THE INFLUENCE OF SUBTLE GRADIENT CHANGES ON DEEP-WATER GRAVITY FLOWS: A CASE STUDY FROM THE MOROCCAN TURBIDITE SYSTEM

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ABSTRACT: The Moroccan Turbidite System is unique in that individual gravity-flow deposits can be correlated across distances of several hundred kilometers, both within and between depositional basins. An extensive dataset of shallow sediment cores is analyzed here, in order to investigate the influence of gradient changes on individual siliciclastic gravity flows passing through this system in the last 160,000 years. The largest flows (deposit volumes > 100 km$^3$) are capable of travelling for more than 1000 km across slopes of less than 0.1°. The deposits of these flows display significant lateral heterogeneity as a consequence of changes in seafloor gradient. Increases in gradient can lead to sediment bypass and/or erosion, and unconfined flows may become channelized. Decreases in gradient can lead to significant changes in sand–mud ratio and the deposition of thick mud caps, while small-volume flow deposits may pinch out completely. One of the largest flows shows evidence for multiple transformations as it crossed the Agadir Basin, with the resulting deposits switching laterally from (1) a gravel lag and cut-and-fill scours (representing bypass and erosion across a slope of 0.05°), to (2) a thick linked turbidite–debrite bed containing a muddy sand debrite (in response to a decrease in slope to < 0.01°), to (3) a normally graded turbidite (following a subtle increase in slope to 0.02°). Although the changes in slope angle described here appear remarkably subtle, the relative changes in slope are significant, and clearly exert a major control on flow behavior. Such variations in slope would not be detectable in outcrop or subsurface sequences, yet will generate significant complexity in deep-water reservoirs.

KEYWORDS: Turbidity current, gravity flow, channel, deep-water, slope, gradient, Morocco

INTRODUCTION

The processes and deposits of deep-water gravity flows, e.g., turbidity currents and debris flows, are influenced by a wide variety of controlling factors, both allocyclic and autocyclic. The latter category includes seafloor gradient, which affects the ability of gravity flows to erode, transport, and deposit sediment. This has been investigated through laboratory experiments (e.g., Garcia and Parker, 1989; Mulder and Alexander, 2001) and studies of modern sedimentary systems (e.g., Ercilla et al., 2002; Wynn et al., 2002a). Recent work, largely driven by the needs of the hydrocarbons industry, has highlighted the influence of seafloor gradient on turbidity currents passing through deep-water channels, particularly in complex slope environments that are modified by diapirism or tectonism (e.g., Friedmann et al., 2000; Pirmez et al., 2000; Adeogba et al., 2003, 2005; Ferry et al., 2005; Wynn et al., 2007). However, the absence of a spatially extensive bed-correlation framework in these examples means that the influence of gradient on individual flows is poorly understood.

Here, we present new data documenting the influence of seafloor gradient on individual submarine gravity flows in the Moroccan Turbidite System, which is located on the Atlantic margin of northwest Africa (Fig. 1; Wynn et al., 2000, 2002b). During the late Quaternary, the Moroccan Turbidite System has hosted some of the largest gravity flows in the world, with individual flow deposits exceeding 100 km$^3$ and flow run-out distances reaching 1500 km (Wynn et al., 2002b; Talling et al., 2007; Frenz et al., 2009). The huge scale of these flows was fully realized only after a detailed chronostratigraphic framework was established throughout the Moroccan Turbidite System, linking deposits in adjacent depositional basins (Wynn et al., 2002b). This contribution specifically focuses on siliciclastic gravity flows derived from the Moroccan continental margin. These flows passed through the 450-km-long Agadir Canyon before reaching the Agadir Basin. The largest flows then continued westwards across the basin floor, and a small number subsequently overspilled into the 600-km-long Madeira Distributary Channel System (referred to herein as “Madeira Channels”) before depositing their remaining sediment load on the enclosed Madeira Abyssal Plain (Fig. 1). In addition, some flows overspilled the sedimentary ridge in the northeast Agadir Basin, and continued onto the Seine Abyssal Plain (Wynn et al., 2002b).
CORE DATASET AND BED CORRELATIVE FRAMEWORK

The Moroccan Turbidite System is unique in that a comprehensive chronostratigraphic framework has allowed individual flow deposits (beds) to be correlated over distances > 1000 km between the Agadir Basin and the Madeira Abyssal Plain (Wynn et al., 2002b). The primary dataset is a series of shallow piston cores (penetration < 15 m) collected during numerous research cruises to the study area. Core data collected prior to 2002 are summarized in Wynn et al. (2002b). More recent core data, mostly from the Agadir Basin, were collected during research cruises on RRS Charles Darwin in 2004 (CD166) and RRS James Cook in 2008 (JC027).

