The Holocene shallowing-upward parasequence of north-west Andros Island, Bahamas

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ABSTRACT

Many pre-Mesozoic records of Earth history are derived from shallow water carbonates deposited on continental shelves. While these carbonates contain geochemical proxy records of climate change, it is the stratal architecture of layered carbonate units that often is used to build age models based on the idea that periodic astronomical forcing of sea-level controls the layering. Reliable age models are crucial to any interpretation of rates and durations of environmental change, but the physical processes that actually control this stratal architecture in shallow water carbonates are controversial. In particular, are upward-shallowing stacks of carbonate beds bounded by flooding surfaces (‘parasequences’) truly a record of relative sea-level change? The purpose of this study is to examine a tidal flat that is actively accumulating carbonate stratigraphy, and to determine the relative importance of tidal channel migration (poorly known, but investigated here) and Holocene sea-level rise (well-known) in controlling post-glacial parasequence architecture. This work represents a field study of peritidal carbonate accumulation at Triple Goose Creek, north-west Andros Island. By integrating surface facies maps with differential global positioning system topographic surveys, a quantitative relationship between facies and elevation is derived. Sedimentary facies are sensitive to elevation changes as small as 5 cm, and are responding to both internal (distance to nearest tidal channel) and external (sea-level rise) controls. The surface maps also are integrated with 187 sediment cores that each span the entire Holocene succession. While flooding of the Triple Goose Creek area should have occurred by ca 4500 years ago, preservation of Holocene sediment did not begin until 1200 years ago. The tidal channels are shown to be stationary, or to migrate sluggishly at up to 6 cm per year. Therefore, while the location of tidal channels is responsible for the modern mosaic of surface facies, these facies and the channels that control them have not migrated substantially during the ca 1200 years of sediment accumulation at Triple Goose Creek. Once the region was channellized, vertical and lateral shifts in facies, such as the landward retreating shoreline, expanding mangrove ponds and seaward advancing inland algal marsh, are driven by changes in relative sea-level and sediment supply, not migrating channels. While stratigraphic columns look different depending on the distance to the nearest tidal channel, the overall parasequence architecture everywhere at Triple Goose Creek records an upward-shallowing trend controlled by the infilling of accommodation space generated by post-glacial sea-level rise.

Keywords Carbonate, cyclostratigraphy, Holocene, tidal channel, tidal flat.
INTRODUCTION

The sedimentology, physical stratigraphy, palaeobiology, geochemistry and palaeomagnetism of carbonate rocks record information about palaeoclimate, palaeoceanography and palaeogeography. In order to interpret this information, one must find a way to tell time in the sedimentary host rock, and thus calculate rates. Without rates, it is not possible to constrain the viable physical, chemical or biological processes responsible for the creation of the geological record. Unfortunately, the dearth of high resolution absolute chronometers in carbonate rocks and the low preservation potential of volcanic ashes in shallow water settings makes it difficult to jump from qualitative description of environmental change to quantitative understanding of process.

The apparent cyclic arrangement of lithological units has been documented in a variety of palaeoenvironmental settings, ranging from fluvial to tidal flat to deep sea. In peritidal carbonates, which represent the most abundant and best preserved records of Earth history in pre-Mesozoic rocks, metre-scale (sometimes referred to as fifth order) upward-shallowing parasequences are ubiquitous (Laporte, 1967; Read, 1973; James, 1984; Grotzinger, 1986a; Sarg, 1988; Koerschner & Read, 1989; Pratt & James, 1992; Goldhammer et al., 1993; McLean & Mountjoy, 1994; Lehmann & Goldhammer, 1999). Upward-shallowing metre-scale parasequences come in many flavours, but often are composed of, from bottom to top: (i) a basal packstone or winnowed-lag overlying a ferruginized and brecciated hardground or subaerial exposure surface; and (ii) a condensed to non-existent transgressive unit culminating in (iii) a maximum flooding surface/zone within shale or carbonate mud rock. The maximum flooding surface is followed by the upward-shallowing portion of the parasequence composed of some combination of carbonate (iv) wavy laminite and/or wackestone, (v) stromatolite and/or packstone, (vi) grainstone and (vii) microbialite, capped by another hiatus or brecciated subaerial exposure surface.

Numerous models have been proposed to link the origin of these lithological cycles to periodic or aperiodic changes in relative sea-level of various time scales (Goodwin & Anderson, 1985; Grotzinger, 1986b; Read et al., 1986; Goldhammer et al., 1990). However, some have questioned the paradigm that repeated metre-scale lithological parasequences in shallow water carbonates are controlled by a periodic sea-level forcing (Cloyd et al., 1990; Drummond & Wilkinson, 1993; Wilkinson et al., 1997, 1999; Burgess, 2001, 2008). For example, Markov chain analysis of lithofacies transitions from a number of Palaeozoic carbonate successions suggested that metre-scale parasequences record potentially random upsection lithofacies transitions (Wilkinson et al., 1997). The thickness-frequency distributions of these same carbonate successions include an exponential decrease in unit frequency with linear increase in unit thickness (Drummond & Wilkinson, 1993; Wilkinson et al., 1999). This type of exponential distribution was interpreted to be characteristic of Poisson processes, where each lithological unit is an independent event (i.e. it has no memory of prior deposition). However, in a more recent analysis, 40 out of 56 carbonate successions analysed showed systematic deviations from an exponential distribution (Burgess, 2008). Furthermore, recognizing a specific statistical distribution alone does not uniquely identify the processes responsible for the stacking pattern. Nevertheless, obvious examples of non-periodic processes that could control lithofacies stacking on peritidal carbonate platforms include the avulsion (Jerolmack & Paola, 2007) and migration of sinuous channels. In fact, Cloyd et al. (1990) reported sedimentary evidence for laterally migrating carbonate grainstone channel deposits within peritidal carbonates.

Are metre-scale parasequences telling us about subsidence, eustasy or both? Or are metre-scale parasequences recording a complex mosaic of facies controlled by proximity to migrating and avulsing tidal channels?

Triple Goose Creek (TGC), on the north-west edge of Andros Island in the Bahamas (Fig. 1), has long captured the imagination of carbonate sedimentologists because it is a rare Holocene example of the type of carbonate environment with a well-developed channellized intertidal zone (Fig. 2) that may have been more globally abundant during much of Earth history. The major physiographic and facies elements of TGC were first described in pioneering work more than 40 years ago (Shinn et al., 1969; Gebelein, 1974; Hardie, 1977). This study extends their work by quantifying the relationship between facies and elevation, and by integrating surface observations of modern processes with the subsurface record of Holocene sedimentation. Given the well-documented record of Holocene sea-
level change (Flemming et al., 1998; Toscano & Macintyre, 2003; Milne et al., 2005) and assuming slow and steady thermal subsidence (0.1 mm year$^{-1}$) over the past 10 000 years, the Holocene sediments of TGC are used to determine the relative importance of channel migration on the stratigraphic architecture of peritidal carbonate parasequences.

GEOLOGICAL SETTING

The Bahamas archipelago (Fig. 1A) is a carbonate depositional environment consisting of several isolated platforms at the southern extremity of North America’s eastern continental margin. Andros is the largest island in the Bahamas, located on the raised eastern rim of the western
Fig. 2. False-colour (4,3,2) pan-sharpened Quickbird satellite image acquired 27/12/2003 by DigitalGlobe. The UTM grid is WGS84 and axis labels are in metres. Cores are located with green circles, and core-transect figures are keyed to numbers. Six tide gauges are located at the numbered triangles keyed to Fig. 10. The DGPS base station is located at the black star. Physiographic elements of Triple Goose Creek (Shinn et al., 1969; Hardie, 1977) include, from west to east, the subtidal Great Bahama Bank (blue), the central intertidal channelized zone, and the supratidal inland algal marsh (brown).
half of the Great Bahama Bank, and bounded by seaways to the east (Tongue of the Ocean) and west (Straits of Florida) (Fig. 1A to C). The positive relief of Andros island is composed of Pleistocene carbonate dunes (Ball, 1967; Carew & Mylroie, 1995a; Brooke, 2001) overlying karsted and cemented Pleistocene coral reef (for example, Morgan’s Bluff; Carew & Mylroie, 1995b). Andros Island has a tropical-maritime climate, with wet summers, dry winters and fairly uniform monthly average temperatures of ca 25°C. Annual average rainfall is ca 130 cm year⁻¹, making Andros significantly wetter than other Holocene peritidal carbonate environments such as the Abu Dhabi region (annual average rainfall ca 2 cm year⁻¹). Prevailing easterly trade winds are strongest and most variable in March and April. Strong winds frequently control water height in the bays and channels of north-west Andros, overwhelming the ca 60 cm signal from the semi-diurnal lunisolar tide.

