INTRODUCTION

Mixed turbidite and contourite depositional systems exist on many continental margins (e.g., Locker and Laine, 1992; Carter and McCave, 1994; Rebescos and Stow, 2001; Knutz and Cartwright, 2003; Hernández-Molina et al., 2006; Uenzelmann-Neben, 2006). Deposition in such systems is focussed along the lower continental slope and rise and occurs under the influence of relatively short-lived, sporadic, gravity-driven currents, as well as long-lived steady along-slope currents (Heezen et al., 1966; Locker and Laine, 1992; Mulder et al., 2008). The interaction of gravity currents and bottom currents is most apparent where low-frequency alternations of contourite and turbidite predominance are preserved (Mulder et al., 2008), and depositional elements are resolved in seismic-reflection data. In such cases, it is likely that pre-existing morphology influenced the character of deposits originating from both down-slope gravity currents and along-slope bottom currents (Locker and Laine, 1992; Uenzelmann-Neben, 2006; Mulder et al., 2008). For instance, in the case of gravity flows, it is recognized that receiving-basin configuration is one of the primary factors controlling lithofacies distribution as gravity flows transit from shelf areas into deeper water (Prather, 2003; Steffens et al., 2003). Similarly for sediment drifts, pre-existing seafloor irregularities appear to be loci for the initiation of deepwater bottom-current deposits (Faugères et al., 1999; Viana et al., 2007; Hopfauf and Spieß, 2001; Wynn and Masson, 2008).

Although conceptual models of the interaction of sedimentary processes in mixed turbidite and contourite systems have been proposed (Heezen et al., 1966; Locker and Laine, 1992; Hernández-Molina et al., 2008; Viana, 2008), relatively few published studies have used 3D seismic data to clearly demonstrate the effects of morphological heritage in alternating gravity-current and bottom-current deposition. Viana et al. (2007) indicate that most studies of deepwater seismic geomorphology focus on describing and understanding depositional elements associated with downslope sediment transport primarily because of exploration interest in turbidite reservoirs (e.g., Posamentier and Kolla, 2003; Deptuck et al., 2003; Davies et al., 2004). Recently, a number of authors have demonstrated the effectiveness of applying 3D seismic data to sediment-drift studies (Knutz and Cartwright, 2003; Holbein and Cartwright, 2006; Viana et al., 2007) and have emphasized the need for more detailed examination of these
deposits. In many areas, sediment drifts contribute significantly to the depositional record along the lower continental slope and rise (Wynn and Stow, 2002; Hernández-Molina et al., 2008).

Across much of the western Scotian margin off eastern Canada (Fig. 1), seismic-reflection data reveal evidence of alternating down-slope and along-slope deposition throughout the Neogene

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**Fig. 1.**—Location maps of study area. A) Regional map of the geomorphology of the continental margin off Nova Scotia. (Modified from Shaw and Courtney, 2004). B) Location map of the continental margin off western Nova Scotia, Canada, showing coverage of 2D and 3D seismic reflection data, the buried edge of the Abenaki carbonate bank, and the location of cross section A–A’. C) Location map of study area showing position of various seismic profiles discussed in the text.
GEOLOGICAL SETTING AND PREVIOUS WORK

The study area spans the middle continental slope to upper continental rise off southwestern Nova Scotia, Canada, in present-day water depths between 500 and 2900 meters below sea level (mbsl) (Fig. 1). The morphology of the modern Scotian margin includes a > 200-km-wide continental shelf consisting of transverse troughs and intervening banks, the product of repeated glaciations through the Quaternary (Piper, 1988). The shelf break lies at 80–130 mbsl (Mosher et al., 2004) and trends subparallel to the coastline for ~1000 km, from Northeast Channel in the west to Laurentian Channel in the east. The present-day slope extends from the shelf break to a decrease in gradient at 2000–2500 mbsl. The modern western Scotian margin has the characteristics of a bypass margin (Ross et al., 1994) with a steep (>4°) upper slope that extends to ~1000 mbsl and a gentler (<2°) lower slope and rise (Fig. 2).