In the Agadir Basin, a series of large-scale gravity-flow deposits (bed volumes > 10 km³) has been emplaced during the last 160,000 years (Fig. 2). Bed 1 is the youngest to be intensively studied (dated at ~ 1000 years BP) and Bed 14 is the oldest (dated at ~ 160,000 years BP). Several of these beds can be correlated onto the Madeira Abyssal Plain, e.g., Beds 1, 2, 5, 7, 10, 12, and 14 (Wynn et al., 2002b). The majority of beds are siliciclastic and derived from the Moroccan margin, although some are sourced from adjacent volcanic islands and seamounts, e.g., Beds 2, 10, and 14 (Fig. 2; Frenz et al., 2009).

It should be noted that previous studies used letters rather than numbers for individual beds on the Madeira Abyssal Plain (see Weaver et al., 1992). In this study we use a numeric scheme for beds throughout the Moroccan Turbidite System (Fig. 2), although beds on the Madeira Abyssal Plain are referred to both numerically and alphabetically for consistency with previous work (e.g., Bed 5(d); see Wynn et al., 2002b for details). References to deposit volumes of individual beds across the Moroccan Turbidite System are largely based upon the results of Wynn et al. (2002b) and Frenz et al. (2009). “Small-volume” beds are those with deposit volumes of < 10 km³, e.g., Beds 3, 1, 32, and 33. “Medium-volume” beds are those with deposit volumes between 10 and 100 km³, e.g., Beds 3 and 11. “Large-volume” beds are those with deposit volumes > 100 km³, e.g., Beds 5, 7, and 12 (Fig. 2).

Individual bed correlations in the Moroccan Turbidite System are based upon a multi-proxy approach, incorporating: (1) micropaleontological dating, using coccoliths and planktonic foraminifera tied to the oxygen isotope curve, (2) relative stratigraphic position, i.e., depth below seafloor, (3) composition of the sand fraction, (4) mud geochemistry and color (visual and scanned), (5) magnetic susceptibility, and (6) lateral thickness trends. These correlations are outlined in detail in Wynn et al. (2002b) and Frenz et al. (2009).

Estimates of bed volume were calculated following construction of sediment isopach maps for individual beds in each depositional basin. It is assumed that thickness variations within individual beds between core sites are linear. Each basin was gridded into 10 km x 10 km squares and the volume was calculated for each square. Bed volumes outside of depositional basins are not included due to a lack of core control and the complex depositional environment, e.g., distributary-channel networks. Therefore estimates of bed volume should be taken as minimum values.

RESULTS

The data presented here were collected from three broad areas along the main flow pathway: (1) the Agadir Basin, (2) the proximal Madeira Channels, and (3) the distal Madeira Channels and the Madeira Abyssal Plain. In each of these areas we docu-
Fig. 2.—Core log panel for CD166/12, illustrating the stratigraphy of gravity-flow deposits (beds) in the Agadir Basin for the last 160,000 years. Core location is shown in Fig. 4. Individual bed boundaries are defined by color shading. Graphic log is based upon visual logging. Grain-size classes are as follows: A = clay and silt (< 63 µm); B = fine sand (63–250 µm); C = medium sand (250–500 µm); D = coarse sand (500–2000 µm); E = gravel (> 2000 µm). Measured grain-size data (mean and mode) were collected using a Malvern Mastersizer 2000. Details of provenance and grain size are from Frenz et al. (2009). Note the upward-fining signature of most large-volume beds, contrasting with the ungraded character of the muddy sand “debrite” in Bed 5.
ment the response of siliciclastic gravity flows to changes in seafloor gradient.