Triple Goose Creek is located on the north-western edge of Andros Island (Fig. 1B). Most of the sediment delivered to the intertidal and supratidal zones of TGC is mud derived from aragonite-producing codiacan algae such as *Halimeda*, *Penicillus* and *Udotea* (Stockman et al., 1967). CaCO₃ production is greatest on the outer edge of the Great Bahama Bank, where cold supersaturated open-ocean waters spill onto the bank and warm. Production is lower on the inner parts of the bank, with CaCO₃ production decreasing as water residence time increases to >250 days adjacent to TGC (Broecker & Takahashi, 1966; Morse, 1984; Demicco & Hardie, 2002). Aragonite mud is transported shoreward and deposited in the intertidal zone by wind and tide driven currents. While gastropods, foraminifera and bivalves inhabit the mud flats, the only skeletal sands (<1 mm broken shell fragments) in the entire system are found on beaches and on channel bottoms. Very fine carbonate sand can be found on levée crests within 500 m of the shore and in channel bars. Without a riverine flux of siliciclastic sediment, the only significant quartz in the TGC system is derived from plumes of dust originating in Saharan West Africa (Moore et al., 2002).

**TRIPLE GOOSE CREEK PHYSIOGRAPHY**

On the landward, eastern side of TGC, a flat, mostly supratidal 3 to 8 km wide inland algal marsh (IAM) is characterized by luxuriant tufted filamentous cyanobacterial mat communities, colloquially referred to as *Scytonema*, and desiccated carbonate mud flats left by storm surges (Fig. 9). The stratigraphic thickness of Holocene mud increases seaward, from emergent cemented Pleistocene dune sands in the east to 1 m of Holocene mud along the western fringe of the IAM.

A central intertidal zone, 1 to 4 km wide, is characterized by tidal channels containing very distinctive depositional bars designated here as ‘butterfly bars’. Channel thalwegs frequently rest on bedrock and are littered with sand and broken shells. Unlike point bars in meandering channels, butterfly bars in TGC form in the straight channel reaches between bends, and reside in the middle of the channel, allowing tides to flush flood and ebb discharges around opposite sides of the bar; their name derives from their shape in plan view, similar to the wings of a butterfly (Fig. 3). Bars are dominated by bioturbated mud, but also contain thin layers of fine sand, and rare lenses of skeletal sand and broken shells.

With increasing distance normal to channel margins, elevation decreases from levée crests (LC) to levée backslopes (LBS) to high (HAM) and low (LAM) algal marshes to shallow mangrove ponds (MP) (Figs 4 and 11). Levée crests are draped with thin (0·1 to 1·0 mm) flat mats formed by filamentous cyanobacteria colloquially referred to as *Schizothrix*, display a low diversity of grasses dominated by a species colloquially referred to as ‘sawbush’, contain eroded black mangrove root systems and rarely are bioturbated (Fig. 6). Levée crests most proximal to the seaward beach ridges (transitional and coloured the same in Fig. 4) receive coarser silts and fine sands as overbank deposits during large tides and storms, and may be rippled and mudcracked (Fig. 5B). The levée backslope is similar to the levée crest, but is marked by the first appearance of *Monanthochloe littoralis* ‘sawgrass’, and an increase in grass diversity, black mangrove roots and bioturbation (Fig. 6).

High and low algal marshes are dominated by *Scytonema* filamentous cyanobacterial tufted mats (Fig. 7B), have a higher diversity of grasses and a greater abundance of mangroves and are only weakly bioturbated by *Morphysa* worms (Fig. 7C). Mangrove ponds contain abundant mangroves (Figs 7A and 8A) and are colonized by a low diversity assemblage of *cerithid* gastropods (Fig. 7D), *Geloina* bivalves, *Peneroplis* foraminifera (Fig. 8B), and burrowing worms and crabs. High ponds surround the tips of first-order

Fig. 3. ‘Butterfly bars’ (red stars) are bright white accumulations of carbonate mud, broken shells and skeletal sand within the channels (dark blue). Like fluvial point bars, butterfly bars are regions of sediment deposition associated with sinuous channels. However, unlike classical point bars that form crescent-shaped deposits attached to the apex of the lobate inner parts of sinuous bends, butterfly bars form elliptical to rhomboidal shapes in the centre of straight segments of channels flanking cuspatate inner bends. Butterfly bars are asymmetrical in shape and distance from the channel banks, and usually reach the elevation of mean tide level (MTL). In rarer circumstances, parts of butterfly bars reach above MTL, are stabilized by mangrove trees, and become isolated from the active bar by cross-cutting channels (E). The UTM grid is WGS84 and axis labels are in metres.

streams and contain nearly homogeneous white mud with little organic matter. Low ponds are similar to mangrove ponds in their assemblage of shelly fauna, but lack abundant mangrove roots. Deep ponds are too deep to explore on foot, but appear similar to low ponds with an even more depauperate fauna.

METHODS

Digital topographic survey

This study involved five weeks of centimetre-scale topographic surveying coupled with surface geological observations of sediment composition, sedimentary structures, and the distribution of microbial mats and vegetation. Elevation data were collected using the combination of a Leica Total Station TCR1203 with Leica GPR121 circular prism (Leica Geosystems Inc., Norcross, GA, USA) and a Trimble 4700 Differential Global Positioning System (DGPS) system with Trimmark Ile base radio modem (Trimble Navigation Limited, Sunnyvale, CA, USA). Survey tripods for both instruments were stabilized by placing tripod legs on 40 cm long sections of 3.8 cm diameter PVC pipe that had been driven 95% into the mud. Over the course of three field seasons, ca 70 000 survey points were collected (Fig. 4).

In order to determine relative positional accuracy and reproducibility of the spatial dataset, 10
specific landmarks for each station were resurveyed once or twice a day for the duration of the project. The resurvey measurements were distributed normally with 1σ errors averaging 1±4 mm east–west, 1±5 mm north–south and 1±5 mm in elevation for the total station in reflector mode, and 1±9 cm east–west, 2±1 cm north–south and 4±1 cm in elevation for the DGPS.

In the absence of nearby benchmarks, all survey data were referenced back to the DGPS base station (Figs 4 and 13A). The absolute accuracy (in world map coordinates) of this control point is estimated to be better than 20 cm east–west and north–south and better than 40 cm in elevation. While relative positional accuracy is crucial to this study, the absolute accuracy does not affect the results. The elevation data are further calibrated by taking z = 0 cm as the mean tide level (MTL), observed over the course of six days at the DGPS base station in March 2005.

For each survey and core-transect, MTL was determined by planting a graduated rod in the channel and monitoring water level throughout the day. Core-top elevations were measured with respect to this local tide level at specific times and then calibrated to the tidal cycle at the DGPS station. It was assumed that there was no significant tidal lag along the channel reach (see below).

**Tidal range**

Mean water level and amplitude of change at TGC are a complex function of the lunisolar semi-diurnal tide and the strength and direction of the wind. For example, it is not uncommon to have a week without high tide when strong easterly winds settle in. To better characterize the tides, six YSI Level Scout pressure transducers (1 mm accuracy; YSI Incorporated, Yellow Springs, OH, USA) were deployed to measure the relative timing and amplitude of the tide at six locations along a single channel network (Fig. 2). The tide gauges recorded water pressure and temperature every 2 sec. Water depth was calculated after a correction for local variation in air pressure recorded by a Davis Instruments Vantage Pro2 weather station (Davis Instruments Corporation, Hayward, CA, USA) setup at tide gauge station #2 (Fig. 2).

The tides are virtually identical in timing and amplitude all the way from a site near the creek mouth on the open bank (1), to three sites along 2±5 km of a single fourth-order channel (2, 3 and 4), to two sites in second-order streams (5 and 6), suggesting that the elevations measured along different core-transects at different times generally are compatible (Fig. 10). The survey indicates an average tidal range of 0±61 m, with 0 to 6 cm of damping upstream over 2±5 km. Average time lag varies from 7 to 15 min, with lags up to 15% longer during ebb tide, and generally longer for sites further east (Fig. 10). At all sites, the tides are strongly asymmetrical in duration, with the flood tide lasting 4±7 ± 0±4 h, and ebb tide lasting 7±6 ± 0±4 h (Fig. 10). This asymmetry is typical in frictionally dominated convergent estuaries.
(Wells, 1995; Lanzoni & Seminara, 2002) and leads to net landward transport of sediment.

