Canyon morphology on a modern carbonate slope of the Bahamas: Evidence of regional tectonic tilting

T. Mulder[1], E. Ducassou[1], H. Gillet[1], V. Hanquiez[1], E. Tournadour[1], J. Combes[1], G.P. Eberli[2], P. Kindler[3], E. Gonthier[1], G. Conesa[4], C. Robin[5], R. Sianipar[2], J.J.G. Reijmer[6], and A. François[1]
[1]Université de Bordeaux, UMR 5805 EPOC, 33405 Talence cedex, France
[2]Division of Marine Geology and Geophysics, University of Miami, Miami, Florida 33149, USA
[3]Section of Earth and Environmental Sciences, University of Geneva, 1205 Geneva, Switzerland
[4]Géologie des Systèmes et Réservoirs Carbonatés, Université de Provence, Marseille, France
[5]Géosciences Rennes, Université de Rennes 1, Rennes, France
[6]Faculty of Earth and Life Sciences (FALW) Department of Sedimentology and Marine Geology, VU University Amsterdam, Amsterdam, Netherlands

ABSTRACT

New high-quality multibeam data presented here depict the northern slope of the Little Bahama Bank (Bahamas). The survey reveals the details of large- and small-scale morphologies that look like siliciclastic systems at a smaller scale, including large-scale slope failure scars and canyon morphologies, previously interpreted as gullies and creep lobes. The slope exhibits mature turbidite systems built by mass-flow events and turbidity currents. The sediment transport processes are probably more complex than expected. Slope failures show sinuous head scarps with various sizes, and most of the scars are filled with recent sediment. Canyons have amphitheater-shaped heads resulting from coalescing slump scars, and are floored by terraces that are interpreted as slump deposits. Canyons rapidly open on a short channel and a depositional fan-shaped lobe. The entire system extends for ~40 km. The development of these small turbidite systems, similar to siliciclastic systems, is due to the lack of cementation related to alongshore current energy forcing the transport of fine particles and flow differentiation. Detailed analyses of bathymetric data show that the canyon and failure-scar morphology and geometry vary following a west-east trend along the bank slope. The changing parameters are canyon length and width, depth of incision, and canyon and channel sinuosity. Accordingly, failure scars are larger and deeper eastward. These observations are consistent with a westward tectonic tilt of the bank during the Cenozoic.

INTRODUCTION

Siliciclastic turbidite systems have been intensively studied as oil exploration targets. Longitudinally, they are made up of a canyon incising the upper continental slope and the shelf. The system begins with a canyon head that usually shows evidence of rotational shallow slope failures. The canyon head opens on a deep valley with sharp edges, ending at a canyon mouth usually located at the base of the continental slope. There, the canyon opens on a channel-levee complex where both erosion in the channel and deposition on the levees coexist. The system ends with a depositional lobe complex, which is channelized over most of its length. In siliciclastic systems, most of the canyons are directly or indirectly connected to a river source (Reading and Richards, 1994). Some other canyons are fed by littoral drifts in the uppermost part of the continental shelf (0–20 m; Burke, 1972). The length of the canyon is an indicator of canyon maturity (Shepard and Dill, 1966). For V-shaped canyons, canyon width is correlated with canyon depth; a larger width indicates greater erosion. A greater height of the canyon edge also indicates greater incision.

Canyons located on carbonate slopes are substantially different from those receiving river input. They have no riverine sediment source, and the only sediment sources are either autochthonous, including mainly sediment productivity, or allochthonous, including sediment load from shelf currents and wind (Playton et al., 2010).

In this paper we present new high-resolution morphological data collected along the northern slope of Little Bahama Bank. We show that the turbidite systems found there are far more complex than previously supposed, and show some similarities to siliciclastic systems. We test to ascertain if the lateral evolution of the system morphology is consistent with regional tilting of this part of the bank.