Published studies of the Cenozoic geological history of the Scotian margin focus on the area near Sable Island, east of the study area, where data density is greatest (e.g., Swift, 1985; Swift et al., 1986; Swift, 1987; Ebinger and Tucholke, 1988; Wade et al., 1995; Long, 2002; Thomas, 2005; MacDonald, 2006; Fensome et al., 2008; Brake, 2009). In general, sediment accumulation appears to have been broadly controlled by changes in relative sea level; however, local patterns of sediment distribution were strongly influenced by local variations in sediment supply, slope morphology, erosion by bottom currents, and salt tectonics. Below the modern continental shelf and upper slope, episodes of localized outer-shelf progradation occurred in the Paleocene, Eocene, and Miocene (Wade et al., 1995; MacDonald, 2006), and major periods of channel formation and canyon incision occurred in the Paleocene, Eocene, Oligocene, Miocene, and Pleistocene (Thomas, 2005; MacDonald, 2006; Fensome et al., 2008; Brake, 2009). On the lower continental slope and rise, Swift (1987) and Ebinger and Tucholke (1988) reported a number of erosional unconformities in seismic-reflection data and suggested a bottom-current origin with inferred ages of Oligocene, Early Miocene, Middle Miocene, and Pliocene.

The Miocene to Pliocene seismic stratigraphy of the study area is described by Campbell and Mosher (2010). In the study area, the edge of the Jurassic Abenaki carbonate bank appears to have controlled the maximum seaward position of the shelf break throughout the Cenozoic and contributed to the development of a steep upper slope (Fig. 2) (Campbell and Mosher, 2010). In contrast, farther east the carbonate bank edge controlled the steepness of the paleoslope until the Late Miocene, after which the relief of the bank was healed over (Swift 1987). Seaward of the carbonate bank, a widespread erosional unconformity was formed in the Miocene and is tentatively correlated to horizon CS2 of Swift (1987) (Fig. 2). The unconformity is complex and developed through repeated erosion episodes, with the oldest recognized erosion forming the basal surface of a sediment drift (Campbell and Mosher, 2010). Swift (1987) suggested that CS2 was of lower to lower-middle Miocene age and equivalent to horizon Merlin from the U.S. continental margin dated at ~12 Ma (Tucholke and Mountain, 1986), although no direct seismic reflection tie of Merlin to the Scotian margin exists. Both CS2 and Merlin are attributed to erosion by bottom currents (Swift, 1987; Tucholke and Mountain, 1986), and the position of the horizon at the base of a sediment drift in the study area supports this interpretation. Two subsequent and extensive periods of mass wasting, as well as channel incision, further eroded the surface in the study area. The net result of these erosion events was the creation of a submarine embayment near the seaward edge of buried autochthonous salt deposits (Fig. 3A). In the study area, CS2 is onlapped by mass-transport and sediment-drift deposits.

DATA AND METHODS

Both 2D and 3D multichannel seismic-reflection data were interpreted for this study and integrated with biostratigraphic (Fensome et al., 2008) and geological (Petro-Canada Inc., 1985) information from the Shelburne G-29 exploration well (Fig. 1). A regional grid of 80 to 106 fold 2D seismic-reflection profiles was used to map the extent of depositional elements, interpret regional structures, and correlate seismic horizons into Shelburne G-29. These profiles were acquired by TGS-Nopec Geophysical Company in 1998 and 1999, with a line spacing of 8 km in the strike direction and 4 to 8 km in the dip direction. Hydrophone streamer length was 6 to 8 km, and the acoustic source was a 130 L tuned airgun array.

A large 3D seismic survey was used to map five key surfaces and to subdivide the drift deposits into sub-units. The surfaces were also used to study the geomorphological evolution of the drift deposits. The 3D seismic volume, termed the Barrington 3D survey, was acquired in 2001 by EnCana Corporation. The survey covers an area of approximately 1790 km². Data were recorded by eight 6000-m-long hydrophone streamers, each with 240 channels. Channel separation was 25 m, and bin spacing was 12.5 m. The acoustic source was a 62 L tuned airgun array generating a peak frequency of 70 Hz and a bandwidth of 5–100 Hz, implying about 5.5 m vertical resolution at water velocity. Processing of both datasets was conducted by the data owners prior to this study.

Seismic-reflection data were interpreted using SeismicMicro Technologies Kingdom Suite and Schlumberger GeoFrame software packages. For the 3D seismic dataset, reflection horizons were mapped and gridded to produce continuous surfaces with a grid cell size of 25 m. Seismic-reflection travel time, amplitude, and dip of maximum similarity attributes were used to interpret the seismic geomorphology. The dip of maximum similarity is a geometric attribute computed by first determining similarity (e.g., semblance) of adjacent traces over a sliding time window and range of dips, and then extracting the dip of maximum semblance within the time window. It is useful for identifying structural discontinuities. Plan-view dimensions of morphological features are given in SI units, and vertical dimensions, such as thickness and height, are given in milliseconds (ms) of two-way travel time (TWTT) or, in some cases, in meters using an interval velocity of 1800 m/s for conversion based on measured velocities for the interval from the Shelburne G-29 well.

