Do hyperpycnal-flow deposits record river-flood dynamics?

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ABSTRACT

Hyperpycnal flows are turbid river plumes that can plunge to form turbidity currents where they enter a water body of lesser density. Because these flows provide one of the most direct connections between terrestrial sediment sources and marine depositional sinks, their deposits might preserve an important record across a variety of climatic and tectonic settings. A leading hypothesis assumes that hyperpycnal-flow velocity scales directly with river discharge, such that individual turbidites record the rising and falling discharge of a flooding river. We tested this hypothesis using a one-dimensional flow model and found that turbid river flow must move through a backwater zone, depth-limited plume, and plunging zone before becoming a turbidity current. These zones can extend tens of kilometers offshore and significantly affect the transfer of momentum from river to turbidity current. Counter to the proposed hypothesis, our results suggest that expected bed forms and sediment grading patterns in hyperpycnal-flow deposits can record multiple flow accelerations and decelerations even during a simple single-peaked flood. This occurs because of spatial acceleration and deceleration within the three transitional zones and because their boundaries move in response to flow discharge and suspended sediment concentration. Results also suggest that the criterion used to identify plunging hyperpycnal flows (a flow density in excess of the ambient fluid) is a necessary, but not sufficient condition. The basin also must be deep enough, in some cases greater than tens of meters, in order for the plume to collapse and form a turbidity current.

INTRODUCTION

Hyperpycnal river plumes are one of the most direct routes by which fluvial sediment sources can be linked to marine depositional basins. Hyperpycnal plumes occur where turbid river waters are denser than an ambient lake or ocean, allowing them to plunge and runout as turbidity currents (e.g., Mulder et al., 2003). Because these flows are generated directly from a riverine source, their deposits are a potential record of river flood characteristics and dynamics, and they might contain vital clues linking terrestrial landscape and marine stratigraphic evolution to tectonics and climate change.

Although common in freshwater reservoirs, hyperpycnal plumes are relatively rare in marine basins because the river must be sufficiently charged with sediment to overcome the density of saltwater. River plumes with a volume concentration of suspended siliciclastic grains >1.5% (or ~40 g/L) are thought to be capable of plunging behavior, assuming density equivalence and a seawater density of $\rho_a = 1025 \text{ kg/m}^3$, although some have suggested much lower thresholds due to mixing (Parsons et al., 2001; Felix et al., 2006). Using a rating-curve analysis, Mulder and Syvitski (1995) showed that many small- and medium-sized rivers (average annual discharge <~300 m³/s) can exceed this threshold on centennial or shorter time scales. The Huang He River, China, for example, is probably the world's largest river capable of producing hyperpycnal flows annually, where sediment concentrations have exceeded 8% during floods (Wright et al., 1988; van Gelder et al., 1994).

Few direct observations are available that link marine hyperpycnalflow deposits (hyperpycnites) to particular river floods. In their review,

Mulder et al. (2003) summarized the intuitive hypothesis that hyperpycnites accurately record the time evolution of a flooding river. In this model, hyperpycnal flows initially accelerate to the flood peak and then decelerate, resulting in a sediment deposit that is reverse to normally graded and that contains sedimentary structures indicating waxing to waning flow (Fig. 1). These deposits are different from the classic Bouma Sequence for turbidity currents generated from mass failures, which are unlikely to preserve an initial accelerating phase of the flow (e.g., Lamb et al., 2008), although turbidity currents triggered by retrogressive sliding (van den Berg et al., 2002), reflections due to topography (e.g., Lamb et al., 2004) and other mechanisms (e.g., Best et al., 2005) might also explain beds preserving the signals of waxing and waning flow. The hypothesis that hyperpycnites record the time evolution of a flooding river is powerful because, if correct, it can be used to reconstruct river processes from modern (e.g., Mulder et al., 2001a, 2001b) and ancient (e.g., Lamb et al., 2008; Myrow et al., 2008) turbidites. This model, however, implicitly assumes that hyperpycnal-flow velocity scales directly with river discharge, an assumption that has not yet been tested.

In this study, we use a simple numerical model for a plunging hyperpycnal river plume to test the hypothesis that hyperpycnal flows accurately reflect the time evolution of a flooding river. We investigate the transitions that must take place for normal river flow to evolve into a turbidity current, including a fluvial backwater zone, a depth-limited river plume, and a plunging river plume (Fig. 2). We specifically focus on the transfer of momentum across these zones and movement of their boundaries



Figure 1. A: Hypothetical flood hydrograph that generates a hyperpycnal river plume; after Mulder et al. (2003). Once critical concentration is exceeded, turbidity current is expected to mirror rising and falling river discharge, producing an inverse to normally graded bed with sedimentary structures indicating waxing to waning flow. B: Schematic drawing of an event bed. C: Photograph of an event bed from the Pennsylvanian Minturn Formation, Colorado, interpreted to be from a hyperpycnal plume (Lamb et al., 2008; Myrow et al., 2008), showing initial deposition of ripples (r) in very fine sand (vfs), then higher-energy plane bed (p) in fine sand (fs), and finally ripples (r) in very fine sand. Pencil for scale is 14 cm long.

