Holocene sea-level fluctuations inferred from the evolution of depositional environments of the southern Langebaan Lagoon salt marsh, South Africa

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Received 27 September 1999; revised manuscript accepted 13 December 2000

A HOLOCENE RESEARCH PAPER

Abstract: A Holocene sea-level curve is constructed from the facies distribution and radiocarbon ages of sediment recovered from the distal, southern salt marsh of Langebaan Lagoon, on the southwest coast of South Africa. Calibrated radiocarbon analyses of an oyster-rich bioclastic gravel indicate that the Flandrian Transgression flooded the lagoon to 0–3 m above present-day levels by 6750 years ago (6.8 ka). Organic matter and shell material dated in distal lagoonal sediments indicate that sea level returned to present-day levels by 4.9 ka and have since remained within ±1 m of present-day levels. Bleached shell and a hiatus in sedimentation suggest an approximate 1 m sea-level lowstand between 2.5 and 1.8 ka. Changes in the macrobenthos assemblage since 1.5 ka that include the loss of the oyster *Ostrea atherstonei*, razor clam *Solen capensis*, brown mussel *Arcuatula capensis* and periwinkle *Oxystele variegata* reflect loss of hard substrate, decreased tidal-flow velocities as reworked sands prograded into the southern lagoon and possibly cooler sea-surface temperatures. Calibrated radiocarbon ages of bulk organic matter from diatom-rich, *Zostera* muddy quartzose sands indicate a 0.5 m sea-level highstand at 1.3 ka followed by a 0.5 m lowstand at 0.7 ka. Dating of fining-upward, organic-rich (2 wt % TOC) noncalcareous muds indicates that the present-day salt marsh has grown by aggradation (~1 mm y^{-1}) and progradation since 0.7 ka.

Key words: Sea level, macrobenthos, salt marsh, South Africa, Holocene.

Introduction

Melting of continental ice sheets resulted in a rapid 120 m rise in sea level, the Flandrian Transgression, between 19 and 7 ka (Bard *et al.*, 1996). Isostatic adjustments, ice-volume changes and local tectonic movements have produced highly variable regional sea-level curves for the far more subtle (<3 m) fluctuations in sea level since 7 ka (Pirazzoli, 1996). For much of the Southern Hemisphere, geophysical isostatic deformation models predict a gradual emergence of continental shorelines to present-day positions since maximum flooding by the mid-Holocene sea-level highstand of 2–4 m (Clark *et al.*, 1978; Clark and Lingle, 1979; Peltier, 1998). An estimated increase in Antarctic ice volume suggests a lowering of sea level by 1 m between 4 and 2.5 ka (Goodwin, 1998). Evidence of sea-level change from coastal areas of southern Africa generally support a mid-Holocene (6–4 ka) highstand and a lowstand from 3 to 2 ka (Miller *et al.*, 1993; Illenberger, 1988). Previous sea-level curves from South Africa were limited by poorly constrained elevated beach deposits and indirect archaeological evidence, and the need was recognized for dating of more precise sea-level indicators from lagoonal and estuarine deposits (Miller *et al.*, 1995). Studies of beachrock along the east coast (Ramsay, 1995) and estuarine deposits at Verlorenvlei on the west coast (Baxter and Meadows, 1999) have increased the resolution of sea-level change in South Africa. Refinement of local sea-level curves is important for testing isostatic and eustatic models, detecting regional variations in tectonic uplift and predicting the impact of future sea-level changes on coastal environments.

Langebaan Lagoon is a narrow, protected extension of the sea with no significant freshwater input (Figure 1). The expansive intertidal sandflat and supratidal salt marsh environments of the lagoon make it a good site for recovery of precise sedimentary sea-level indicators. Changes in local upwelling intensity would have had a minor impact on sea level, but may have affected the benthic organisms of the lagoon by varying water temperature and nutrient levels (Cohen *et al.*, 1992). Previous geological studies of Langebaan Lagoon focused on the sediment dynamics of the intertidal, subtidal and tidal channel environments (Flemming, 1977a; 1977b; 1977c). The purpose of this study is to examine