**Agadir Basin**

The Agadir Basin lies at a water depth of 4300–4500 m and covers an area of ~35,000 km² (Figs. 1, 3, 4). Siliciclastic flows sourced from the Moroccan continental margin enter this basin from the east via the Agadir Canyon (Fig. 1). Previous work (Wynn et al., 2002a; Wynn et al., 2002b) has revealed that axial slope angles throughout much of the lower Agadir Canyon are in the region of 0.2–0.4° (note that for consistency with previous studies on the Moroccan Turbidite System, slope angles are quoted in degrees). The canyon mouth is characterized by a zone of intense seafloor scour, interpreted to result from rapid flow expansion of large-volume turbidity currents (Wynn et al., 2002a; Wynn et al., 2002b; Talling et al., 2007).

The overall axial (ENE–WSW) gradient across the floor of the Agadir Basin is <0.1°, with depth increasing to the WSW (Figs. 3, 4; Wynn et al., 2002b). Two flat sub-basins can be identified (east and west), separated by a minor ramp (Fig. 4). Flows exiting the Agadir Canyon initially pass over a slope of >0.05° prior to reaching the flat eastern sub-basin floor (Figs. 4, 5). While crossing this slope, large-volume flows eroded the seafloor or bypassed their sediment load. For example, in core CD166/49, Bed 5 comprises a gravel lag, while in cores CD166/48 and CD166/51 it is represented by erosional hiatuses overlain by upward-finishing turbidites containing Bouma divisions Te–Ta (Fig. 5; Bouma, 1962). Erosional hiatuses are identified through comparison with adjacent stratigraphy, and represent localized erosion of up to 1.0 m of sediment. In contrast, small-volume flows deposit thin mud turbidites in this area, e.g., Beds 1.5 and 3.1–3.3.

The eastern margin of the eastern sub-basin occurs at ~4390 m water depth and is marked by a reduction in slope from 0.05° to <0.01° (Figs. 4, 5). This slope break has a major influence on flow behavior and the resulting deposit character. Several of the small-volume beds pinch out abruptly, e.g., Beds 1.6 and 3.1–3.3 (Fig. 5). Medium-volume beds, e.g., Beds 3 and 11, show a marked decrease in sand–mud ratio across the slope break. However, the most spectacular response is shown by the large-volume flow that deposited Bed 5, which changed from an erosive (and dominantly bypassing) turbidity current to a depositional two-phase flow over a distance of a few kilometers (Fig. 5). In the eastern sub-basin, Bed 5 comprises a distinctive “linked turbidite–debrite” deposit up to 2.5 m thick (Talling et al., 2004; Talling et al., 2007). This deposit contains: (1) a basal lag or graded turbidite sand (Bouma Ta–Tb), overlain by (2) ungraded muddy sand “debrite” containing randomly distributed mud clasts, overlain by (3) ungraded turbidite mud (Bouma Te). This tripartite bed structure is attributed to flow transformation of the parent flow. An initially (weakly or strongly) turbulent bypassing flow, loaded with mud through erosion of fine-grained seafloor sediments in the Agadir Canyon and eastern Agadir Basin, decreased in velocity as it crossed the slope break. The head of the flow remained relatively turbulent and dilute, and slightly eroded the seafloor and/or deposited a clean turbidite sand through rapid “layer-by-layer” aggradation (Figs. 2, 5). However, the decrease in velocity led to the higher-concentration central portion of the flow depositing its load in mass, forming the ungraded muddy sand deposit. The tail of the flow was a fine-grained suspension cloud, which settled out slowly to form the ungraded mud cap (see Talling et al., 2007, for a more detailed discussion of this process).

The only other large-volume bed that was cored either side of the slope break is Bed 7, which is a more mud-dominated deposit than Bed 5. This bed shows a fourfold thickness increase across the slope break, from <15 cm in cores CD166/48 and CD166/49 on the ramp (note that Bed 7 is absent in core CD166/51 due to erosion beneath Bed 5) to ~65 cm in cores CD166/57 and CD166/34 on the step. Unlike Bed 5, Bed 7 maintains its low sand–mud ratio and upward-finishing aggradational turbidite character across the slope break (Fig. 5).