SETTING AND METHODS

We focused our study on the northern windward margin of Little Bahama Bank, north of the island of Grand Bahama (Bahamas; Fig. 1A). Carbonate deposition in the Bahamas has occurred since the Cretaceous and perhaps since the Jurassic (Eberli and Ginsburg, 1989): the platform has aggraded ~1500 m since the Miocene. Little Bahama Bank is located on the passive margin of the North American plate, but is close to the Blake-Bahama Outer Ridge and the Blake Escarpment, which has subsided at a rate of 20 m/m.y. during the Cenozoic (Freeman-Lynde et al., 1981). South of the Bahamian archipelago, tectonic structures include the Cuban fault system, including the Nortecubana and Cauto-Nipe faults (Cottila Rodriguez et al., 2007), and the Bahama Escarpment, a Jurassic rift structure. The study area is only 500 km away from the active North American–Caribbean plate boundary. Recent tectonic activity in the Bahamian archipelago has been substantiated near Walkers Cay (Mullins and van Buren, 1981) and along Mayaguana Island (Kindler et al., 2011). The linear edge of Mayaguana could be related to a fault, and the island may have undergone tectonic tilting ~500 k.y. ago. An average subsidence rate of 1–2 m/100 k.y. was estimated for the Bahama Banks for the past 30 m.y. (Mullins and Lynts, 1977). It has been shown that in the Cretaceous and early Cenozoic, the evolution of the Great Bahama Bank was strongly influenced by global tectonic events related to the opening of the Atlantic Ocean, Caribbean plate motion, and the collision between Cuba and the Bahamian platform (Masaferro and Eberli, 1999).

Between Little Bahama Bank and the Blake Plateau, the Antilles Current flows westward with alternate bands of low (<10 cm/s) and high (10–40 cm/s) bottom velocity (Ingham, 1974; Fig. 1A), and has formed large and thick detached contourite drifts (Mullins et al., 1980). The drifts thicken in the distal part of the bank where the current joins the Florida Current to form the Gulf Stream. The windward margin and slopes of Little Bahama Bank are exposed to northward-moving Atlantic storms and energetic swells, and receive coarse- and fine-grained sediment exported from the bank (Lantzsch et al., 2007).

Methods and Acquired Data

The Carambar cruise was conducted from 31 October to 29 November 2010 on the R/V *Le Suroît* (Mulder et al., 2012). Its main objective was to better understand the transport of carbonate particles along submarine slopes. The second part of the cruise focused on Little