RESULTS

The study interval is located between seismic reflection horizons A and E (discussed below) and comprises a buried succession of wavy to concordant reflections on the lower slope and rise.
Fig. 2.—A) Schematic dip-oriented cross section of Cenozoic deposits and older structural elements in the study area. The studied interval is highlighted in gray. Location of profile is given in Figure 1B. B) Dip-oriented seismic reflection profile and line interpretation through the central part of the study area. Note the major erosional unconformity and onlapping deposits which are the focus of this study (Horizons A–E). Location of profile is given in Figure 1C. Data are courtesy EnCana Corp. and TGS-Nopec.
interpreted to represent alternating deposits from bottom currents and sediment gravity flows (Figs. 2, 4–7). In the 3D survey area, the interval is up to 450 ms thick and thins to zero thickness upslope. The basal bounding surface of the succession is the top of a regional mass-transport deposit (MTD) and associated erosional unconformity (CS2), which together form a widespread marine onlap surface (Fig. 2; see also Campbell and Mosher, 2010). The upper bounding surface of the succession is defined by an abrupt transition to chaotic and transparent reflections interpreted as an interval of stacked mass-transport deposits. Five seismic horizons were correlated over the 3D seismic dataset and include the upper and lower bounding surfaces (Table 1). At Shelburne G-29, this interval correlates to upper Miocene to lower Pliocene strata (Fensome et al., 2008).

**Horizon CS2/Merlin**

The seismic geomorphology of horizon CS2 is presented in Campbell and Mosher (2010). CS2 is interpreted to be an erosional unconformity and appears as a strong, negative reflection that overlies a highly faulted interval (Figs. 2, 4). The unconformity is complex, formed by multiple, coalesced erosion surfaces within the area of the 3D seismic data. The seismic geomorphology of CS2 in the 3D seismic dataset shows evidence of pronounced downslope erosion (Fig. 8). In the southwestern corner of the 3D dataset, a long arcuate escarpment more than 220 m high corresponds to the headscarp of a large MTD. Immediately north and east of the escarpment, long, linear grooves up to 14 km long and 400 m wide are oriented down-dip and characterize much of the unconformity. Five domal structures are present on the CS2 surface and are the subsurface expression of underlying salt diapirs. Up-dip from the grooved zone, the seabed is scoured, with numerous smaller arcuate escarpments 0.5 to 4 km wide and 50 m high. This area passes up-dip to a zone that is much smoother with some subtle scars. The shelfward transition from erosional to conformable character of the surface is marked by a sharp change in gradient (Fig. 2). In the eastern part of the 3D seismic dataset, a channel eroded down to the level of CS2. The channel is up to 3 km wide and 370 m deep. The channel walls are scalloped, and a sinuous channel thalweg is apparent.

**Horizon A**

Horizon A forms the lower boundary of the succession examined in this study (Fig. 9). The horizon can be broadly divided into two zones: an upslope portion corresponding to the first peak reflection above the prominent trough of the underlying CS2 unconformity (described above), and a downslope portion that corresponds to the top boundary of the final mass-transport deposit associated with CS2. The morphology of the upslope portion of Horizon A is very similar to the underlying erosional surface where it directly drapes CS2 (Figs. 5, 8, 9), consisting of

**Fig. 3.—**Time structure maps of the regional Late Miocene to recent evolution of the western Scotian margin. **A** Time structure of the lower unconformity and the location of the seaward extent of autochthonous salt (from Shimeld, 2004) and a submarine embayment (Modified from Campbell and Mosher, 2010). **B** Time structure of the top of the sediment drift. Note the broad terrace. **C** Morphology of present seafloor (modified from Shaw and Courtney, 2004). Contour interval for all maps is 100 ms (ttw). Data are courtesy TGS-Nopec.
Fig. 4.—Series of three strike-oriented seismic profiles and line interpretations from the study area. Profile C–C’ is located just seaward of the break in slope associated with the top of the sediment drift and shows that gravity flows (gray stipple) at Horizon E are concentrated in sediment-wave troughs. Profile D–D’ is 7 km seaward of C–C’ and shows that gravity-flow focusing in wave troughs is less apparent. Profile E–E’ is 10 km seaward of D–D’ and shows little evidence of the thick, high-seismic-amplitude facies of the Horizon E complex. Note that in most cases sediment-wave migration direction is from west to east. Locations for profiles are given in Figure 1C and Figures 8–14. Data are courtesy of EnCana Corp.
Fig. 5.—Seismic-reflection profile and line interpretation F–F'. Profile is a dip-oriented line down the axis of sediment Fairway 1. The Horizon E “complex” is thickest above three steps formed by the inherited morphology of the underlying sediment drift and MTD. Note the apparent up-slope sediment-wave migration of type 2 bedforms. Location of profile is given in Figure 1C and Figures 8–14. Data are courtesy of EnCana Corp.