The western margin of the eastern sub-basin (at ~4405 m WD) is marked by a very subtle increase in gradient, from <0.01° to ~0.02° (Fig. 4). This gradient change appears to have had little effect on the majority of flows; for example, Beds 3.7, 11, and 12 all show little variation across this boundary. However, Bed 5 shows evidence for transformation back into a dominantly turbulent flow, resulting in deposition of a relatively thin but continuous fining-upwards Bouma sequence (Tb to Te) at core site CD166/31 (Fig. 5). As the flow moved across the gently sloping (~0.02°) floor of the western sub-basin it remained dominantly turbulent and relatively thin, although in some

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**Fig. 3.**—Gradient profile along the axis of the Moroccan Turbidite System at two different vertical exaggerations (55x and 8x). Vertical scale is in meters. Profile follows the main pathway for siliciclastic flows sourced from the Moroccan margin, and is based upon GEBCO bathymetry. See Fig. 1 for location of profile line (X–Y). Once beyond the mouth of Agadir Canyon, siliciclastic flows traveled for hundreds of kilometers across slopes of <0.1°.
Fig. 4.—A) Shaded bathymetry map and B) gradient map of the Agadir Basin, based upon GEBCO data. For location see Figure 1. Bathymetric contours are spaced at 100 m intervals on the bathymetry map and at 500 m intervals on the gradient map. Areas with EM12 multibeam bathymetry data coverage are shown on the bathymetry map; these data were collected during research cruise CD166 and cover the Agadir Canyon mouth and eastern Madeira Channels. Core locations are shown by white circles; those used in this study (all CD166 cores) are numbered and shown in yellow (see Figs. 5 and 7). Red dashed line shows gradient profile line used in Figures 2 and 5. Western (W) and eastern (E) sub-basins of Agadir Basin are picked out on the gradient map, and the flat floor of each sub-basin (<0.02°) is outlined with white dashed line. SAP = Seine Abyssal Plain.
Fig. 5.—Core correlation panel across the Agadir Basin, hung off A) seafloor and B) the basis of Bed 5. For location of profile line, and illustrated cores, see Figure 4. Seafloor profile in Part A is taken from GEBCO bathymetry data. Bed classification and graphic-log grain-size scales are the same as those used in Fig. 2. Dashed vertical red lines delimit areas of different gradient, marked in Part A as ramps and steps (although note that calculation of precise slope angles in the western Agadir Basin is hindered by poor bathymetric data resolution). Precise water depths measured at each core location are shown in brackets below core numbers in Part A. Note that the vertical scale of the core logs is at a scale different from the present-day seafloor profile in Part A. Asterisks mark erosional hiatuses below Bed 5 in cores CD166/48 and CD166/51. Note the marked changes in bed character between cores CD166/51 and CD166/57 (where present-day slope decrease from 0.05° to < 0.01°), and between cores CD166/24 and CD166/22 (where present-day slope increases from 0.02° to 0.06°).
places the middle part of the flow resumed deposition of a muddy sand “debrite”. For example, in cores CD166/12, CD166/28, and CD166/29 (Figs. 2, 5), a well-developed basal laminated sand ($T_b$) is overlain by variably developed muddy sand “debrite” and capped with turbidite mud ($T_e$). It appears that the slight increase in gradient between the two sub-basins led to an increase in velocity and a temporary return to turbulent flow conditions within the Bed 5 flow. However, while crossing the almost flat floor of the western sub-basin, the velocity locally decreased and the high-concentration central portion of the flow underwent en masse deposition once more.

**Proximal Madeira Channels**

The western margin of Agadir Basin is marked by an increase in gradient and development of a distributary channel network (Figs. 1, 3, 4, 6). These channels, here referred to as the Madeira Channels, connect the Agadir Basin and the Madeira Abyssal Plain, with the largest flows from the Moroccan margin continuing westwards along this pathway. Previous work by Masson (1994) and Wynn et al. (2002b) showed that axial slope angles along the proximal (eastern) Madeira Channels range from 0.05° to 0.14°, and that the channels were dominated by sediment bypass in the late Quaternary. New data collected during research cruise CD166 allow this transition from unconfined to partially confined flow to be examined in more detail (although note that a detailed geophysical and sedimentological study of the Madeira Channels is in preparation as part of a separate contribution).