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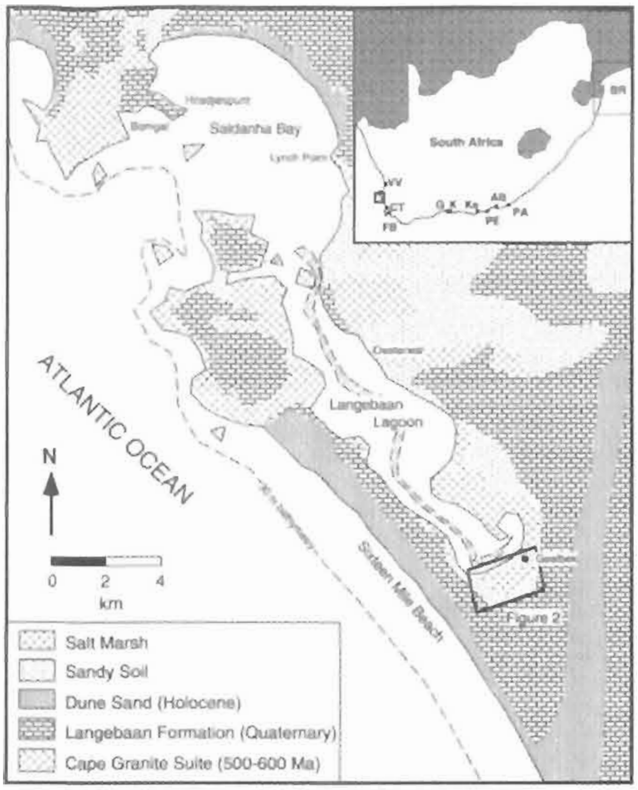

Figure 1 General geological map of the Saldanha Bay and Langebaan Lagoon area located on the southwest coast of South Africa. Present-day subtidal channel of the central Langebaan Lagoon is shown in dashed lines (Flemming, 1977a). Inset map of southern Africa shows the location of Verlorenvlei (VV), Algoa Bay (AB), east coast beachrock samples (BR) of Ramsay (1995), Knysna estuary (K), Keurbooms estuary (Ke), Groenvlei (G), False Bay (FB), Cape Town (CT), Port Elizabeth (PE) and Port Alfred (PA).

lagoonal and salt marsh sediment from the distal Langebaan Lagoon for evidence of Holocene sea-level fluctuations and associated changes in the benthic community.

Geological setting and sample sites

Langebaan Lagoon is 15 km long and up to 4 km wide and parallels the west Atlantic coast of South Africa situated 100 km north of Cape Town (Figure 1). The lagoon is separated from the Atlantic Ocean by granite headlands and Pleistocene to Recent coastal dunes and aeolianites. It is connected to the ocean via Saldanha Bay through a narrow tidal inlet in the north. Adjacent to the lagoon are Late-Precambrian to Cambrian granite hills with Pleistocene marine and aeolian sands having extensive calcrete sheets (Tankard, 1976; Rogers, 1980; Dale and McMillan, 1999). Modification of the lagoon margin by human activities has been minimal and includes pastoralism since 1600 years ago and farming by European settlers since 300 years ago (Smith *et al.*, 1991). Parts of the eastern and southern lagoon were dredged for fossil oysters and a trench was cut through the salt marsh (Figure 2). The study area has been a nature reserve since 1984.

Langebaan Lagoon is dominated by tidal currents and includes tidal channels, subtidal flats, intertidal flats and supratidal salt marsh environments (Flemming, 1977a; 1997b; 1977c). Tides are semi-diurnal and microtidal with a mean spring tidal range of 1.4 m and a maximum astronomical tidal range of 2 m (Day, 1981a). The lagoon is located in a Mediterranean, semi-arid climate with an annual rainfall of 240 mm received mostly in the winter months (May–August). Freshwater springs occur along the lagoon margin, but there is no significant river water or sediment input to the lagoon. Most of the aeolian sand is trapped by Holocene barrier dunes west of the lagoon (Figure 1).

At the distal, far southern end of the lagoon is a 3 km^2 salt marsh separated from the open lagoon by a 2 m high ridge that is cut by a single large channel through which tidal waters enter and leave the marsh (Figure 2). Salt marshes also exist along the southwestern and eastern edges of the lagoon. Cores for this study come from the southernmost salt marsh and from its main channel out to the low-water spring tide level at the edge of the open lagoon (Figure 2). A hand-held vibrational coring device consisting of polyvinyl tubing and an expandable rubber plug was used to extract 1–2 m long sediment cores (Heron, 1997).