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Figure 2. Cartoon illustrating five zones of flow from river to turbidity current, as well as locations of backwater point, shoreline, and plunge point.

as a function of river discharge and sediment concentration. Although sometimes neglected in models for hyperpychal flows (e.g., Mulder et al., 1998), these transitional zones are found to be important filters on the transfer of momentum, which, in some cases, produce anticorrelations between hyperpychal-flow velocity and river discharge.

MODEL DEVELOPMENT

In the most upstream zone, the river is approximated as normal flow where the water-surface slope (S_w) parallels the bed slope (S_b) (Fig. 2). Further downstream, in the backwater zone, flow diverges as the watersurface slope tends to horizontal because it is affected by the stagnant water beyond the shoreline. In most cases, a hyperpychal river plume will not plunge at the shoreline because a specific depth must be attained for the plume to be unstable depending on the discharge and sediment concentration (Akiyama and Stefan, 1984). We refer to the flow in between the shoreline and the plunge point as a depth-limited plume. At the plunge point, the turbid flow is unstable, and because it is denser than the neighboring ambient fluid, it collapses in the plunge zone. Downstream of the plunge zone, the flow is by definition a turbidity current—a bottom-hugging density flow driven by the submerged weight of suspended sediment.

Our goal is to use the simplest possible model that accurately captures the transfer of momentum from normal river flow to turbidity current. We model flow using one-dimensional (1-D), depth-averaged, steady-state conservation equations for mass and momentum (Henderson, 1966), assuming a constant friction factor (see the GSA Data Repository¹). Because the model is 1-D, mixing of ambient fluid and lateral spreading of the depth-limited plume are neglected. Following theoretical and experimental work on plunging behavior in reservoirs and lakes (Akiyama and Stefan, 1984; Parker and Toniolo, 2007), the plunge point is defined by a constant densimetric Froude number of $F_d \sim 0.5$ (Lee and Yu, 1997):

$$F_{\rm d} = \frac{q}{\sqrt{\Delta \rho g h^3 / \rho_{\rm a}}}$$

where q is the discharge per unit width, q = Uh, U is the velocity averaged over the depth h, $\Delta \rho$ is density of the current in excess of the ambient fluid density ρ_a , and g is the acceleration due to gravity. The same workers also determined that the fully collapsed turbidity current has $F_d = 1$, a thickness $h_d \approx 0.75h_p$, where h_p is the flow depth at the plunge point, and that the collapse occurs over a length $L_p \approx 10h_p$. These values are used to determine the location and size of the plunge zone. The values of flow depth, velocity, and sediment concentration of the collapsed plume should serve as the boundary conditions for a turbidity current model to route flow and sediment further seaward (e.g., Parker, 1982). Because our emphasis is on the transitional regions, we assume steady and uniform flow in the turbidity current zone for simplicity. In all zones, erosion and deposition of sediment are neglected. Deposition rates are often small enough that they do not significantly impact the dynamics of turbidity currents, especially for plumes dominated by mud.

Although more sophisticated three-dimensional (3-D) models are available for plunging flows (e.g., Kassem and Imran, 2001), our approach is an appropriate first-order test for the hypothesis that hyperpycnal-flow velocity scales directly with river discharge. Even with this simple model, we find that significant filtering occurs as momentum is transferred from river to turbidity current.

MODEL RESULTS

Model results are shown for a hypothetical river scaled roughly after the Huang He River because the occurrence of hyperpycnal flows is well documented there (Wright et al., 1998). Following Wright et al. (1998) and van Gelder et al. (1994), the bed gradients are 5×10^{-4} , 3×10^{-3} , and 6×10^{-4} for the fluvial reach (x < 2 km), delta front (2 < x < 4 km), and the delta rise (x > 4 km) (Fig. 3A). The water surface elevation is fixed at sea level at the downstream end of the modeling domain (x = 60 km), which results in a water depth during low flow of $h_s = 4$ m at the shoreline (x =0). For the first three model runs (T1 – T3), sediment concentration (c_0) was set to be just above the critical concentration necessary for plunging ($c_0 = 1.7\%$ and $\rho_a = 1025$ kg/m³) and the discharge (q_0) was varied from



Figure 3. Model results of (A) flow elevation and (B) depth-averaged flow velocity as a function of distance from shoreline for four model runs (T1–T4). C: Model runs of local flow velocity at five distances from shoreline as function of input fluvial discharge for same model setup as in A and B with $c_0 = 1.7\%$. D: Hypothetical hyperpycnal plume deposit resulting from a simple flood hydrograph (e.g., Fig. 1A) within region of anticorrelation (e.g., x = 30 km in C) showing multiple stacked units of ripple and parallel lamination and three cycles of reverse to normal grading.