The Madeira Channels have two main branches composed of multiple incisional channels, but only the northern branch is a pathway for siliciclastic flows exiting the Agadir Basin (Fig. 1). Two individual channels are incised into the western margin of Agadir Basin (Fig. 6), and both of these channels start at the precise point where slope angles increase from < 0.02° to > 0.06° (Figs. 3, 4; Masson, 1994). This gradient increase, combined with the transition from unconfined to confined flow, exerts significant influence on flows passing westwards across the

**Fig. 6**—Shaded bathymetry map of the Madeira Channels, based upon GEBCO data. Overlaid are EM12 multibeam bathymetry data collected during research cruise CD166. For location see Fig. 1. Bathymetric contours are spaced at 100 m intervals. Core locations are shown by white circles; those used in this study (all CD166 cores) are numbered and shown in yellow (see Figures 5 and 7). The white dashed line (X–Y) across the northern channel shows the location of the core correlation panel (Fig. 7).
“knickpoint”. In particular, large-volume Beds 5, 7, and 12 all show evidence for significant bypass at the heads of the channels. For example, in core CD166/22, Beds 5 and 7 are completely absent, and Bed 12 is reduced to a thin laminated sand (Bouma Tb; Fig. 5). All of these beds are present farther west on the Madeira Abyssal Plain, and the regional bathymetry shows that the northern Madeira Channels are the only potential route westwards (Fig. 1).

A further transect of cores was collected across the larger of the northern channels, some 50 km downslope from the spill point (Figs. 6, 7). Initial analysis of these cores reveals that the channel axis is a zone of dominant bypass; for example, core CD166/17 contains no siliciclastic turbidite but does contain indications for a thin bioturbated mud deposit corresponding to volcaniclastic “marker” Bed 14 (identified by its distinctive geochemical signature; Frenz et al., 2009). However, two cores located on the northern margin of the channel do contain siliciclastic sands that appear to correspond to Beds 5 and 12, but these are only thin rippled layers (Bouma Tc) with no mud caps (Fig. 7). The (almost) complete bypass within the channel, and

Fig. 7.—Core correlation panel and 3.5 kHz profile across the northern Madeira Channel. For location of profile line, and illustrated cores, see Figure 6. The core correlation panel is hung off the present-day seafloor, with the seafloor profile derived from 3.5 kHz echo-sounder data. Bed classification and graphic-log grain-size scales are the same as those used in Figure 2. Three distinctive glacial clay units that aided visual correlation of the lower sections of the cores are also shown. Note the almost complete absence of any of the studied beds in the channel axis (CD166/17). Bed 14 is identified in this core by its distinctive geochemical signature (Frenz et al., 2009). Two thin siliciclastic sand beds, identified in core CD166/19 some 5 km from the channel, are probably Beds 5 and 12. Note that the 3.5 kHz profile shows increased acoustic penetration within the channel compared to the channel margins, supporting the finding that coarse material has wholly bypassed the channel axis. Also note the presence of the U-shaped channel form to a depth of at least 20 m below seafloor.
partial bypass on the channel margins, is supported by 3.5 kHz echo-sounder profiles, which show increased acoustic penetration within the channel compared to the channel margins (Fig. 7). The above interpretations assume that the U-shaped channel form has been present throughout the last 200,000 years. This is also supported by 3.5 kHz profiles, which indicate that the channel form is persistent in the subsurface to a depth of at least 20 m (Fig. 7).

Distal Madeira Channels and Madeira Abyssal Plain

The distal (western) section of the northern branch of the Madeira Channels has an average axial slope angle of 0.11° (Fig. 8; Masson, 1994; Wynn et al., 2002b). The northernmost channel (which follows the line of profiles in Figures 1 and 3) runs around the southern margin of Madeira Rise and continues to the northwest, before terminating on the northeast margin of the flat Madeira Abyssal Plain (Figs 1, 3, 8). The slope profile shows a marked increase in gradient associated with termination of the spur at the southwest edge of Madeira Rise (Figs. 1, 3); this gradient increase corresponds to a knickpoint, and an associated decrease in channel width and an increase in channel incision from 5–10 m to 20 m (Masson, 1994).