Fig. 6.—Seismic-reflection profile and line interpretation G–G'. Profile is a dip-oriented line down the axis of sediment Fairway 2. A perched basin was created by a large type 2 bedform that migrated up-slope and was filled by a thick succession of gravity-flow deposits (gray stipple). Location of profile is given in Figure 1C and Figures 8–14. Data are courtesy of EnCana Corp.
curvilinear and arcuate escarpments and elongated scours that developed during earlier mass wasting. The downslope portion of Horizon A is separated from the underlying CS2 by an interval up to 300 ms thick dominated by chaotic mass-transport deposits. The downslope morphology is rugose, due to the surface relief of massive blocks associated with the underlying MTD. This MTD is up to 280 ms thick, and individual blocks have lengths up to 4 km and relief up to 100 ms (Fig. 9). A large sidewall escarpment is preserved at the eastern margin of the MTD (herein referred to as the eastern escarpment) (Fig. 4A, B) and is over 20 km long with a maximum height of 150 ms.

In addition to relief due to gravity-flow processes, five buried salt diapirs create domal structures on this surface (Fig. 9C). Some of the salt doming is inferred to have been present at the time of
formation of the pre-drift surface, based on the subtle diversion of gravity-flow striations around diapirs (Campbell and Mosher, 2010). It is also possible that younger folding of strata above the salt diapirs could have produced this pattern.

**Horizon B**

Horizon B is a prominent reflection that onlaps Horizon A along the middle slope and also onlaps mass-transport deposits and salt diapirs where they have high positive relief (Fig. 2, 4). The isochron map of the Horizon A–B interval (Fig. 10C) illustrates that deposition was focused in the depression between the eastern escarpment and an area of blocky mass-transport deposition, leading to the development of a terrace. There was little or no deposition on the salt-diapir crests and areas east of the eastern escarpment.

The Horizon A–B interval shows early development of several sediment waves and the smoothing of relief associated with the basal Horizon A surface (Fig. 10). Much of the blocky and irregular relief apparent on the Horizon A surface is subdued or even absent at the Horizon B level. Sediment waves that are developed near the eastern escarpment (Figs. 4, 10) are continuous over distances up to 10 km, with up to 1 km of northeastward wave-crest migration between Horizons A and B. Wave amplitudes are up to 90 m, with an average wavelength between the two crests of 4 km. Wave-flank gradients are on the order of 5°. Sediment waves southwest of the salt diapirs are curvilinear, with wave crests that can be followed for up to 10 km, and average wavelengths of 3 km. These sediment waves have crests that bend northward around the northern curved flank of underlying diapirs (Fig. 10). Near the updip limit, where Horizon B onlaps Horizon A, additional smaller bedforms have an east–west crest orientation and an apparent south-to-north migration direction (Figs. 5, 10). The wave field in the southwestern corner of the study area overlying the buried mass-transport deposit is complex, with highly discontinuous wave crests with variable orientations.

**Horizon C**

Horizon C is a high-seismic-amplitude wavy reflection in the middle of the sediment-drift sequence (Figs. 2, 4). Horizon C appears to correspond to a time of maximum bedform relief

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**Table 1.—Seismic stratigraphy. Age estimates are from correlation to the Sheburne G-29 well and biostratigraphy from Fensome et al. (2008).**

<table>
<thead>
<tr>
<th>Horizon</th>
<th>Age</th>
<th>Geological Significance</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Late Miocene</td>
<td>Basal surface. Regional unconformity and top of associated MTD.</td>
</tr>
<tr>
<td>B</td>
<td>Late Miocene–Zanclean</td>
<td>Onset of sediment-drift development.</td>
</tr>
<tr>
<td>C</td>
<td>Zanclean</td>
<td>Middle of sediment-drift sequence.</td>
</tr>
<tr>
<td>D</td>
<td>Zanclean</td>
<td>Top of sediment-drift sequence.</td>
</tr>
<tr>
<td>E &quot;complex&quot;</td>
<td>Late Pliocene</td>
<td>Return to gravity-flow predominance.</td>
</tr>
</tbody>
</table>
across most of the study area. An isochron map between Horizons B and C shows that maximum deposition is located at the two large sediment waves west of the eastern escarpment and immediately west of the salt diapirs (Fig. 11C).