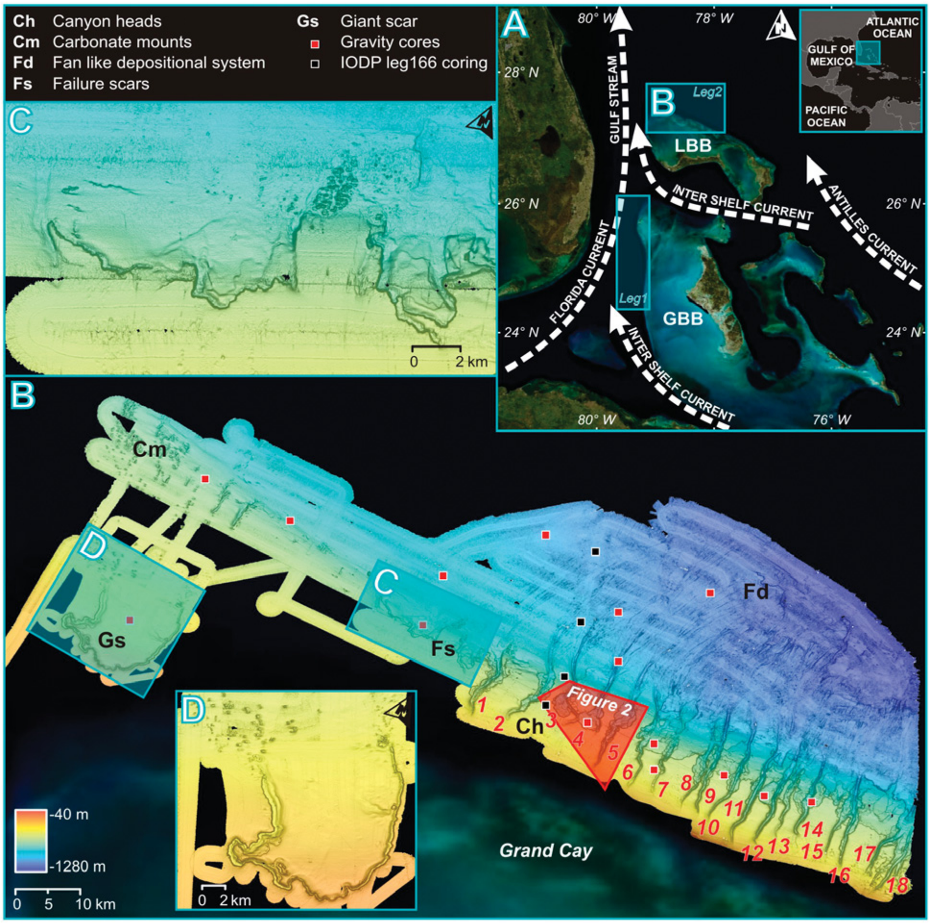

Figure 1. A: Location of two study legs of Carambar cruise. White dashed arrows are trajectories of main oceanic currents in western part of Bahamian archipelago. LBB—Little Bahama Bank; GBB—Great Bahama Bank; IODP—Integrated Ocean Drilling Program. B: Bathymetric map of carbonate slope located north of LBB; canyon numbers are in red. C: Detail of failure scar area. D: Detail of giant scar.

Bahama Bank (Fig. 1A). Data collected during this leg include more than 4560 km^2 of multibeam bathymetry and acoustic imagery (Kongsberg EM302 echosounder), 1390 km of very high (echo subbottom profiler) and high-resolution (multichannel) seismic profiles, and 16 Kullenberg gravity cores (Fig. 1B). This new data set on carbonate slopes adds to the extensive data for carbonate banks, including academic and industrial seismic lines and the results of Ocean Drilling Program (ODP) Leg 101 (Austin et al., 1988).

RESULTS AND DISCUSSION

Sedimentary Processes Along Little Bahama Bank

The northern slope of Little Bahama Bank extends over 75 km, and is ~24 km wide. It comprises an upper slope (300–900 m) and a lower slope (900–1300 m) with different gradients (3%–4% and 1%–1.5%, respectively; Mullins et al., 1984). The new multibeam bathymetric survey shows sedimentary morphologies more complex than the previously described gullies and creep lobes (Harwood and Towers, 1988). Complete turbidite systems are identified along the slope, extending between ~300 m and ~1300 m of water depth with a canyon, a short channel, and a depositional lobe. True canyons are on the upper slope (Fig. 1B); some open rapidly on a short (10–15 km long) channel that in places has levee-shaped structures and ends with a depositional lobe. In these lobes, the soft sediment cover appears to be very thin, as attested to by the low penetration of the acoustic signal in very high resolution seismic images and high backscatter on acoustic imagery. The absence of substantial deposition could result from four processes: (1) removal of the fine grains by storms and tidal currents; (2) pirating by along-slope contour currents; (3) elutriation by downslope turbidity currents; (4) early diagenesis and cementation on the top of the slope. All systems do not extend over more than 40 km and all are deflected in an eastward direction, toward the deepest basin floor. Because no point source exists, no large submarine system forms, as in siliciclastic-fed submarine slopes. Their size compares to those developed along the margin of eastern Corsica (Bellaiche et al., 1994). The presence of true turbidite systems suggests that initial slope failures occurring at the canyon head can differentiate into sediment flows and turbidity currents, as suggested by the fine-grained turbidites described by Mullins et al. (1984). Consequently, longitudinal sediment sorting can result from this flow transformation, as occurs in siliciclastic-fed systems. This similarity to siliciclastic systems is possible due to the lack of carbonate cementation along the middle and base of the slope because of along-slope current activity (Harwood and Towers, 1988). Noncemented fine-grained carbonate particles can then be transported individually, allowing for downslope flow transformation.