**Subsurface geology**

A 1 cm diameter stainless steel probe was used to measure the depth to Pleistocene basement rock (cemented limestone grainstone) at a 566 point subset of survey points. One hundred and eighty seven cores, 5 cm in diameter and ranging from 0·2 to 3·4 m in length, were extracted using a stainless steel and aluminium Livingstone piston corer from the University of Minnesota Limnology laboratory. Extruded cores were split and...
logged at 0·1 to 5 cm vertical resolution using the simplified carbonate facies described in Fig. 13. Based on observations from modern channel bottoms, broken shells and sand in cores were used to infer ancient channel locations.

**Radiocarbon dating**

Benthic *Peneroplis proteus* (Fig. 8B) are the most abundant foraminifera species in both TGC pond environments and the near shore subtidal regions off north-west Andros Island. Foraminifera were picked from core samples and surface channel and pond sediments. The foraminifera were sonicated in a 3% peroxide and sodium hexametaphosphate solution and rinsed with DH₂O to clean the tests. Three different 4 mg replicates from each of eight foraminifera collections were run at the National Ocean Sciences Accelerator Mass Spectrometry (NOSAMS) Facility for ¹⁴C ages and δ¹³C. The three replicates from each collection yielded a narrow 0·1 to 0·5‰ range in δ¹³C (compared with a 3·5‰ range in δ¹³C from 200 individual foraminifera specimens from surface environments within TGC), and a 20 to 100 year range in ¹⁴C ages. ¹⁴C ages were converted to calendar ages using the Marine04 calibration (Hughen *et al.*, 2004). Finally, calendar ages were adjusted for the approximate regional difference from the average global reservoir age (*Stuiver* & *Braziunas*, 1993) in the Bahamas of ΔR≈96 years determined from sites 294 and 297 (http://calib.qub.ac.uk/marine/, 12 December 2010). Calendar ages are reported with 1σ errors in Figs 13 to 20.

**RESULTS**

**Surface geology**

Most, if not all of the facies elements are deposited in an overlapping range of water depths (Fig. 11), leading to lateral equivalence of facies and a complex mosaic of facies elements (Fig. 4). Nevertheless, the mean elevation of each facies is statistically distinct, and facies elevations tend to decrease with increasing distance down the back of the levée (Fig. 11). These results are consistent with Markov chain analysis depicting highly ordered lateral facies transitions (*Rankey*, 2002). The topographic relief that defines this geobiological zonation in the intertidal zone ranges from 0·41 m above mean tide level (MTL) for high (75th percentile) levée crests to 0·10 m above MTL for high (75th percentile) mangrove ponds, to 0·11 m
below MTL for some of the deeper (25th percentile) low ponds, to 1.5 m below MTL for some of the deeper channel bottoms (Fig. 11). As Shinn et al. (1969) and Ginsburg et al. (1977) have noted, facies are sensitive to changes in elevation as small as 5 cm (Fig. 11).

Other variables such as sediment supply (Wright, 1984; Rankey, 2004) also affect the distribution of facies. Under many circumstances at TGC, sediment supply is a function of the distance from the nearest channel and the size (i.e. stream order) of that channel. The average distances between different facies elements and channels of different stream orders form fairly complex populations that are not always normally distributed (Fig. 12). However, mean distances between facies and channels are statistically distinct, and a few general patterns emerge. First, low elevation facies like high ponds and mangrove ponds form adjacent to first-order and maturing second-order streams, respectively. These environments adjacent to low-order

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Fig. 9. (A) The supratidal inland algal marsh (IAM) is found east of the channellized intertidal zone and is composed of luxuriant Scytonema mats (A) and (C) sometimes buried in mud during storm surges and spring tides (B). The laminated inland algal marsh sediments record a history of past storm surges, mud deposition, mud-cracking and recolonization by Scytonema mats (D). The red circle with black letter in (A) indicates the location of photograph (C). In (D), buried layers equivalent to surface units in (B) and (C) are labelled with red circles and black letters. In (A) the mangroves are 20 cm tall. In (B) and (D) the pen is 13.5 cm long. In (C) the pencil is 15 cm long.
streams receive a small but steady flux of suspended mud with all but the weakest tides and generally remain within 10 cm of MTL. These pond facies are found further away from higher order streams compared with levée crests and levée backslopes (i.e. positively sloping linear fits in the bottom three panels of Fig. 12B). Levées and levée backslopes tend to form adjacent to fourth and fifth-order streams and open ocean shorelines (Fig. 12) where the sediment delivery is zero during normal tides, but where the sediment flux becomes large during very high tides and storm surges that overtop the levée crests and beach ridges. High and low algal marshes appear to occupy a sweet spot in mean elevation and distance to high-order channels where they receive enough sediment to aggrade above MTL and avoid gastropod grazing in the ponds (Fig. 7), and not enough sediment to kill the Scytonema mats and create a levée backslope environment (Fig. 6).

A number of specific landforms also suggest that the current surface distribution of topography and facies is evolving, perhaps due to Recent changes in sea-level (Gornitz & Lebedeff, 1987; Maul & Martin, 1993; Douglas, 1997, 2001), ocean acidification (Gledhill et al., 2008) and sediment supply. For example, at the western margin of the channellized belt, beaches vary from expansive shallow regions of bioturbated mud (for example, north 27°72500-27°7330; Fig. 2) to steep winnowed slopes dominated by skeletal sand and eroded beach-rock terraces (for example, Figs 5 and 14; Shinn et al., 1969; Hardie, 1977). These observations are consistent with embayed channel mouths (Fig. 2) and truncation of channel lengths at the highest (fifth) stream order (Horton, 1945; Strahler, 1957), all of which suggest landward (eastward) movement of the erosional shoreline. A comparison of 1943 aerial photographs with 2001 IKonos satellite imagery suggests that the shoreline has retreated eastward by up to 50 m during this interval (Rankey & Morgan, 2002).

Within the channellized belt, the distribution of facies also is changing. In particular, the mangrove ponds appear to be expanding into shallower water regimes at the expense of low and high algal marsh environments. It is common to find steep (up to 5 cm vertical relief) scalloped downslope margins of low algal marshes transitioning abruptly into mangrove ponds (Figs 7 and 15). This transition often is delineated by concentrations of cerithid gastropods left just above MTL near grazed Scytonema mats of the low algal marsh (Fig. 7A and D). This mangrove pond expansion occurs in response to an increase in water depth, either due to rising sea-level, or to diminished levée-overbank mud supply to the algal marshes (Figs 11 and 12). Diminished mud supply to the marshes could be related to channel evolution or to less frequent storm surges. The TGC-wide shrinkage of algal marsh environments is consistent with comparison of 1943 aerial photographs with a 2001 IKonos satellite image, where up to ca 250 m of lateral

Fig. 10. Water height with respect to mean tide level recorded at six tide gauges (1 to 6) located in Fig. 2. The first tide gauge began recording on 21 March 2007 at 2:28 pm. Although amplitude damping and time lags are very small over 2.5 km upstream, there is a strong tidal asymmetry, with the ebb tide lasting, on average, 38% longer than the flood tide. wrt = with reference to.
algal marsh shrinkage is seen in some places (Rankey & Morgan, 2002).

Mangrove ponds also may be expanding into deeper water regimes at the expense of low and high ponds, however, this conclusion is more ambiguous. Where most second and third-order channels enter ponds, the channels bifurcate and birdfoot deltas of freshly deposited white aragonite mud (HP; Fig. 4) prograde into the ponds. These generally landward prograding (‘inside-out’; Shinn et al., 1969) deltas (for example, north of core C25; Fig. 13) represent accumulation in the intertidal zone of suspension-transported aragonite mud from the Great Bahama Bank under the influence of the strong TGC flood tides (Fig. 10). During all but the most extreme flood tides, channel waters carrying suspended aragonite mud do not flow the high levees and these pond deltas represent the terminal sink for sediment. As the mud deltas aggrade and approach mean tide level (MTL), they are colonized by mangroves and stabilized as mangrove pond environments fringing long reaches of first-order and eventually second-order streams (Fig. 4). Less common first and second-order headward-eroding dendritic channel networks drain the deep ponds during ebb tide (for example, Fig. 2, 200 m west/northwest of Fig. 13 and Fig. 4). However, the sluggish ebb tide (Fig. 10) has significantly less power than the flood tide and rarely develops sufficient shear stresses to suspend cohesive aragonite mud from the pond environments.