Both the direction and the magnitude of wave-crest migration vary between Horizons B and C. Near the eastern escarpment, the dominant migration direction is to the northeast, with up to 2 km of wave-crest migration near the eastern escarpment (Fig. 4). Interestingly, at least one wave crest migrated in the opposite direction, towards the southwest (Fig. 11E) and is likely caused by changing current parameters as the bottom current interacted with the eastern escarpment and the buried salt diapir to the south. Variations in the amount of wave-crest migration on most sediment waves also result in a subtle change of orientation of the wave crests at Horizon C compared to Horizon B. Wave amplitudes, particularly near the eastern escarpment, increased significantly from < 90 m at Horizon B to > 150 m at Horizon C, with the steeper flanks of the bedforms showing a sharp increase in gradient to 9°.
Sediment waves southwest of the salt diapirs continue to develop in a curvilinear pattern around the diapirs. At Horizon C, these sediment waves are linked to low-amplitude two-dimensional waves north of the diapirs that trace up dip to where they intersect or transition to east-west-oriented waves (Fig. 11). East-west-oriented waves are more common at the Horizon C level than Horizon B and have continuous crests for up to 5 km. In the area overlying the buried MTD in the southwest, many of the discontinuous bedforms observed at Horizon B have become linked, forming larger and more organized bedforms with curvilinear to circular planform shapes (Fig. 11).

**Horizon D**

Horizon D (Fig. 12) coincides with a decrease in sediment-wave growth and migration (Figs. 4–6). The horizon C–D interval shows evidence of sediment-wave migration and growth, as well as some draping and ponding in bathymetric lows (Fig. 4). The isochron map of the interval between horizons C and D (Fig. 12C) illustrates that sedimentation was greatest along the crests of large sediment waves in the eastern part of the study area, in contrast to the western part of the study area, where sediment draping dominated in the wave troughs and little deposition occurred at wave crests.

Along the eastern escarpment, compared to Horizon C, the waves are of lower amplitude and there is bifurcation of one of the wave crests (Fig. 12). Large sediment waves continue to develop southwest of the salt diapirs and the crests of the east-west-oriented waves are more continuous, reaching lengths up to 10 km. Above the buried MTD in the southwest, two large curvilinear sediment waves are apparent.

Time structure maps show that cumulative deposition of the sediment drift resulted in the construction of a bathymetric step in the southern half of the 3D survey area (Fig. 12B). The step extends up to 50 km seaward, forming a prominent regional terrace (Fig. 3B).

**Horizon E Complex**

Over much of the study area, Horizon E consists of a single high-positive-amplitude reflection; however, in some paleobathymetric lows, the horizon splits into two or more high-positive-amplitude reflections (Figs. 4–7). Horizon E marks an abrupt change in the seismic facies character and geomorphology within the study interval. Much of the wavy and undulating
nature of the underlying sediment drift was healed over at the Horizon E level, with only sediment waves that are adjacent to salt diapirs still present (Fig. 13A). A seismic-reflection line along strike to the regional slope near the foot of the ramp shows that deposition of the Horizon E complex is focused in sediment-wave troughs (Fig. 4A).

The Horizon D–E interval generally lacks the internal sediment-wave geometry of underlying deposits. The interval is divided into two parts: a lower low-amplitude and reflection-free section that immediately overlies Horizon D, and a high-amplitude section assigned to the Horizon E complex (Figs. 4–7). An RMS amplitude map of the Horizon D–E interval (Fig. 13D) and the amplitude map of Horizon E show several amplitude anomalies that have ribbon to sub-ellipsoid shape (Figs. 13D, 14). They generally have a downslope orientation, with trajectories that avoid the crests of salt diapirs. Highest seismic amplitudes are located at the foot of the steep ramp at the up-dip limit of the sediment drift. Deposits associated with the Horizon E complex...
are thickest and have the greatest amplitude in the troughs of bedforms in the underlying sediment drift (Figs. 4, 14).