Submarine Canyon Morphology

Among the 22 submarine canyons identified by Mullins et al. (1984), 18 were surveyed in this study (Fig. 1B), and 16 systems have been totally imaged. Detailed canyon morphology was studied after processing the bathymetry data with Caraibes software (Ifremer) and the ArcGIS system (Esri) (Table 1). Canyon heads usually exhibit an amphitheater shape (Fig. 2A);

TABLE 1. SYNTHESIS OF MAIN MORPHOMETRIC VALUES FOR 16 CANYONS

Canyon	1*	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Water depth of canyon head (m)	440	500	550	535	460	440	490	490	480	480	410	410	480	460	500	490
Canyon length (km)	12	14.94	15	15.16	19.6	15.05	13.76	12.65	17.16	16.88	18.63	18.3	16	16.41	14.07	15.5
Canyon width (m)	440	438	440	512	810	900	427	408	400	519	656	666	630	1170	1010	762
Thalweg depth (m)	63	38	40	46	48	52	35	35	39	43	64	72	35	80	68	62
Canyon sinuosity	1.09	1.12	1.16	1	1.03	1.07	1.11	1.04	1.02	1.03	1.03	1.06	1.05	1.09	1.05	1.01

*See Figure 1B for canyon numbering.

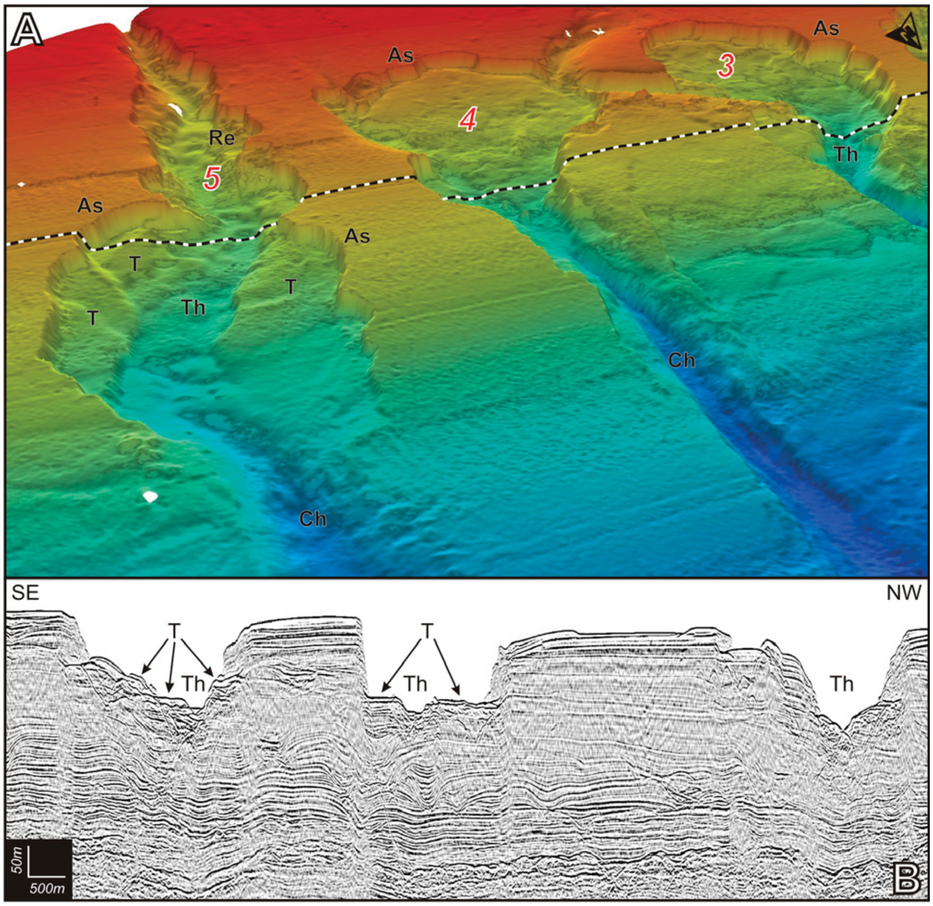

Figure 2. A: Detail of canyon head showing coalescing arcuate scars (As) forming amphitheater envelope, retrogressive erosion (Re), terraces (T), thalweg incision (Th), and channel (Ch). Canyon numbers are in red. See Figure 1B for location. B: High-resolution multichannel seismic profile showing thalweg incision through morphologic terraces. See Figure 1A for location (black and white dashed line).