Results

Radiocarbon ages

Conventional radiocarbon ages were determined on bulk organic matter and shell samples by the Quaternary Dating Laboratory in Pretoria and are referred to here as yrs BP (Table 1). The CAL4C programme was used to obtain calibrated ages along with their 2σ range (Talma and Vogel, 1993). Shell material was calibrated using the marine data set of Stuiver and Braziunas (1993) assuming a reservoir age of 550 years for Atlantic coastal waters and 400 years for southern and eastern coastal waters of South Africa (J. Vogel, personal communication, 1998). Bulk organic carbon ages were calibrated using the Northern Hemisphere tree-ring data, adjusted by 40 years for the Southern Hemisphere (Vogel *et al.*, 1993). Calibrated ages are referred to as ka, thousands of years before the present (1950).

In addition to analytical errors reported in Tables 1, 2 and 3, there are uncertainties in the source of the organic carbon analysed. Primary production in the lagoon is estimated to be 55% macrophytes, 23% phytoplankton and 22% benthic diatoms (Christie, 1981; Fielding *et al.*, 1988). Primary production in the salt marsh is dominated by macrophytes that are assumed to take up their carbon directly from the atmosphere with no reservoir effect. However, bulk organic carbon dates may require a reservoir correction. For example, the calibrated ages of bulk organic carbon from samples TOP83 and TOP159 would be 138 to 275 years younger if 25 to 50% of their carbon had a marine rather than an atmospheric source. Bulk organic carbon samples may also include older, reworked organic matter from bioturbation or erosion and are considered to represent maximum ages. Bulk organic carbon $\delta^{13}C$ values range from −25.8 to −16.4‰ (Table 1) and suggest a mixture of C3 and C4 plant sources. Some carbonate shell samples were etched and pitted, but there was no evidence of recrystallization or calcite overgrowths. Radiocarbon ages are best viewed as mean ages for the sediment interval sampled because up to 300 g of sediment and 30 g of shell were required for organic and inorganic radiocarbon analyses, respectively.

Stratigraphy and sedimentology of the southern salt marsh pans

The salt marsh surface is covered by a diverse and abundant macrophyte community whose distribution is largely related to elevation and salt tolerance (Day, 1981b). The rich macrophyte (*Arthocnemum*) cover of the salt marsh is flooded fortnightly by high-water spring (HWS) tide levels 0.7 m above mean sea (tide) level (MSL). Locally depressed pans and channels, elevated up to 0.5 m above the salt marsh, are flooded only at maximum astronomical tide (HAT), during storm surges or by runoff from winter rains. These pans lack macrophytes but contain algae when flooded in winter, with salinities of 25 to 37‰. In summer, surface waters in the pans reach halite precipitation (280‰). Salt pan sediments are calcareous and have porewater salinities of around 100‰, whereas salt marsh sediments are noncalcareous and have salinities between 35 and 42‰ year round (Heron, 1997).