Two depositional fairways are recognized from the seismic-amplitude maps (Fig. 14) and seismic-reflection data (Figs. 4–6). These are interpreted to mark the transition back to gravity-driven deposition with sediment transport dominantly oriented in the downslope direction. Fairway 1 is the westernmost depositional fairway. It changes orientation downslope, from a southeast to a southwest direction (Fig. 14). An axial dip profile down the fairway reveals deposits of variable thickness between horizons D and E (Fig. 5). In plan view on the RMS amplitude map (Fig. 14), the fairway becomes a narrow, < 500-m wide ribbon downslope where it intersects profile E–E’ (Fig. 4C). Fairway 2 is located in the central part of the study area, and its plan-view geometry appears similar to Fairway 1 (Fig. 14). Like Fairway 1, Fairway 2 changes orientation downslope, from southeast to southwest. An axial dip profile down Fairway 2 shows that the Horizon E complex bifurcates into a number of reflections immediately updip of a large sediment-wave crest (Fig. 6).

Fig. 12.—Seismic geomorphology of Horizon D. A) Shaded-relief map. B) Time structure contours (50 ms (TWTT) contour interval). C) Isochron map of the Horizon C–D interval. D) Isochron map of the entire sediment drift (Horizon A–D interval). Data are courtesy of EnCana Corp.
In addition to the two depositional fairways, two other depocenters are recognized (Figs. 4, 14). In the extreme western part of the study area, a moderate to high-amplitude lobe is present within a depression formed by the erosion and deposition of the large MTD associated with Horizon A (Fig. 14). The downslope limit of the high amplitude is controlled by a ridge formed by a sediment-wave crest. Downslope from the ridge, the Horizon D–E interval consists of lower-amplitude reflections. In the eastern part of the study area, a large depression located within the trough of a sediment wave is a second depocenter (Fig. 7). The depression fill (Horizon E complex) thins upslope and downslope and onlaps the flanks of the sediment wave. Reflection amplitudes within the fill show evidence of channel development, with meandering channel forms and high-amplitude sheets (Fig. 7).

**DISCUSSION**

**Bedform Classification**

The geomorphology of Horizons A through D is dominated by sedimentary bedforms that are broadly divided into three types (Fig. 11A, B). Type 1 bedforms are most common and...
comprise two-dimensional sediment waves with crests oriented northwest–southeast (Fig. 11A). Their crests extend up to 20 km in length, their wavelengths are 2–4 km, and their amplitudes are up to 150 m. Wave amplitude is greatest immediately west of abrupt negative changes in seafloor relief, such as west-dipping escarpments and west-dipping diapir flanks. Type 2 bedforms are two-dimensional sediment waves with crests oriented east-west. They are confined to areas of steeper slopes and gullies updip of Type 1 bedforms. Type 2 bedforms have wave crests that typically extend for 5 km, wavelengths of 3 km, and wave amplitudes up to 75 m (Fig. 11A). Type 3 bedforms are circular to curvilinear bedforms located on the west side of three closely spaced salt diapirs in the 3D seismic survey area (Fig. 11A–C). This bedform type is less organized than Types 1 and 2.

Wave migration through the Horizon B–C interval is interpreted from comparison of changes in wave-crest position at the Horizon B and C level, as well as the intervening asymmetrical and sinusoidal seismic reflection geometries. Most Type 1 bedforms migrate from west to east (Fig. 4). An exception includes the second wave crest west of the eastern escarpment,
which appears to migrate from east to west along part of the bedform (Fig. 11E). Additionally, wave-crest migration can vary along an individual wave, for example, along the large wave immediately west of the eastern escarpment. On this bedform, the wave appears to have remained stationary at the southern end of the wave and migrated over 1 km at the northern end of the wave crest (Fig. 11E). Type 2 bedforms migrate up slope, as revealed by arbitrary seismic profiles perpendicular to the wave crest (Figs. 2, 5, 6). Migration direction of type 3 bedforms is highly variable, from laterally climbing bedforms to vertically aggrading bedforms (Fig. 4).

**Alternation of Depositional Modes and the Effects of Inherited Morphology**

**Widespread Slope Failure and Mass-Transport Deposition.**

Horizon CS2, a regional deepwater unconformity, formed in the Middle to Late Miocene in the study area through periods of major slope failure and, to a lesser extent, bottom-current erosion and channel incision (Campbell and Mosher 1993). Gravity-driven mass wasting had a marked effect on subsequent depositional patterns. The erosional surface modified the pre-existing slope morphology and created a broad ramp on the middle to lower slope. The steep, upper portion of the ramp created a bypass zone that persisted for millions of years and likely restricted the accumulation of gravity-flow deposits to areas farther seaward. The last phase of slope failure resulted in the formation of the eastern escarpment, which controlled the development of large sediment waves to the west.