¹GSA Data Repository item 2009268, model description, steady state assumption, and velocity-discharge scaling, is available online at www.geosociety. org/pubs/ft2009.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

the measured average annual discharge $(q_0 = 4 \text{ m}^2/\text{s}, \text{ assuming a constant}$ channel width of w = 400 m) to the largest discharge on record $(q_0 = 22 \text{ m}^2/\text{s})$. Note that the maximum possible discharge on the Huang He over geologic time could be as large as 140 m²/s (Mulder and Syvitski, 1995). For the fourth run, the discharge was held constant at $q_0 = 22 \text{ m}^2/\text{s}$, and sediment concentration was increased to the maximum observed value of $c_0 = 8.4\%$. Each profile in Figure 3A is at steady state and can be considered to be either different flood events or stages within a single event because the time scale for the plunge point to move (approximately hours) is small compared to a typical flood duration (approximately days) (see the Data Repository).

For any given profile, the flow velocity in the normal-flow zone is uniform, as expected (Fig. 3B). Velocity decreases in the backwater zone as the water surface slope and bed slope diverge. In the depth-limited plume, the flow decelerates rapidly due to the divergent bed and water-surface slopes. The plume plunges at $F_d = 0.5$ and accelerates throughout the plunge zone until the condition $F_d = 1$ is achieved.

The model indicates that velocity in the normal-flow zone and the turbidity-current zone scale with discharge to the one-third power (see the Data Repository). Thus, even though the magnitude of velocity in the turbidity current is smaller than in the river due to the reduced driving force of the submerged current, an increase in discharge should produce an equivalent relative response in velocity in both zones. The intervening velocity profiles are more complex than this, however, because the locations of the transition zones are also a function of discharge.

The maximum length of the backwater zone scales with flow depth at the shoreline (h_s) divided by the channel-bed slope (S_b) . Thus, for the Huang He $(h_s \approx 4 \text{ m}; S_b = 5 \times 10^{-4})$, the backwater zone can extend tens of kilometers upstream of the shoreline (Fig. 3A). However, for high discharge events, $S_w = S_b$, the backwater point is pushed toward the shoreline. This results in an amplified change in flow velocity near the shoreline as compared to the normal-flow zone further upstream. For example, comparing T1 to T3, the 5.5-fold increase in discharge causes velocity in the normal flow zone (x < -8 km) to increase by a factor of 1.8 and velocity at the shoreline to increase by a factor of 4.7.

An increase in discharge also pushes the plunge point seaward. Given $F_{d} = 0.5$, the depth required for plunging is a function of discharge and sediment concentration at the plunge point (Fig. 4A). The plungepoint depth can be as large as 40 m for the measured discharge of 22 m²/s on the Huang He River. Because the gradient of the delta rise is small, the plunge point moves ~50 km offshore to achieve the necessary depth. The translation of the plunge point results in flow velocity being anticorrelated with river discharge at some locations on the bed. For example, for the region 18 < x < 23 km, the flow velocity decreases as discharge increases from T1 to T2, and in the region 37 < x < 49 km, flow velocity decreases as discharge increases from T2 to T3. This can be seen more clearly in Figure 3C, where the relationship between local flow velocity and inlet discharge is anticorrelated depending on the discharge magnitude and the distance offshore. This anticorrelation occurs because the plunge point moves seaward with increasing discharge, and flow velocity in the turbidity current is faster than that in the depth-limited plume near the plunge point.

Higher suspended sediment concentrations cause the current to become unstable at shallower depths, thus moving the plunge point landward (Fig. 3A). In profile T4, the sediment concentration is set to the maximum on record: $c_0 = 8.5\%$. The result is translation of the plunge point from its location at T3 landward by 47 km to within 3 km of the shore-line. The increase in sediment concentration from T3 to T4 also causes a significant increase in turbidity current velocity from 0.9 m/s to 2.9 m/s due to the greater flow density (Fig. 3B). This implies that turbidity current velocity can be correlated, anticorrelated, or even uncorrelated with river discharge, depending on how sediment concentration and discharge covary during the course of a flood (Fig. 4B).