In contrast to the landward migrating shoreline, the inland algal marsh shows evidence for seaward (westward) expansion over pond facies (see description of Subsurface geology below). In the comparison of 1943 aerial photographs to 2001 Ikonos imagery, parts of the inland algal marsh have expanded ca 60 m seaward over that interval (Rankey & Morgan, 2002).

The integrated topographic and facies map (Fig. 4) and the facies-elevation-stream distance data (Figs 11 and 12) represent a modern snapshot of the distribution of subtidal to supratidal environments at TGC. A few of the landforms probably represent active processes responsible for the most rapidly changing facies patterns, and generally are consistent with
Fig. 12. The facies polygons in Fig. 4 were converted to a 2 m grid, and the shortest distance between each grid point and the nearest channels of each stream order were determined for the seven major facies depicted here (LC, levee crest; LBS, levee backslope; HAM, high algal marsh; LAM, low algal marsh; MP, mangrove pond; HP, high pond; LP, low pond). The number of samples for each facies is labelled as the italicized grey number in the upper left panel of (B) and is the same for all the panels in (B). (A) depicts ‘distance to nearest stream’ histograms for each facies, with colour-keyed vertical lines showing the 95% confidence interval on the mean calculated for each distribution. In some examples, distances between facies units and channels are normally distributed with statistically distinct means [for example, left hand panel of (A)]. However, even roughly normally distributed data may have very heavy tails, such as the distance between mangrove pond facies and first-order streams [upper left panels in (A) and (B)]. In other examples, distances between facies units and channels do not have a simple normal distribution, and may even appear multimodal [right panel in (A)], making the best interpretations of single population means suspect. Nevertheless, a general pattern emerges where, for first, second and third-order streams, the distance to the nearest channel decreases from high to low elevation facies (for example, from LC to HP; Fig. 11). For fourth and fifth-order streams and the shoreline to the open ocean, the distance to the nearest channel or shoreline increases from high to low elevation facies. This behaviour is consistent with the hypothesis that levées and their associated pondward facies (LBS, HAM, LAM) form adjacent to higher order streams, while low elevation facies like high pond and mangrove pond form adjacent to first-order and maturing second-order streams, respectively.
visual comparison of aerial imagery from 1943 and 2001 (Rankey & Morgan, 2002). However, in order to decipher the relative influence of sea-level change and channel migration on long-term development of the full Holocene stratigraphy at TGC, the subsurface must be probed.

Subsurface geology

The stratigraphy (Figs 13 to 20) of 187 core sites (Fig. 2) reveals the parasequence architecture of TGC. Note that the term ‘parasequence’ formally is defined as a sedimentary sequence bounded by marine flooding surfaces (Van Wagoner et al., 1990). The Holocene parasequence at TGC is not complete, and it could be argued that, with more time, a supratidal flat, aeolian deposit or palaeosol would develop before the next flooding. Nevertheless, the inland algal marsh may represent the parasequence cap and the package at TGC is sufficiently complete to merit the term ‘parasequence’.

Across TGC, the depth of Pleistocene basement decreases from >280 cm below MTL near the beach ridge to 20 cm above MTL in the eastern inland algal marsh. Over the same distance, the Holocene succession thins from up to 340 cm thick near the beach ridge (Fig. 14) to 0 to 10 cm thick in the eastern inland algal marsh (Fig. 20). At the eastern transition from the channellized belt to inland algal marsh, the Holocene package is 1 m thick (Fig. 20).

Eighty percent of cores penetrated to lithified Pleistocene basement. Ten percent of these cores retrieved 0-3 to 2 cm diameter basement fragments and, in all cases, the samples were cemented lime-grainstone. The cores that did not reach basement stopped in a coarse skeletal sand and breccia unit that was difficult to penetrate or sample with a piston corer. In all cases, this coarse sand was located in a depression or pit, 15 to 75 cm deeper than the mean surrounding bedrock depth. Figure 16 includes a basement topography map that depicts ca 50 cm of local relief on the highly irregular basement surface. Based on the grinding sound made by the probe used to measure depth to basement, most of the depressions are ca 50 to 100% filled with the coarse sand and breccia. Foraminifera are rare in these sands filling the basement relief, and a reliable age range for this discontinuous basal deposit was not determined.

Virtually all cores in the channellized belt (except those in active channels) have a basal shell-rich mud resting directly on basement, or transitioning abruptly from the pit-filling coarse sand. The shell-rich mud unit is medium grey in colour, 1 to 30 cm thick, composed of 1 to 5 cm intercalations of packstone and wackestone, and fully bioturbated. The shells in this packstone/wackestone unit are dominated by cerithid gastropods, but also may contain Peneroplis foraminifera and rare bivalves. In contrast to the shells found in modern beach and channel-bottom environments and the shells in basement karst pits, the shells in this basal packstone are not broken, and the bivalves even may be articulated. This basal packstone/wackestone unit contains abundant fibrous organic material and has a two to four times higher density of fossil mangrove roots than any other part of the Holocene stratigraphy. The mud-shell-root assemblage most resembles the modern mangrove pond environment (Figs 7 and 8), with packstones representing shallower pond-edge gastropod death assemblages (Fig. 7D), and wackestones representing shallow subtidal ponds. No evidence was found for the fossil supratidal marsh preserved at depth suggested by Shinn et al. (1969) and Shinn (1986).

Reservoir-corrected radiocarbon ages from Peneroplis proteus foraminifera tests from this basal packstone in five widely spaced stratigraphic cores (Figs 16 to 19) indicate that pond environments at TGC were first established ca 1200 years ago. These radiocarbon ages also suggest sediment accumulation rates of 1-5 to 2 mm year$^{-1}$ averaged over the full thickness of the Holocene stratigraphy in these cores.

The packstone/wackestone unit fines upward into a grey mudstone and wackestone containing rare 1 to 4 cm thick gastropod and/or foraminifer packstone layers. This mudstone and wackestone unit makes up 50 to 90% of the Holocene stratigraphy, is up to 1·5 m thick at the beach (Fig. 14), and wedges-out just east of the transition between the channellized belt and the inland algal marsh where lithified basement ramps upward towards the surface and marks the eastern boundary of Holocene intertidal deposits (Fig. 20). Like the basal packstone and modern pond environments, the shells in this wackestone are not broken and the sediment is fully bioturbated. Fossil mangrove roots are present in this unit, but are not as abundant as in the basal packstone. This mudstone and wackestone unit intersects the surface without discernible facies change in modern pond environments.

Beneath shallower intertidal environments such as levée crests, levée backslopes and algal
Fig. 13. (A) Core location map (false-colour 4,3,2 Quickbird satellite image) and (B) lithostratigraphy (generated using MatStrat; Lewis et al., 2011) near the DGPS station. Photographs depict, from left to right, faintly laminated levee crest *Schizothrix*-mud couplets, *cerithid* gastropod wackestone and channel sand with broken shells. In the legend, packstones are defined as having >20% shell content (grain-supported), wackestones are defined as having 5 to 20% shell content and mudstones are defined as having <5% shell content, roughly equivalent to the classic Dunham (1962) classification for lithified carbonates. When sedimentary units contain multiple elements (for example, >5% sand, 5% gastropods and 5% foraminifera tests; see example in legend), the lithofacies box contains wedges of different colours representing each facies element present. Channel sediments (highlighted in red) are defined by the presence of broken shells (black triangles) and sand. Black boxes in Figs 16 to 19 are reservoir-age corrected radiocarbon dates from *Peneropolis proteus* foraminifera tests. Note that channel sands and broken shells are limited to the active modern channel, and no lateral channel migration is evident. The very shallow ‘abandoned-looking’ channel crossed by cores C22 to C24 contains no record of sand or broken shells and probably was never a deep, well-developed channel. Such shallow channels sometimes represent failed bifurcations of the landward-propagating birdfoot delta. However, the highly sinuous reaches in some of these shallow channels remain mysteries. See Fig. 2 for core locations within the larger TGC area.
Shallowing-upward parasequence, Bahamas
marshes (Fig. 4), cores reveal that grey bioturbated mudstones and wackestones with mangrove roots transition within ±10 cm of MTL up into beige, faintly laminated mud with virtually no shells or roots. Below modern levée crests, the mud becomes whiter and better laminated, eventually culminating in up to 15 cm of 1 to 2 mm thick alternations of \textit{Schizothrix} bacterial mat and white mud laminations with grass roots (Fig. 6B) identical to the modern levée crest environment. The thickest unambiguous levée crest laminated mud unit found is 50 cm thick. Beneath modern high and low algal marshes, grey bioturbated wackestones also transition upward into beige muds; however, only very faint laminations are evident below the upper few centimetres that are dominated by thick fibrous \textit{Scytonema} mats.