**The Development of a Sediment Drift.**

On a regional scale, the construction of a sediment drift contributed to the infilling of the submarine embayment related to the erosion of CS2 (Fig. 3A) and created a broad terrace up to 50 km wide (Fig. 3B), forming an abrupt break in slope where the drift onlaps Horizon A (Fig. 2). The drift was constructed through the stacking of sediment-wave packages and, within the area of 3D seismic data, reached a maximum thickness of ~380 ms (Fig. 12D). The drift appears to have grown from the deeper part of the basin and backstepped up the slope rather than developing a plastrered component over the steep ramp. This architecture suggests that some of the sediment constituting the sediment-wave field was likely pirated from turbidity currents that would have bypassed the ramp, only to perhaps become entrained in a bottom current on the rise, forming sediment waves. Such a process has been suggested by He et al. (2008), and the formation of muddy sediment waves would require a bottom current with a velocity of only 0.05–0.25 m/s (Stow et al., 2009). Alternatively, the backstepping and onlapping architecture may have resulted from bottom-current acceleration across the ramp, preventing deposition, although there is no evidence of along-slope erosion on the ramp.

The initiation process for fine-grained sediment waves generated by bottom currents is not well understood (Flood, 1988; Blumskack, 1993; Hopfauf and Spieß, 2001; Wynn and Stow, 2002), but in general it is attributed to the presence of seafloor irregularities which cause the required bottom-current velocity perturbations needed for wave development (Hopfauf and Spieß, 2001; Blumskack, 1993). Detailed analysis of the Horizon A–D interval in this study provides some insight into deep-water sediment-wave-forming processes. Seafloor irregularities at Horizon A resulted from the surface expression of five salt diapirs, and rugose erosional and depositional features (escarpments, blocky MTDs) associated with seabed failure (Fig. 9). Sediment-wave development is greatest immediately southwest of these features (Figs. 10–12). On basin floors, wave crests are usually oriented perpendicular to current direction (Wynn and Stow, 2002). For the study area, this fact implies a northeast to southwest bottom-current flow direction, approximately perpendicular to the regional slope. On steeper slopes at the northern limit of the sediment-wave field, Type 2 bedforms migrate in a north to a northwest direction. This observation implies that sediment waves on slopes migrate up slope and up current (cf. Wynn and Stow 2002). The geomorphology of Horizon C shows an abrupt transition from Type 1 to Type 2 bedforms, apparently related to an increase in the gradient of the underlying seabed (Figs. 6, 11). Type 3 bedforms exhibit complicated geometries that likely result from bottom currents flowing over complex topography. For example, bedform geometry around the salt diapir shown in Figure 11D suggests flow separation and formation of lee vortices on the downcurrent side. This phenomenon was modeled for fluid flow around three-dimensional objects (Smolarkiewicz and Rotunno, 1989), and is known to occur in alluvial channels (Best and Brayshaw, 1985).

The sediment drift is not sampled in the study area. Fifty kilometers east of the study area at the Sherburne G-29 well, sidewall cores of age-equivalent strata are composed predominantly of dark-gray brown to dark gray-green claystone (Petro-Canada Inc., 1985), however, sediment waves are not present at the well site. Biostratigraphic data at the Sherburne G-29 well (Fensome et al., 2008) shows that the drift formed during the Late Miocene to Middle Pliocene and is therefore of age similar to that of a large sediment-wave field on the New Jersey continental rise (Pong and Mountain, 1987). Off New Jersey (e.g., Deep Sea Drilling Program site 603), the sediment-wave interval consists of dark grayish green hemipelagic claystone, with entrainment of fine-grained turbidity currents and hemipelagic sediments in a southwest-flowing bottom current suggested as the mechanism for formation of the sediment-wave field (Locke and Laine, 1992).