Figure 4. Contours of predicted (A) plunge point depth (h_p) in meters and (B) depth-averaged velocity of a turbidity current after plunging (m/s) as function of inlet sediment concentration (c_0) and flow discharge per unit width (q_0). In both cases, critical concentration required to plunge is set to 1.5%.

The transfer of momentum between the transitional zones of flow and the translation of their boundaries can lead to varying depositional signatures. Although our interest is to link flow dynamics with deposits, herein this is done only by assuming that expected depositional sediment sizes and bed forms track local depth-averaged velocity. At low flow, the backwater zone is a region of divergent flow and is likely depositional. However, during high discharge, the velocity can increase substantially due to movement of the backwater zone, which could result in rapid evacuation of sediment further charging the turbid plume. Seaward of the plunge zone, turbidity current velocity is expected to scale with the river discharge to the one-third power (for constant sediment concentration). Thus, an increase in river discharge should result in coarsening of the bed or higher-energy bed forms in deep water. For the example shown (T1-T3, Fig. 3B), the velocity in the turbidity current zone (x > 50 km) increases with the increasing river discharge, although the absolute magnitude of the response is muted in comparison to the river. Nonetheless, this change in velocity (from ~0.5 to 0.9 m/s) should result in ripples at T1 that wash out to the upper plane bed at T3 for most sand sizes (Southard, 1991). However, such bed form changes can also be accomplished by increasing the sediment concentration independent of discharge, as shown by T4, where turbidity current velocities of ~2.9 m/s are well into the upper-plane-bed stability field.

The depositional patterns between the shoreline and the maximum extent of the plunge point (e.g., 0 < x < 50 km in Fig. 3A) are likely to be complex. Because velocity can be anticorrelated with discharge,

depositional sequences over the course of a river flood (e.g., Fig. 1A) might contain multiple stacked inverse and normally graded units and bed forms, such as alternating stacked rippled and parallel laminations, indicating pulsating flow within a single event bed (Fig. 3D). Such deposits have been interpreted previously to be the result of a flood with multiple peaks or turbidity current reflections due to complex topography (e.g., Lamb et al., 2008; Myrow et al., 2008). Our results suggest that multiple accelerations and decelerations can occur due to the intrinsic phenomenon of plunge point translation for even the most simple flood hydrograph and bed topography.

DISCUSSION

The depth-limited plume extends tens of kilometers offshore in the example shown (Fig. 3). This is in part because of the low slope of the Huang He shelf. The distance between the plunge point and the shoreline is $\sim h_p/S_b$ and, all other things equal, the plunge point would be closer to shore on a steeper slope (Fig. 4). Our model assumes that there is no lateral spreading within the depth-limited plume, which makes it most accurate for plunging within a fjord or submarine canyon (e.g., Mulder et al., 1998). A lateral spreading plume should plunge closer to shore, and this depends on the width of the river mouth, spreading angle, bed topography, and mixing across the spreading front (e.g., Johnson et al., 1987; Kassem et al., 2003). The maximum extent of the plunge point also can be diminished if sediment concentration covaries with discharge (e.g., Figs. 3A and 4A), as it often does in flooding rivers. This relationship can be complex, and it depends on the sediment supply, grain size, and history of flooding (e.g., Topping et al., 2000).

Our model suggests that depth-limited hyperpycnal plumes might be more important in the geologic record than previously recognized. The depth needed to plunge (h_p) can be large (tens of meters) when the sediment concentration is just above critical or when discharge (per unit width) is large (Fig. 4A). Our model suggests an important end-member case where a hyperpycnal plume might never plunge if h_p exceeds the maximum depth in the ocean or lake basin. If this occurs, the depth-limited plume will advance across the basin until it dissipates (or becomes hypopycnal) due to deposition. Thus, a plume density in excess of the ambient fluid density is a necessary, but not sufficient, condition to form a turbidity current. In addition, the basin must be deep enough for the current to become unstable.

CONCLUSIONS

Where hyperpycnal rivers enter oceans or lakes of sufficient depth, the flow moves through a backwater zone, a depth-limited plume, and a plunging plume, and this significantly affects the transfer of momentum from river to turbidity current. Importantly, the boundaries of these zones shift during the course of a river flood, resulting in a potentially large regime, in space and time, where local hyperpycnal plume velocities are anticorrelated with river discharge. The result is a complex depositional signature including multiple stacked inverse and normally graded units, in some cases, even from a simple single-peaked hydrograph.

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