefficient
\( F_s \) resuspension flux of sediment at the bed
\( F_{\text{Fr}} \) densimetric Froude number
\( H \) flow thickness
\( H_r \) reference flow thickness
\( L_s \) grain settling length in the absence of resuspension
\( Q_s \) volumetric sediment flux per unit width
\( Q_{s0} \) volumetric sediment flux per unit width at inflow boundary
\( Q_w \) volumetric water flux per unit width
\( Q_{w0} \) volumetric water flux per unit width at inflow boundary
\( R_g \) submerged specific weight of sediment
\( R_{b0} \) bulk Richardson number
\( r_0 \) near-bed concentration parameter
\( S \) bed slope
\( S_c \) graded bed slope
\( S_f \) friction slope = graded slope in the absence of fluid entrainment
\( S_t \) clinoform topset slope
\( T \) reference time scale
\( t \) time
\( U \) layer-averaged flow velocity
\( x \) streamwise distance
\( \eta \) bed elevation
\( \lambda \) bed porosity
\( \rho \) density of ambient fluid
\( \rho_s \) sediment grain density
\( \tau_b \) bed shear stress
\( \tau_c \) critical bed shear stress for full suspension
\( \theta \) bed slope angle \( = \tan^{-1}(S) \)
\( \theta_c \) graded slope angle \( = \tan^{-1}(S_c) \)
\( \theta_f \) friction bed slope angle \( = \tan^{-1}(S_f) \)
\( \omega_s \) grain settling velocity
\( \omega_{sc} \) critical settling velocity for sediment resuspension
\( \tilde{\eta}, \tilde{Q}_s, \tilde{F}_s, \tilde{E}_w \) Scaled, dimensionless variables
\( \tilde{Q}_s, \tilde{\tau}, \tilde{S}, \tilde{T} \)