close-up views show that the amphitheater results from coalescing, arcuate slump scars, suggesting that canyons form by successive failures of the wall edges of the canyon head at the slope top (Fig. 2A). Some canyon heads (e.g., 3 and 4 in Fig. 1B) show that slump scars extend upslope in a preferential direction and transform into a shallow small valley. These structures were interpreted as sediment transport direction by Harwood and Towers (1988). Our data interpretation is slightly different; we suggest that the canyons form by retrogressive erosion along an initial gully (see Shepard, 1981). Canyon heads appear at the slope top at water depths ranging from 410 m to 535 m (Fig. 1B; Table 1). The absence of canyons on the uppermost part of the slope is likely related to a high degree of early cementation (Mullins et al., 1984) that increases shear resistance and prevents sediment from failure. The morphological data show significant variations in geomorphologic parameters from east to west (Table 1). In the western part, canyons 1–8 are shorter than 15.2 km (except for canyon 5) while canyons 9–16 are longer than 16 km in the eastern part (except for canyons 15 and 16). Canyons are also wider in the eastern part than in the western part: canyons 1–11 are narrower than 520 m, with the exception of canyons 5 and 6. Canyons 12–16 are wider than 600 m. The depth of incision is greater for the eastern canyons than for the western ones, except for canyon 1. All canyons in the studied area are straight; however, the sinuosity of canyons is greater in the western part (canyons 1–7; but canyons 4 and 5 have a sinuosity >1.06) than in the eastern part (canyons 8–16, but canyon 14 has a sinuosity <1.06). Geomorphological data analysis indicates that thalweg incision and canyon length and width increase eastward while canyon sinuosity decreases eastward.

Canyons are floored with flat elongated morphologies interpreted as terraces (Fig. 2A). Terraces are situated along the flank of the canyon headwalls. They do not have similar elevations and form rough seafloor morphology. Terrace distribution shows no relationship with water depth and their locations at the toes of the scars suggest that they are mass-flow deposits resulting from failures occurring along the canyon head, as suggested by Harwood and Towers (1988). In addition, high-resolution multichannel seismic profiles perpendicular to the canyon incisions show that terraces are covered by a thin sediment drape, suggesting that after failure terraces may act as nested levees where carbonate turbidites from spilling turbidity currents are deposited (Pichevin et al., 2003; Fig. 2B). Thalweg incision between terraces suggests erosion by downcanyon sediment-flow processes (Fig. 2A).

Failure Scars

In contrast to the conclusions of Harwood and Towers (1988), we find that failure scars are not restricted to canyon incisions. Analysis of geomorphologic data in the failure scar area (Fig. 1C) shows that the height of the scars, the depth of the distal part of the scar, the scar surface, and the scar length increase eastward while the depth of scar head decreases. This suggests that slumping processes are more intense in the eastern part of Little Bahama Bank than in the western part.

The largest slope failure is visible in the westernmost part of the upper slope (Fig. 1D), between 275 and 460 m water depth. Only a sinuous depression corresponding to the head scar is visible. Evidence to interpret this structure as a failure scar includes the following: (1) the vertical edge of the structure is always the upslope edge, (2) the echofacies are different inside and outside of the scar, and (3) the scar vanishes distally and ends with topographic depressions containing a mound inside; the depressions are interpreted as pockmarks and each relief is interpreted as a cold-water coral mound.

Tectonic Tilt of Little Bahama Bank?