Miocene and Pliocene sediment drifts along the Scotian margin are underreported in the literature. Apart from identification of small zones of sediment waves (e.g., Swift, 1987; MacDonald, 2006), the role of bottom currents in shaping the margin has focused on recognition of seismic-reflection horizons attributed to bottom-current erosion (Swift, 1987; Ebinger and Tucholke, 1988). During construction of the sediment drift in this study, paleo-current direction indicators suggest that mean bottom-current flow was from the northeast to the southwest. It therefore appears likely that the sediment drift was deposited under the influence of a western boundary current. Using the sediment-drift categories defined by Rebesco (2005), the sediment drift in this study could be defined as a sheeted or infill drift, based on its low relief, large area, and the observation that it fills the submarine embayment formed by the erosion of CS2.

**Return to Gravity-Flow Deposition.**

The Horizon E "complex" is interpreted to represent a transition back to gravity-flow-dominated deposition, based on a change in reflection character and geometry at this level. The complex is correlated on 2D seismic data to the Sherburne G-29 well, where it corresponds to an ~50-m-thick interval of interbedded, well-sorted, fine-grained sandstone and claystone (Petro-Canada Inc. 1985). The greatest seismic amplitudes of Horizon E, in terms of both magnitude and area, are located just seaward of the break in slope (Fig. 14). Seismic-reflection onlap relationships and thickness variations suggest that gravity-flow deposits pref-
erentially filled sediment-wave troughs. This observation is most apparent proximal to the break in slope (Fig. 4). In Fairway 1, gravity-flow deposits form a series of three elongated, stepped basins (Fig. 5). Sills were formed by inherited relief from sediment-wave crests and the buried irregular MTD surface. The fairway was laterally confined either by position within wave troughs or by structural highs related to salt diapirs. In Fairway 2, a perched basin was formed by a large Type 2 sediment wave at the break in slope (Figs. 6, 14). In this example, a Type 2 wave crest created a sill, and gravity-flow deposits accumulated behind the sill until it was breached. In the large sediment-wave trough in the eastern part of the study area (Fig. 7), repeated gravity flows filled the depression, creating a high-amplitude wedge. A similar model for formation of ponded basins by sediment-wave morphology was suggested by Viana (2008).

There are several places in the study area where large depressions at the Horizon D surface were filled with low-seismic-amplitude deposits instead of the high-amplitude Horizon E complex (Fig. 4). The precise position along slope where gravity flows bypassed the ramp and reached the terrace was controlled by factors farther up slope outside the study area (e.g., the position of failures or turbidity currents generated at the shelf edge or upper slope). Once flows reached the break in slope, presumably they accumulated in the closest depression, which, once filled, overflowed into the next depression downslope in a cascading manner. Large depressions on the Horizon D surface that were not connected to gravity-flow inputs were filled with lower-seismic-amplitude deposits of the Horizon D–E interval. Therefore, it appears that proximity to the input point of gravity flows, proximity to the base of slope, and connectivity between adjacent depressions were more important factors than depression size for accumulating the high-seismic-amplitude deposits associated with the Horizon E complex.

SUMMARY AND CONCLUSIONS

In the study area, the seismic stratigraphy reveals abrupt changes in depositional styles and sedimentary processes. Initially, seabed failure and deposition of the large MTD associated with the pre-drift surface represented rapid erosion of the middle to lower slope and sediment transfer into the deeper basin. This period was followed by sediment-drift development and construction of a large terrace at the foot of slope. Finally, an apparently sudden transition to gravity-flow (turbidity current) input occurred that was contemporaneous with a termination of sediment-wave development, suggesting that either bottom-current intensity decreased or gravity-flow input overwhelmed the bottom-current signal.

Within the study area, inherited morphology affected subsequent deposition patterns for both along-slope and down-slope processes. Regionally, the sediment-wave field developed within a submarine embayment created by mass wasting and channel incision. Locally, seafloor irregularities due to mass wasting and protruding salt diapirs controlled the precise locations of maximum sediment-wave growth. With the return to gravity-flow-dominated deposition, the highest seismic amplitudes occur at the break in slope formed by the drift terrace. At the break in slope proximal to gravity-flow input points, local topographic features at the scale of individual sediment waves controlled the distribution of gravity flows.

The studied interval on the western Scotian margin represents alternation from gravity-flow-dominated to bottom-current-dominated to gravity-flow-dominated deposition. The results demonstrate the importance of morphological heritage in controlling subsequent deposition patterns. The morphologies of the bedforms observed in the study area provide insights into the hydrodynamics of bottom currents and sediment-wave-forming processes. With the onset of gravity-flow deposition, the regional terrace formed by the sediment drift effectively trapped gravity flows that may otherwise have been transported to deeper parts of the basin. Locally, sediment-wave morphology strongly influenced the locations of gravity-flow conduits and depocenters.

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