Pioneering dye experiments conducted between 1968 and 1971 found sedimentation rates of 0.3 to 3.0 mm year\(^{-1}\) on levée crests within the channellized belt (Hardie, 1977), consistent with roughly averaged sedimentation rates over the past 1200 years based on radiocarbon ages from foraminifera in the basal packstone (see above). Assuming an average sedimentation rate of 1 to 2 mm year\(^{-1}\), and average levée crest laminated sediment thicknesses of 35 to 40 cm, levée crest environments above MTL (and thus significant channels) would have been established at least 200 to 400 years ago.

The transition from bioturbated grey wackestone to laminated white microbialite/mud across MTL also corresponds to a major geochemical change. The dominant iron mineral in the laminated white muds is magnetite, but microbial iron reduction begins to dissolve the magnetite at and below MTL (Maloof \textit{et al.}, 2007). In the lower quarter of some cores, magnetic iron sulphides precipitate in a bottom zone of sulphate reduction, probably coupled to the oxidation of decaying mangrove roots (Maloof \textit{et al.}, 2007).

Modern beach sediments are characterized by broken shells, skeletal sands (Fig. 5A and D) and rare cemented ‘beach rock’ (Fig. 5C; Shinn, 2009). No beach sediments are preserved in core more than 30 m inland of the modern shoreline (Fig. 14). Instead, a 10 to 30 cm thick veneer of beach sediments overlie levée crest muds before transitioning downward across MTL into grey wackestones. Together, these observations suggest that the beach has not prograded seaward. This result is consistent with modern erosional features such as sculpted terraces (Fig. 5A) and embayed channel mouths (Fig. 2), and truncation of stream lengths at the highest (fifth) stream order (Horton, 1945), all of which suggest landward migration of the shoreline. Interpretation of aerial imagery from 1943 and 2001 also suggest active erosion of the shoreline (Rankey & Morgan, 2002).

The inland algal marsh preserves up to 60 cm of unbioturbated white mud and light brown \textit{Scytonema} mats (Fig. 20) identical to the modern inland algal marsh environment (Fig. 9). The cores are dominated by 0.5 to 1.5 cm couplets of
white, mudcracked mud and light brown fossil Scytonema mat (Fig. 9D). The cores also reveal at least 150 m of westward expansion of the inland algal marsh over shallow subtidal grey wackestones representing ancient pond sediment (Fig. 20). Comparison of aerial imagery from 1943 and 2001 indicates up to 60 m of inland algal marsh expansion (Rankey & Morgan, 2002), suggesting that westward growth of the inland algal marsh may have begun in just the last 100 to 200 years.

In the study area (Fig. 20), the eastward reach of sub-linear first-order channels also represents the eastward extent of pre-IAM grey wackestones, suggesting that these channels are the ones that supplied mud to the ancient pond and have survived inland algal marsh progradation. Some of these streams are rare examples of channels at TGC that show evidence of headward erosion (D’Alpaos et al., 2005) as they extend landward into the inland algal marsh. Cores from tadpole-shaped pools at the heads of linear channels (C42 and C43 in Fig. 20) show a four-fold to six-fold increase in mangrove root density, suggesting that as the inland algal marsh encroaches on mangrove clusters growing along the eastern extent of ponds, the mangroves die and leave behind low regions that capture headward-extending first-order streams.

In the western 70% of the channellized belt, channel thalwegs sit directly on Pleistocene basement and scattered piles of skeletal sand, broken shells, brecciated fragments of cemented grainstone (resembling aeolianite found at localities such as Morgan’s Bluff, Andros Island) and rare flakes of cemented Schizothrix mud from levee crests. In two samples from the bottom of active modern channels, broken foraminifera give 14C dates of 1575 ± 125 (Fig. 19) and 1890 ± 30 AD (Fig. 17). These results are distinct from the much older 14C ages derived from unbroken foraminifera within the basal packstone unit preserved in cores taken outside of channels, suggesting that at least some of the shells being actively incorporated into channel sediments are Recent.

Individual transects show no evidence for significant channel incision, and even the deepest channel thalwegs usually are only 0 to 20 cm deeper than surrounding bedrock topography. In the one region where a full depth to basement survey was conducted (Fig. 16), the channel broadly divides a northern basement high from a southern basement low. The hypothesis that the relief is related to channel erosion can be ruled out because cores reveal a maximum of 10 m of northward channel migration. The relief
Fig. 16. Core location map (false-colour 4,3,2 Quickbird satellite image) and lithostratigraphy (generated using MatStrat; Lewis et al., 2011). See Fig. 2 for core locations in the larger TGC area, and Fig. 13B for the facies legend. The modern pond muds south of the main channel contain a higher fine sand fraction than any other ponds observed at TGC, perhaps sourced by storm surges that flood the nearby steep and eroded beach ridge to the west. This beach ridge has large spillover sand lobes visible as white regions in the otherwise red-brown levee backslope and algal marsh. The elevated sand content also is reflected in cores C116 to 119, which contain up to 10% sand, but do not contain any broken shells typical of channel sediments. The abundance of broken shells in the active channel course and immediately to its south suggests that the eastern channel bend has migrated northward by 5 to 10 m. The depth to basement map (upper left) shows that the channel follows a 0·5 m step from a basement low in the south to a basement high in the north.
may represent either long-wavelength karstic topography or the original shape of a relict Pleistocene dune. Antecedent basement topography probably did not play a significant role in controlling the course of the channel.

Channel bars are composed almost entirely of bioturbated mud, but frequently contain lenses of sand, broken shells and breccia clasts similar to materials found in the channel thalwegs. In the modern environment, these sands and broken shells are unique to active channels and beaches. Therefore, the lenses of sand and broken shells found at depth in cores are hypothesized to represent deposits from ancient channels that have since migrated, avulsed, or just filled in. The lateral extent of ancient channel sands is measured to calculate channel migration rates.

Figure 17 shows the clearest example of channel migration. A 40 cm thick mud and sand body with broken shells found continuously over 40 m south-east of the channel (C140) indicates that the channel has migrated at least 40 m to the north-west. In this instance, broken foraminifera at the top of the ancient channel sand in core C140 date from 1360 ± 70 AD, suggesting a maximum average rate of 6 cm year\(^{-1}\) of channel migration. Core C139 also contains a channel sand body 100 m south of the main active channel. However, depth to basement probing suggests that this sand body is not continuous with C140, and is instead associated with the nearly abandoned channel near C139.

Most channels actually show substantially slower and less significant migration consistent with earlier studies (Shinn, 1986). The channel networks in Figs 13 and 18 show zero migration, the channels in Figs 16 and 19 show 5 to 10 m of migration, and the channel in Fig. 15 shows evidence for the abandonment of a substantial channel once located between cores C26 and C30, but no channel migration. Therefore, seven out of eight channels surveyed show 0 to 10 m of migration over ca 1200 years of sediment accumulation.

**DISCUSSION**

**What does the base of the Holocene parasequence look like at Triple Goose Creek?**

Results from coring have suggested that supratidal marsh sediments extend seaward under the intertidal channelized belt, the beach ridge, and at least 1 km offshore beneath subtidal sediments (Shinn, 1986). Offshore, the supratidal marsh
Sediments are thought to be 2 to 4 m below sea-level and thus document the effect of rising sea-level (Shinn, 1986). The entire tidal flat complex may have existed farther seaward from its present location when sea-level was a metre or more below present level (Shinn, 1986).