To interpret the trend in geomorphologic features observed on the north slope of Little Bahama Bank, we tested the hypothesis of a westward tectonic tilt suggested by Austin et al. (1988). Canyons that are longer, deeper, wider, more incised, and have a lower sinuosity are present more in the eastern part than in the western part. Thus, eastern canyons show more erosion than western ones. Accordingly, they are above their equilibrium state in the east and below in the west. This morphologic difference is interpreted as the result of tilting: canyons have been moved upward on the eastern portion of the bank slope and downward on the westward part. The failure scars show a trend similar to that of the canyons and could be interpreted as the initial stage of canyon formation. Larger scars are located in the east and smaller scars occur in the west consistently with a westward tilt; slope readjustment is more important in the east than in the west. This tilting could result from tectonic activity affecting all Little Bahama Bank and a large part of the Bahamian archipelago, and is consistent with the formation of large mass-failure deposits, such as the buried debrites recovered from ODP cores (Austin et al., 1988) or the mass transport complexes observed both in seismic profiles and on seafloor along the Great Bahama Bank (Mulder et al., 2012). This, in addition to the presence of a large, partially buried, outsized slump scar in the western upper slope of the carbonate bank, suggests that several tectonic pulses affected the bank during the late Cenozoic. Masaferro et al. (2002) demonstrated that the Santaren anticline (Southern Santaren Channel, Great Bahama Bank) shows a series of episodic and nonsteady fold growth, resulting in the specific geometry of growth stratal patterns marking the interplay of sedimentation and tectonic fold uplift. All these failure scars, as well as the large failure scar located at the western end of the Little Bahama Bank upper slope, are filled by thick Quaternary contourite deposits due to the action of the Antilles Current; this suggests that the tilting is not recent.

The submarine morphology of the carbonate slope located north of Little Bahama Bank shows a spectacular set of canyons, small turbidite systems similar to siliciclastic ones, and failure scars. The canyons are definitely related to retrogressive erosion limited upslope by a cementation front defined by the nature of the sediment deposited at the top of slope (Mullins et al., 1984: Lantzsch et al., 2007). This succession shows a significant lateral change in morphology consistent with a westward tectonic tilting of the entire carbonate margin. This tilting can also provide an alternate explanation for the presence of islands mainly on the eastern part of Little Bahama Bank (e.g., Abaco Island). More energetic hydrodynamic conditions (Ball, 1967) would add to the tectonic forcing. The age of the tilting and its causes will be determined from future studies of cores and high-resolution seismic data. Based on the thickness of its sediment fill, the largest western scar could be related to an older tectonic event.

ACKNOWLEDGMENTS

We thank the captain and crew of the RV *Le Suroît* for the quality of the acquired data and Ifremer-Genavir for cruise organization. This work has been supported by the French INSU (Institut National des Sciences de l'Univers) Program "Actions Marges."

REFERENCES CITED

Austin, J.A., Jr., and 20 others, 1988, Proceedings of the Ocean Drilling Program, Scientific results, Volume 101: College Station, Texas, Ocean Drilling Program, 501 p., doi:10.2973/odp.proc.sr.101.1988.

Ball, M.M., 1967, Carbonate sand bodies of Florida and the Bahamas: Journal of Sedimentary Petrology, v. 37, p. 536–591.

Bellaiche, G., Droz, L., Gaullier, V., and Pautot, G., 1994, Small submarine fans on the eastern margin of Corsica: Sedimentary significance and tectonic implications: Marine Geology, v. 117, p. 177–185, doi:10.1016/0025-3227(94)90013-2.

Burke, K., 1972, Longshore drift, submarine canyons, and submarine fans in development of Niger Delta: American Association of Petroleum Geologists Bulletin, v. 56, p. 1975–1983.

Cotilla Rodriguez, M.O., Franzke, H.J., and Cordoba Barba, D., 2007, Seismicity and seismoactive faults of Cuba: Russian Geology and Geophysics, v. 48, p. 505–522, doi:10.1016/j.rgg.2006.08.004.

Eberli, G.P., and Ginsburg, R.N., 1989, Cenozoic progradation of north-western Great Bahama Bank, a record of lateral platform growth and sea-level fluctuations, *in* Crevello, P.D., et al., eds., Controls on carbonate platforms and basin development: Society of Economic Paleontologists and Mineralogists Special Publication 44, p. 339–351.

Freeman-Lynde, R.P., Cita, M.B., Jadoul, F., Miller, E.L., and Ryan, W.B.F., 1981, Marine geology of the Bahama escarpment: Marine Geology, v. 44, p. 119–156, doi:10.1016/0025-3227(81)90115-8.

Harwood, G.M., and Towers, P.A., 1988, Seismic sedimentologic interpretation of a carbonate slope, north margin of Little Bahama Bank, *in* Austin, J.A., Jr., et al., Proceedings of the Ocean Drilling Program, Scientific results, Volume 101: College Station, Texas, Ocean Drilling Program, p. 263–277, doi:10.2973/odp.proc.sr.101.143.1988.