The Madeira Abyssal Plain is an enormous enclosed basin, covering an area of almost 70,000 km² and lying at a water depth of ~5400 m. Slope angles are generally <0.01° across the basin floor, although the plain is punctuated by volcanic seamounts (Fig. 8). These seamounts are draped by pelagic and hemipelagic sediments that are a source for small-volume carbonate-rich turbidites (Weaver et al., 1992). However, the plain is dominantly filled by volcaniclastic and siliciclastic turbidites entering from the east. Previous work on selected cores from the Madeira Abyssal Plain has revealed that deposition of turbidite sand is concentrated in broad terminal lobes at the end of the Madeira Channels, associated with an abrupt decrease in gradient (Wynn et al., 2000).
Four cores were selected along an ~275-km-long transect that extends from near the termination of the northern Madeira Channels to the central Madeira Abyssal Plain. This transect allows us to investigate how flows respond to the transition from channelized to nonchannelized seafloor and the associated decrease in gradient (Figs. 8, 9). Core D11816 is located just south of the distal termination of the northern Madeira Channels, and contains three large-volume silicilastic beds: 5(d), 7(e), and 12(0) (Fig. 9). These beds are all <50 cm thick, and are dominated by upward-fining, ripple- and planar-laminated silts and fine sands (Bouma Tc and Td), overlain by thin mud caps (Bouma Tc). Farther to the southwest these beds show a dramatic increase in mud-cap thickness, while the fine sand and silty bases thin or pinch out (Fig. 9). The thickest mud caps are found in core D10698, in the deepest (and most distal) part of the Madeira Abyssal Plain. Here, Bed 5(d) is 0.8 m thick, Bed 7(e) is ~3.0 m thick, and Bed 12(0) is ~4.0 m thick.

DISCUSSION

Previous studies have demonstrated that natural deep-water gravity flows show a range of erosional and depositional responses to changes in gradient. For example, Friedmann et al. (2000) found that abrupt decreases in local gradient lead to decreased erosive power, decreased flow confinement, rapid deposition, channel widening, and fractional freezing (of debris flows). Abrupt increases in local gradient lead to incision, increased confinement, narrowing of channels, and knickpoint development. The data presented here broadly support these findings, but our ability to identify and correlate individual beds at a basin scale has also provided new insights into the response of individual gravity flows to changes in seafloor gradient.

Perhaps the most significant result of this study is the remarkably low gradient and long distance across which the studied flows traverse. For example, the flows responsible for depositing Beds 5, 7, and 12 travelled over slopes of <0.01° for >1000 km, between the mouth of Agadir Canyon and the Madeira Abyssal Plain (Fig. 3). Related to this is the surprisingly subtle nature of the gradient changes that influence flow behavior. The slope break in the eastern Agadir Basin marks a transition from just <0.05° to <0.01° (Fig. 4), and yet flows passing across this slope break exhibit evidence for dramatic changes in their ability to transport or deposit sediment. Several small-volume beds pinch out completely, while one of the large-volume beds (Bed 5) shows evidence for a dramatic flow transformation (Fig. 5; Talling et al., 2007). The relative change in gradient at these low values is significant. For example, the transition described above represents a fivefold decrease in gradient, and it is this relative value, in low-gradient areas, that appears to be of crucial importance in influencing flow behavior (cf. Friedmann et al., 2000). Other factors, such as the sediment composition, grain size, and volume of the flow, may also be important in some cases; for example, small-volume flows behave very differently from medium- or large-volume flows at a given point in the flow pathway (e.g., Fig. 5; Wynn et al., 2002b).

There are some indications that depositional topography of individual beds may influence slope gradient and subsequent flow deposits, but the present dataset is probably too limited to explore this in detail. For example, Bed 5 is up to 2.5 m thick in the eastern Agadir Basin, in an area where present-day variations in seafloor depth may be <4 m over a horizontal distance of 50 km (e.g., between cores CD166/57 and CD166/33). It seems likely that deposition of Bed 5 would have subtly reduced the gradient at the slope break between cores CD166/51 and CD166/57, possibly influencing the subsequent pinch-out of several small-volume flow deposits out at this location, e.g., Beds 1.6, 3.1, 3.2, and 3.3 (Fig. 5). Prior to deposition of Bed 5, the only small-volume turbidites recovered in cores are Beds 7.1 and 7.2, which pinch out farther upslope of the present-day slope break (Fig. 5). Unfortunately, lack of core penetration means that this hypothesis cannot be fully tested on a statistically robust number of beds. In addition, differential compaction of spatially heterolithic sediments is hard to assess with accuracy, as is the possible influence of basin subsidence/uplift, hindering reconstruction of paleoslopes.