In 90% of cores extracted from the channellized belt (Fig. 2), there is a basal 1 to 30 cm thick unit of wackestone and packstone containing unbroken shells, and abundant decaying mangrove roots and fibrous organic material. This facies assemblage resembles the modern mangrove pond, and perhaps some of the fibrous organic material represents *Scytonema* mat from the low algal marsh. However, this basal unit does not contain the centimetre-thick mudcracked mud layers and early cemented crusts diagnostic of cores taken from the supratidal inland algal marsh. In fact, cores taken from the modern supratidal inland algal marsh are relatively lean in organic matter, rarely contain gastropod shells or mangrove roots, and do not resemble this basal wackestone. Furthermore, the only crusts encountered in 187 cores were from below the modern inland algal marsh itself (Fig. 20), 5 to 30 cm below the beach ridge (Fig. 14), or as very rare, discontinuous lenses within the upper 80 cm from four cores intersecting levee backslopes. Therefore, although there is strong evidence that the peritidal lagoon extended seaward of the modern beach (see below), there is no evidence that a supratidal marsh extended seaward of its current extent as the base of the Holocene parasequence. Instead, the basal mangrove pond and low algal marsh wackestone unit lies directly on cemented Pleistocene basement or skeletal sand filled basement depressions and represents the initial development of a mud-pond system at TGC.

**What is the lag depth/time during flooding of the bank?**

Assuming steady thermal subsidence of ca 0.1 mm year\(^{-1}\) for the Great Bahama Bank, the TGC region should have been flooded by rising Holocene sea-level \(4500 \pm 2000\) years ago (Flemming et al., 1998; Toscano & Macintyre, 2003; Hubbard et al., 2005; Milne et al., 2005). However, the oldest foraminifera found in any sediments within the Holocene TGC parasequence are ca 1200 years old (Figs 16 to 19). Thus, this result implies that the first ca 3300 years of flooding must have been spent under a regime energetic enough to break shells and winnow mud away, but not erosive enough to completely peneplane the cemented grainstone basement. This hypothesis is supported by the depth to basement survey (Fig. 16) combined with coring that revealed irregular (karstic?) topography of up to ca 1 m amplitude, with the only deposition represented by skeletal sand, broken shells and brecciated fragments of cemented grainstone filling topographic depressions.

Numerous models have attempted to reproduce this lag between relative sea-level rise and the beginning of sediment accumulation (Grotzinger, 1986a; Enos, 1991; Wilkinson et al., 1991; Flemings & Grotzinger, 1996; Tipper, 1997). At TGC, lag depth is ca 1 to 3 m and lag time is ca
3300 years. Carbonate production by codiacean algae is most effective in warm shallow water, and the most common conceptual model for lag depth/time states that carbonate production scales with the area of shallowly submerged (i.e. \( \leq 6 \text{ m} \)) continental shelf (Ginsburg, 1971). In this model, initial flooding of a bank will not generate significant carbonate production until threshold subtidal bank depths and areas are achieved. Once sedimentation starts, carbonate accumulation will catch up and then keep up with sea-level rise, and will maintain a shallow environment amenable to continued carbonate production. This model also provides an internal negative feedback to carbonate accumulation, where, if the intertidal zone progrades seaward, subtidal shelf area will be reduced until carbonate production and sediment delivery to the lagoon can no longer keep up with shoreline erosion and shrinkage of the lagoon (Ginsburg, 1971).

However, subtidal shelf area cannot be the only relevant variable because some large carbonate shelves are drowned by rising sea-level and directly overlain with pelagic sediments (Schlager, 1981). Based on a seismic survey of partially drowned Cay Sal Bank (Bahamas), Hine & Steinmetz (1984) suggested that carbonate accumulation also depends on the time spent shallowly submerged during sea-level rise. For example, the top-Pleistocene surface on Cay Sal Bank lies at an elevation 15 to 20 m lower than the TGC and intersected rising Holocene sea-levels ca 8000 years ago when sea-level was rising ca 8 m ky\(^{-1}\) (compared with TGC that was flooded ca 4000 years ago when sea-level was rising ca 0.5 m ky\(^{-1}\)) (Flemming et al., 1998; Toscano & Macintyre, 2003; Hubbard et al., 2005; Milne et al., 2005). One reason that a relatively slow rate of change of sea-level may be important for initiating carbonate accumulation is that a stable seaward barrier, like a reef or a sand ridge, is crucial for developing restricted lagoonal environments that become protected terminal depocentres for carbonate mud (Kendall & Schlager, 1981; Pratt & James, 1992). So, in the case of the Pleistocene-Holocene history of the Bahamas, the rate of subsidence during the long sea-level lowstand of the last ice age determined when and how rapidly each bank-top was flooded, and thus whether the bank drowned or accumulated a thick upward-shallowing peritidal carbonate parasequence. It is worth noting that, while this scenario may help explain mud-dominated tidal flat development, it does not necessarily account...
for the landward-thinning, 1 to 4 m thick packages of coarse shallow-subtidal Holocene sediments accumulating in the lee of semi-continuous reef systems and Pleistocene bedrock islands in high energy environments, such as the wind and current swept margin of the Caicos Platform (Dravis & Wanless, 2008; Wanless & Dravis, 2008; Rankey et al., 2009).

By ca 800 AD (ca 3300 years after the initial flooding of TGC), the rate of sea-level rise had decreased enough to allow for the development of a stationary wave break and the growth of a beach ridge or barrier island west of the modern beach ridge at TGC. The ancient beach ridge would have sheltered TGC and reduced energy in its lee, allowing for the deposition and preservation of aragonite mud carried in suspension by tidal channels into a large peritidal lagoonal area (i.e. mangrove and high ponds) occupying what is now the channellized belt. Such physiography exists today in large areas of central west Andros Island (south-west portion of Fig. 1B). Furthermore, now-isolated remnants of levée crests and other intertidal carbonate muds on Williams Island (Fig. 1B; 45 km south-west of TGC) provide evidence that tidal flats and the ancient shoreline once extended seaward of the modern beach (Shinn et al., 1969; Gebelein, 1974; Shinn, 1986).

**When did the lagoon become channellized?**

In all cores from the modern channellized belt, 85 to 100% of the sediment column below mean tide level (MTL) is composed of bioturbated mudstone and *cerithid* gastropod wackestone with variable mangrove root density (for example, Fig. 13B). This facies association is indicative of modern mangrove pond and high pond environments. One of the key observations linking this pervasive mudstone and wackestone unit to an ancient pond environment is that even the frequent 1 to 8 cm thick packstone layers within the wackestone are composed of unbroken *cerithid* gastropod shells with lesser numbers of *Peneroplis* foraminifera and articulate bivalves. Today, intact shells only are found in ponds, while modern beach and channel bottom environments, as well as the coarse core-bottom lag filling karst pits, are dominated by broken shells and fine skeletal sand (for example, Fig. 13).
For the last 1200 years, this mudstone and wackestone unit with rare packstone horizons has accumulated, and continues to accumulate in modern mangrove, high and low pond environments. Packstones probably represent gastropod death assemblages on pond margins near MTL and are more abundant in the lower 10 to 30 cm of mud, suggesting an initial stage of pond-deepening before nearly 100% mudstone/wackestone deposition began. During the same interval, some regions of TGC became channelized, and by 200 to 400 years ago, channels had built levées above MTL.

Why did the pond become channelized? Today, where most second or third-order channels intersect ponds, the environment is depositional rather than erosional. In fact, there is no evidence for tidal channel extension by headward erosion into regions of greatest topographic slope (Knighton et al., 1991; D’Alpaos et al., 2005). Because flood tides have approximately twice the velocity of ebb tides (Fig. 10), floods carry greater suspended load and there is net deposition in the intertidal zone. As most tides are not large enough to overtop the main levées of higher order streams, the suspended load of aragonite mud is carried in the channels to the ponds, where the mud is dumped. In this way, an ‘inside-out’ delta progrades landward into the pond (Shinn et al., 1969; Pratt & James, 1992).

As cohesive aragonite mud accumulates, the mouth bars obstruct the flood tide and the trunk channel bifurcates into new first-order reaches, generating classic birdfoot delta morphologies. Laboratory experiments have suggested that high cohesion in the sediment and/or stabilization by vegetation are crucial factors in the formation of such birdfoot delta morphologies (Hoyal & Sheets, 2008). This process occurs over and over again, with first-order channels reaching eastward as the delta progrades into the pond. Eventually, the levées of older (seaward) bifurcations build up to MTL and become stabilized by mangroves (as is suggested by observations in Fig. 13). Continued mud accumulation constructs levées above MTL, and as the levées dry out and are colonized by cyanobacteria, the mangroves die and algal marshes and levée crests are created. Inevitably, some bifurcations are abandoned and partly filled with mud while others widen, develop dominant intertidal levée systems and occasionally begin to migrate slowly. Therefore, the higher order channels evolve to step-like forms in plan-view, where each step represents the path of the dominant channel from a former bifurcation (Fig. 21). In essence, TGC is a series of large, overlapping birdfoot deltas expanding eastward as mud is transported from the Great Bahama Bank onto the north-western edge of Andros Island.