Ingham, C., 1974, Velocity and transport of the Antilles current northeast of the Bahama Islands: Fish Bulletin, v. 73, p. 626–632.

Kindler, P., Godefroid, F., Chiaradia, M., Ehlert, C., Eisenhauer, A., Frank, M., Hasler, C.-A., and Samankassou, E., 2011, Discovery of Miocene to early Pleistocene deposits on Mayaguana, Bahamas: Evidence for recent active tectonism on the North American margin: Geology, v. 39, p. 523–526, doi:10.1130/G32011.1.

Lantzsch, H., Roth, S., Reijmer, J.J.G., and Kinkel, H., 2007, Sea-level related resedimentation processes on the northern slope of Little Bahama Bank (middle Pleistocene to Holocene): Sedimentology, v. 54, p. 1307–1322, doi:10.1111/j.1365-3091.2007.00882.x.

Masaferro, J.L., and Eberli, G.P., 1999, Jurassic-Cenozoic structural evolution of the southern Great Bahama Bank, *in* Mann, P., ed., Caribbean basins: Sedimentary Basins of the World 4: Amsterdam, Elsevier, p. 167–193.

Masaferro, J.L., Bulnes, M., Poblet, J., and Eberli, G.P., 2002, Episodic folding inferred from syntectonic carbonate sedimentation: The Santaren anticline, Bahamas foreland: Sedimentary Geology, v. 146, p. 11–24, doi:10.1016/S0037-0738(01)00163-4.

Mulder, T., and 14 others, 2012, New insights into the morphology and sedimentary processes along the western slope of Great Bahama Bank: Geology (in press).

Mullins, H.T., and Lynts, G.W., 1977, Origin of the northwestern Bahama Platform: Review and reinterpretation: Geological Society of America Bulletin, v. 88, p. 1447–1461, doi:10.1130/0016-7606(1977)88<1447:OOTNBP>2.0.CO;2.

Mullins, H.T., and van Buren, H.M., 1981, Walkers Cay fault, Bahamas: Evidence for Cenozoic faulting: Geo-Marine Letters, v. 1, p. 225–231, doi:10.1007/BF02462438.

Mullins, H.T., Neumann, A.C., Wilber, R.J., Hine, A.C., and Chinburg, S.J., 1980, Carbonate sediment drifts in the northern Straits of Florida: American Association of Petroleum Geologists Bulletin, v. 64, p. 1701–1717.

Mullins, H.T., Heath, K.C., van Buren, H.M., and Newton, C.R., 1984, Anatomy of a modern open-ocean carbonate slope: Northern Little Bahama Bank: Sedimentology, v. 31, p. 141–168, doi:10.1111/j.1365-3091.1984.tb01956.x.

Pichevin, L., Mulder, T., Savoye, B., Cremer, M., Piper, D.J.W., and Gervais, A., 2003, The Golo submarine system (east-Corsica margin): Turbidite system morphology and processes of terrace formation from high-resolution seismic Boomer profiles: Geo-Marine Letters, v. 23, p. 117–124, doi:10.1007/s00367-003-0131-4.

Playton, T., Janson, X., and Kerans, C., 2010, Carbonate slopes, *in* James, N.P., and Dalrymple, R.W., eds., Facies models 4: St. John's, Newfoundland, Geological Association of Canada, p. 449–476.

Reading, H.G., and Richards, M.T., 1994, The classification of deep-water siliciclastic depositional systems by grain size and feeder systems: American Association of Petroleum Geologists Bulletin, v. 78, p. 792–822.

Shepard, F.P., 1981, Submarine canyons: Multiple causes and long-time persistence: American Association of Petroleum Geologists Bulletin, v. 65, p. 1062–1077.

Shepard, F.P., and Dill, R.F., 1966, Submarine canyons and other sea valleys: Chicago, Illinois, Rand McNally, 381 p.

Manuscript received 17 February 2012
Manuscript accepted 18 February 2012

Printed in USA