Another key result of this study is the ability of flows to efficiently bypass much of their sediment load across areas of increased seafloor gradient. Our preliminary analysis of data from the Madeira Channels indicates that flows exiting the Agadir Basin switch from a partially depositional state to an almost wholly bypassing state, associated with the transition from unconfined to confined flow and an increase in gradient (Figs. 5, 7). The location of the channels themselves is presumably a response to increased gradient and the ability of individual flows to incise the seafloor (e.g., Friedmann et al., 2000; Pirmez et al., 2000). On the Madeira Abyssal Plain, the exceptionally thick mud caps displayed by Beds 7 and 12 (Fig. 9) highlight the ability of muddy suspensions to bypass right to the very end of the transport pathway, despite having previously passed over the almost flat floor of the Agadir Basin. The exact process driving this phenomenon has been discussed previously (e.g. McCave and Jones, 1988), and is currently being revisited as part of an ongoing study using core data presented here.

Flow transformations in submarine gravity flows are increasingly being recognized in both modern (e.g., Talling et al., 2007) and ancient (e.g., Ito, 2008) deep-water environments. Several models have been proposed outlining the processes involved, but this study adds a new dimension in that multiple transformations can be recognized in a single flow within a discrete basin. The response of the Bed 5 flow to the gradient decrease in eastern Agadir Basin has already been documented, with an initially turbulent flow decreasing in velocity leading to deposition of a tripartite “linked turbidite–debrite bed” in the eastern sub-basin (Talling et al., 2004; Talling et al., 2007). However, a subsequent increase in slope, from <0.01° to 0.02° appears to have been sufficient to initiate a transformation back into a weakly or strongly turbulent flow in the central basin (Fig. 5). Interestingly, this minor ramp appears to have had little influence on other, fully turbulent flows (Fig. 5). The Bed 5 flow subsequently decreased in velocity while crossing the floor of the western sub-basin, and linked turbidite–debrite deposition was resumed. Finally, the flow increased in velocity upon reaching the western basin margin, with spillover into the Madeira Channels being characterized by turbulent flow and sediment bypass (Figs. 5, 7).

CONCLUSIONS

The ability to confidently link gravity-flow deposits to a specific depositional environment is one of the major advantages of data collected from modern deep-water systems. In addition, accurate measurements of gradient can be obtained, which is generally not possible in outcrop or the subsurface. This study has utilized a unique modern dataset from the Moroccan Turbidite System, in order to highlight the important role that slope gradient plays in the transport and deposition of sediment by individual deep-water gravity flows. This is a significant advance on previous studies investigating the influence of gradient on such flows, which have been limited to laboratory experiments or analysis of geomorphological features resulting from single flows.

Some of the results of this study were predictable, and are compatible with previous work. For example, marked decreases
Fig. 9.—Core correlation panel across the northeast Madeira Abyssal Plain. For location of profile line, and illustrated cores, see Figure 8. The profile is hung off the seafloor, with the seafloor profile interpreted from core depth data (due to insufficient GEBCO data quality in this area). Bed classification and graphic-log grain-size scales are the same as those used in Figure 2. Note the marked increase in thickness of siliciclastic beds 5(d), 7(e), and 12(f) in the deepest part of the basin (cores CD10688 and CD10698).
in gradient can lead to small-volume flows dying out and large-volume flows depositing thick mud caps. In addition, marked increases in gradient can lead to sediment bypass and/or erosion and, in some cases, a transition from unconfined to confined flow. However, more unexpected is the evidence for multiple flow transformations related to slope breaks, as shown by Bed 5 in the Agadir Basin. The efficiency of bypass within the northeastern Madeira Channels is also noteworthy, with cores from the channel axis containing little evidence of the flows that passed through them, even though cores farther down-system (from the Madeira Abyssal Plain) indicate that the flows were still carrying significant volumes of both sand and mud. A detailed geophysical and sedimentological analysis of this zone of sediment bypass is currently in preparation.

Although the absolute changes in slope angle across much of the study area appear remarkably subtle, the relative rate of change appears to be particularly significant, e.g., the slope break from 0.05° to < 0.01° in the eastern Agadir Basin actually represents a fivefold decrease in slope and has a major influence on flows of all sizes. However, the specific response of individual flows to changes in gradient is varied, due to variations in sediment volume, composition, and grain size.

Finally, this study has shown that large-scale, dominantly turbulent gravity flows, containing a mixture of sand and mud, are highly efficient at transporting sediment for vast distances across remarkably low slopes. The response of these flows to gradient changes is significant, and can lead to rapid lateral variations in bed geometry and sediment facies. These conclusions have important implications for geoscientists analyzing analogous sequences in outcrop or the subsurface, where knowledge of paleoslope is often limited.

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