Do sinuous tidal channels migrate?

One can see how migrating and avulsing channels on a steadily subsiding platform could lead to lateral sediment accretion and the generation of a: “planform arrangement of multiple different lithological elements, lacking simple trends in spatial arrangement, but showing some statistical relationship between element size and frequency of occurrence” (mosaic; Burgess, 2008). Such a mosaic could preserve vertical facies changes not necessarily associated with sea-level. However, the results presented here show that on the 1200 year time scale of Holocene sediment accumulation at TGC, there has not been substantial lateral migration of the eight studied channel segments. Therefore, vertical facies variability in the Holocene stratigraphy must reflect a competition between sea-level change, channel levée aggradation and landward delta expansion.

Why do the sinuous channels not migrate? Firstly, the tidal inlets breaching the beach ridge at TGC may have been virtually stationary for the last thousand years (Fig. 21). In contrast, in sandier systems with stronger longshore currents, tidal inlets may shift upwards of 60 m year⁻¹ (Kumar & Sanders, 1974). At TGC, the fixed tidal inlets may play some role in keeping fourth and fifth-order channels stationary. Secondly, the greater cohesion in rod-shaped pelletal carbonate mud compared with quartz sand or even carbonate sand probably makes TGC channel banks more resistant to undercutting, failure and lateral migration (although some amount of undercutting and failure is observable even on a day to day basis). One question that is difficult to address directly relates to how important mangroves are for stabilizing TGC channel banks. While mangroves help the initial stabilization of shallow subtidal birdfoot delta mud levées (Fig. 6A) and the inner banks of larger channels (Fig. 6A), the outer banks of channels that are most susceptible to undercutting and erosion are stabilized by cyanobacterial mats (Fig. 6A). Likewise, at least some intertidal siliciclastic systems that lack significant channel migration also have levées that are stabilized by cyanobacterial mats (Luternauer et al., 1995). Nevertheless, mangroves did
not evolve until the late Cretaceous Period (Chapman, 1976; Duke, 1995; Ellison et al., 1999), and it is difficult to say whether pre-Cretaceous tidal channels would have been as stable as those at TGC.

Despite the obvious differences between sub-aerial and submarine channel systems (Klaucke & Hesse, 1996; Straub et al., 2008; Parsons et al., 2010), perhaps there also is something to be learned about bank stabilization from the sinuous channels with prominent clay-dominated deltas common in deep-sea fans. Inclined sand bodies dipping towards the channel axis, perpendicular to the palaeoflow direction, have been identified in integrated 3D seismic and well data from deep-sea channels and interpreted as the deposits of laterally migrating point bars (Abreu et al., 2003). However, channel sinuosity seems to form during this early stage of point-bar accretion and then stabilize, because relatively few meander cutoffs (and oxbows) have been seen in these submarine channel systems (Peakall et al., 2000). Once bends form, perhaps a stable planform geometry is reached, where levees and floodplains aggrade vertically and channel bends do not migrate, as is the case for tidal channels at TGC. The increased cohesion of a clay-dominated system, where the shear stress required to erode clay from channel banks is much higher than the shear stress needed to permit deposition (Mehta et al., 1989), may be important for stabilizing submarine channel bends (Peakall et al., 2000), similar to the way that clay appears to stabilize channel bends in experimental river channels (Smith, 1998), siliciclastic tidal channels (Fenies & Fauquéres, 1998) and perhaps the way that carbonate-mud stabilizes the tidal channels of TGC.

**How is Triple Goose Creek changing today?**

Two secondary processes are actively working to shrink the intertidal channellized belt. Firstly, with each storm surge and the highest tides, muds are transported up onto the eastern supratidal zone, promoting westward progradation of the inland algal marsh. As the inland algal marsh creeps seaward over pond sediments, the supratidal zone captures the eastern reaches of first-order channels and halts the landward extension of birdfoot deltas. Secondly, perhaps in response to renewed Recent sea-level rise (Gornitz & Lebedeff, 1987; Maul & Martin, 1993; Douglas, 1997, 2001): (i) the beach ridge is migrating landward onto the truncated western margin of the channellized belt; and (ii) mangrove ponds are expanding onto the back sides of levées and flooding low algal marshes.

**Sequence stratigraphy of the Holocene parasequence at Triple Goose Creek**

The evolution of TGC can be summarized by placing it within a sequence stratigraphic framework that can be compared directly with the geological record of ancient peritidal parasequences (Fig. 22). The bottom parasequence boundary is a brecciated (karstic?) unconformity. Relief on the basement surface is filled partly with a skeletal sand and intraclast breccia lag. In fact, this coarse lag deposit represents nearly the entire transgressive systems tract, with ca 3300 years of sea-level rise recorded by just a thin veneer of broken shells. Maximum flooding is achieved near the base of the _cerithid_ gastropod wackestone that abruptly overlies the coarse lag, and sometimes

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**Fig. 21.** Sequential cartoons showing the three-dimensional development of Triple Goose Creek stratigraphy over the last 1200 years. During the last ice age, the Pleistocene aeolianite basement of TGC (grey) is karsted. When rising Holocene seas flood the margin (sea water is removed for clarity, but mean tide level would reach the base of the ksig.) the beach ridge forms, protecting TGC from wave swash. Aragonite mud produced by codiacean algae on the open subtidal shelf region to the west is delivered by tides and storms and begins to accumulate as intercalated packstone and wackestone in a mangrove pond environment. Where tidal channels traverse the beach ridge, they empty suspended mud into the pond and produce a birdfoot delta. (A) Under the influence of a strongly flood-dominated tide, net landward accumulation of sediment leads to eastward progradation of the birdfoot delta complexes. Older western portions of the channel complex widen and aggrade until their levées are supratidal. Many birdfoot bifurcations are abandoned and partly filled with mud, leaving a step-like network of higher order channel bends. (C) Perhaps by 1600 to 1800 ad, tidal channels had propagated nearly to their current eastward extent, delivering mud to the westward prograding inland algal marsh. (D) In the last 80 years, accelerated sea-level rise and/or ocean acidification have caused the shoreline to migrate landward and the mangrove pond environment to flood into algal marsh regions occupying the backsides of levées. Apparently out of phase with these transgressive trends, the inland algal marsh continues to rapidly prograde seaward. Some tidal channels continue to extend eastward by headward erosion into the inland algal marsh and are frequently captured by circular ponds once occupied by mangrove clumps before supratidal encroachment.
sits directly on lithified basement. The maximum flooding zone is difficult to place as a discrete surface, but may be just above the lowermost 10 to 30 cm of core that contains frequent packstone layers representing gastropod death assemblages from relatively shallow pond environments. The ensuing highstand systems tract is where the parasequence starts to look more like a mosaic of laterally differentiated facies and less like a layer cake. Across most of TGC, the mudstone/wackestone aggrades vertically with no perceptible ‘upward-shallowing’. However, channelization due to landward birdfoot delta progradation into ponds occurs simultaneously during the highstand systems tract, accelerating mud deposition on levées until levée crest elevations reach the height of average storm surges up to 0·5 m above mean tide level. Depending on proximity to the nearest channel and the size of the channel, the mudstone/wackestone shallows upward into either intertidal algal marsh or supratidal levée crest microbialaminite. The channels do not migrate laterally on the thousand-year time scale for parasequence development. However, as long as sediment supply is maintained, the inland algal marsh will prograde seaward over the channellized belt, capping all the intertidal facies with a desiccated supratidal parasequence top.

Had there not been a Recent acceleration in sea-level rise, Holocene sea-level rise would have continued to decelerate and the TGC intertidal zone probably would have expanded seaward. Ginsburg (1971) suggested that such seaward progradation of the intertidal zone would occur only until so much of the subtidal area had been filled that carbonate production on the shrinking open shelf could no longer fill in the accommodation space created by rising relative sea-level, resulting in shoreline erosion and retreat (Burgess et al., 2001; Burgess & Wright, 2003; Burgess, 2006). In fact, it is difficult to know whether active beach erosion and landward migration of the shoreline today is caused by rising sea-level or decreasing carbonate production related to ocean acidification, or both.

**Fig. 22.** Cartoon stratigraphic columns depicting the Holocene succession that has been (vertical yellow bar) or will be recorded in different locations (channel, levée, pond and inland) at Triple Goose Creek. At the base of the column is a cemented aeolian grainstone, probably remnant from subaerial exposure during the last ice age sea-level lowstand. Karstic relief on the top of this unit is filled with skeletal sand and intraclast breccia, marking the flooding surface developed during post-glacial Holocene sea-level rise. There is a ca 3300 year lag between flooding (dark blue line) and the first preservation of mudstones and wackestones. Subtidal (meaning always below mean tide and in most cases below low tide) mudstones with lenses of wackestones and packstones comprise the thickest unit in the succession. Near channels, intertidal levée systems comprised of *Schizothrix* microbialaminites are the first obvious indicator of an upward-shallowing trend. At inland locations, supratidal *Scytonema* microbialaminite with interlayered storm muds mudcracked and lithified to hard crusts directly overlap subtidal mudstones, with no well-preserved intertidal sediments. Barring a sudden increase in the rate of sea-level rise, this supratidal microbialaminite unit will continue to expand seaward, capping subtidal and intertidal sediments from levées, ponds and channels. Either the supratidal microbialaminite or a grainstone aeolianite deposited during a future ice age sea-level lowstand will form the top of the upward-shallowing parasequence before the next interglacial flooding surface. While the processes controlling sediment delivery, stacking and preservation differ in detail, the stratigraphy of the ‘levée’ parasequence is similar to classic models of upward-shallowing parasequences in humid, low energy, peritidal environments (James, 1984; Pratt & James, 1992).
The combined topography-facies survey result that small changes in water depth control large changes in facies [for example, the transitions from wackestone (MP and HP) to tufting (HAM and LAM) and flat-laminated (LC and LBS microbialites; Fig. 11] emphasizes the fact that seemingly large and/or abrupt facies transitions in ancient carbonate parasequences may require only very small changes in local water depth. However, it is worth noting that the range of facies observed in this modern system, while impressive and partly dependent on water depth, may be obscured due to diagenetic processes upon entering the rock record. Facies defined by biotic components with poor preservation potential or subject to bioturbation may be especially susceptible to erosion. Only those sediments that become well-laminated as a result of primary depositional processes (for example, laminated microbialites) would be distinctive in the rock record. Most other facies might be recorded as just weakly stratified bioturbated mudstone/wackestone.

CONCLUSIONS

Ancient peritidal carbonates host valuable records of environmental, sea water chemical and palaeomagnetic change. In order to interpret these signals, one must understand how time is represented in the ubiquitous metre-scale upward-shallowing carbonate parasequences of the geological record. In order to evaluate the relative importance of external (for example, sea-level) and internal (for example, migrating channels) forcing on facies stacking in shallow water carbonates, the geomorphology and Holocene stratigraphy of Triple Goose Creek (Andros Island, Bahamas) was examined. At Triple Goose Creek (TGC), ca 2 m of peritidal carbonate have accumulated over the past 1200 years under known sea-level forcing and unknown influence from currently active sinuous tidal channels.

The Modern mosaic of facies at TGC is not random. Instead, the distribution of facies is a function of water depth and sediment supply, both of which are controlled by mean elevation and the distance to the nearest channel and the size/order of that channel. Contrary to prevalent models for tidal channel initiation and early development by headward erosion in siliciclastic systems (D’Alpaos et al., 2005), intertidal channelization at TGC is dominantly a depositional process related to landward propagation of bird-foot deltas across ponds. New subtidal channel levées prograding into ponds are stabilized by mangroves until they aggrade to supratidal elevations and mangroves are replaced by cyanobacterial mats. Sequentially, first channels and then supratidal levées propagate landward across the intertidal zone like a wave, depositing sediment in a seaward thickening wedge. Once formed, channels do not migrate, or migrate very slowly. As channel banks aggrade, some reaches are abandoned, converting networks of birdfoot bifurcations into a step-like pattern of channel bends. The most mature fourth and fifth-order channels widen and migrate at the sluggish rate of 0 to 6 cm year^{-1}.

Therefore, models for upward-shallowing peritidal parasequence development that rely on lateral sediment accretion by migrating channels to explain vertical facies stacking should be viewed with caution. On the other hand, it is crucially important to be aware that visually significant changes in facies (for example, a wackestone shallowing upward into a microbialite) may represent just centimetres of water depth change. This centimetre-scale facies sensitivity to elevation thresholds probably is a fact recorded in the ancient stratigraphic record by sharp contacts between facies.

What other lessons are there to learn for interpreting ancient metre-scale upward-shallowing parasequences? Perhaps most importantly, such parasequences may record just a small fraction of the time represented by a stratigraphic column, and parasequence architecture may record little information about total sea-level change (Fischer, 1964; Grotzinger, 1986a; Enos, 1991; Wilkinson et al., 1991; Flemings & Grotzinger, 1996; Tipper, 1997). For example, at TGC, the only sedimentary record of the last glacial sea-level lowstand was a veneer of cemented and karsted grainstone. Then, even as sea-level rose to inundate the TGC region, most of the transgression is recorded only by discontinuous piles of broken shells that collected in bedrock depressions. Only the last 1.2 ky of a 100 ky ice age cycle are recorded by the upward-shallowing ca 2 m thick TGC parasequence.

Admittedly, the large amplitudes of Pleistocene sea-level oscillations are virtually unique to ‘icehouse’ periods in Earth history, and upward-shallowing parasequences developed during ‘greenhouse’ worlds may not have left as much time unrecorded. However, the highly condensed transgressive systems tract and relatively thick upward-shallowing highstand are ubiquitous.
characteristics in ancient peritidal carbonate parasequences. It has long been postulated that the fundamental processes controlling sediment accumulation in these carbonate environments, regardless of the amplitude of eustatic change, inevitably lead to a substantial lag depth/time between platform flooding and sediment accumulation, and then punctuated upward-shallowing depositional events (Ginsburg, 1971; Goodwin & Anderson, 1985; Grotzinger, 1986a; Enos, 1991; Wilkinson et al., 1991; Flemings & Grotzinger, 1996; Tipper, 1997).

Triple Goose Creek was flooded ca 4500 years ago (Flemming et al., 1998; Toscano & Macintyre, 2003; Hubbard et al., 2005; Milne et al., 2005), but actual accumulation of TGC sediment occurred exclusively during the last 1200 years when total eustatic change was <1 m. Two hundred kilometres to the south-west, Cay Sal Bank began 15 to 20 m lower than TGC and was flooded ca 8000 years ago when sea-level was rising ca 8 m ky\(^{-1}\), but has been left barren without any significant accumulation of carbonate mud (Hine & Steinmetz, 1984). Once carbonate accumulation is initiated, it is thought that carbonate production can keep up with any rate of sea-level change (Schlager, 1981). However, a relatively slow rate of sea-level rise through the critical depth range for carbonate production may be important for building a stable seaward barrier, like a reef or a sand ridge, that can shelter lagoonal environments amenable to carbonate mud preservation. For the Pleistocene–Holocene history of the Bahamas, the rate of subsidence during the long ice-age lowstand determined when and how rapidly each bank-top was flooded, and thus whether the bank drowned or accumulated a thick upward-shallowing carbonate parasequence.

It has long been realized that teasing a reliable physical record of external forcing (for example, eustasy) out of metre-scale upward-shallowing carbonate parasequences is complicated (Grotzinger, 1986a,b; Read et al., 1986; Koerschner & Read, 1989; Goldhammer et al., 1990; Wilkinson et al., 1991; Drummond & Wilkinson, 1993; Goldhammer et al., 1993; Flemings & Grotzinger, 1996; Read, 1998). However, sediment accumulation at TGC is certainly at least partly a function of eustasy, and landward migration of the shoreline and lateral expansion of the pond into algal marsh environments may be direct responses to the recent acceleration of sea-level rise and/or ocean acidification. Crucial to deciphering the sea-level control on these sedimentary responses is not just the study of vertical stacking of facies in a stratigraphic column, but also metre and kilometre-scale lateral facies variability within a parasequence.

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