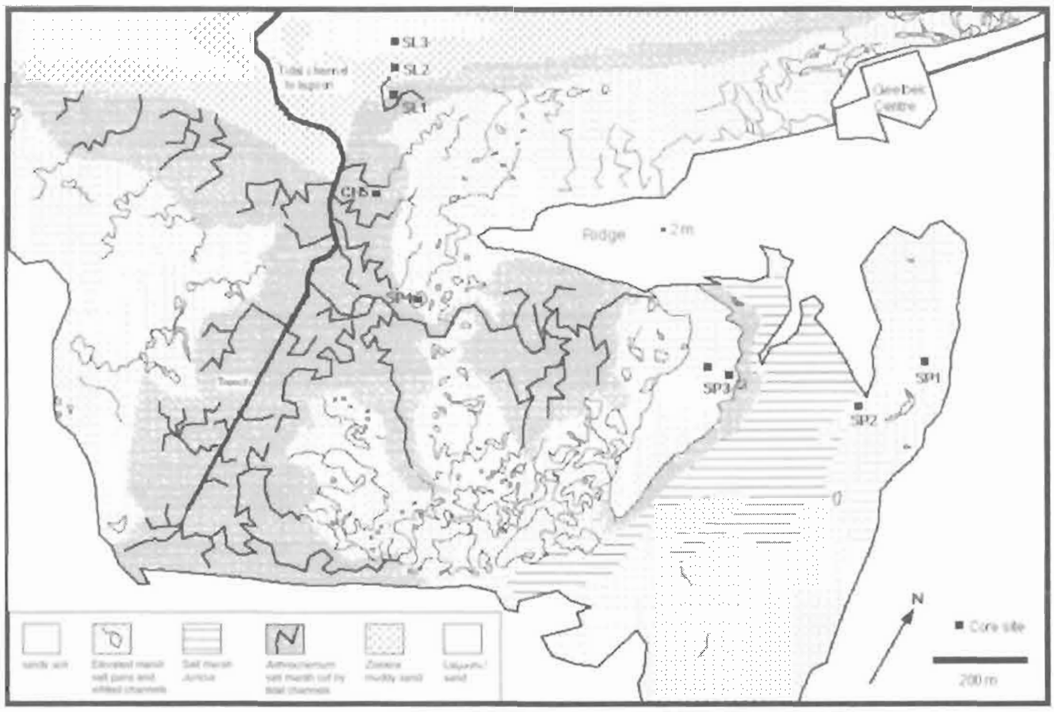

Figure 2 Southernmost salt marsh of the Langebaan Lagoon showing the distribution of depositional environments, active (heavy line) and relict (light line) tidal channels, evaporative salt pans and core locations. Two, essentially identical, cores were recovered at SP3. (Based on a 1989 colour aerial photograph from Council for Scientific and Industrial Research, Stellenbosch.)

Table 1 Radiocarbon analyses of shell and organic matter (OM) from the southern Langebaan Lagoon

Sample	Core: depth (cm)	Material	Analytical no. Pta-	$\delta^{13}C$ ‰ PDB	Conventional ^{14}C age yr BP	Calibrated ^{14}C age cal. yr BP	2σ range cal. yr BP	Indicated sea level relative to present day (m)
SP1-87	SP1:70–80	OM, bulk	7570	−23.8	4260 ± 80	4829	4529–4879	0 to −0.7
SP2-48	SP2:18–28	OM, bulk	7576	−25.0	560 ± 45	529	498–558; 612–627	−0.2
SP2-82	SP2:50–60	OM, bulk	7579	−25.8	840 ± 45	698	662–776	−0.5
SP3-99	SPE:88–92	*P. capensis*	7557	+1.9	4850 ± 70	4873	4800–5084	0 to +1
TOP83	SL1:25–35	OM, bulk	7589	−20.6	450 ± 70	488	299–539	−0.4
TOP159	SL1:68–86	OM, bulk	7597	−16.4	1390 ± 50	1280	1171–1322	0 to +0.7
SL2-105	SL1:85–105	Shell, mixed	7773	+1.0	3470 ± 60	3158	2973–3321	−0.5 to +1.0
SL3-48	SL3:30–48	Shell, mixed	7771	+1.0	2920 ± 50	2464	2333–2684	−0.5 to +1.0
BOT126	SL3:80–90	Shell, mixed	7558	+1.6	4510 ± 50	4487	4370–4606	−1 to +1
BOT176	SL3:125–135	*O. atherstonei*	7564	+1.1	6460 ± 70	6750	6617–6900	0 to +3

Cores SP1 to SP4 were recovered from salt pans on the eastern edge of the central tidal channel of the salt marsh (Figure 2). Holocene sediment in the southern salt marsh is generally less than 1 m thick, and in cores SP1, SP2 and SP3 is deposited on Upper Pleistocene calcarenites (Figure 3). The recovered sediment contains one or more fining-upward sequences (Figure 3). The basal salt marsh sediment consists of bioturbated sands (<10% mud) and muddy sands (10–50% mud) that are noncalcareous, low in organic carbon (<0.5 wt%) and commonly contain abundant fossil benthic diatoms. These sands are overlain by calcareous sandy muds (5–50% sand) that contain abundant ostracods, carbonate-cemented grains, roots, pyrite and bulk organic carbon contents of around 2 wt %. The sand fraction of the sediment is primarily reworked aeolian and aeolianite quartz grains plus calcareous bioclasts. The mud fraction consists of quartz, diatom frustules, organic matter, sponge spicules and the clay minerals illite, kaolinite and smectite. The transition from sand- to mud-rich facies can be abrupt (cores SP1 and SP2) or gradational (cores SP3 and SP4). Where the transition is abrupt, the underlying sand is yellow to orange rather than grey to tan in colour. Radiocarbon ages of shell and organic matter recovered from the salt pan cores range from 4.9 to 0.5 ka (Table 1; Figure 3).

Stratigraphy and sedimentology of the southern lagoonal salt marsh

Four cores were recovered along a transect from the mean high-water spring (HWS) level salt marsh banks of the main channel to the mean low-water spring (LWS) level intertidal sand flats at the southern end of the lagoon (Figure 2). Two cores (CH5 and SL1) are from HWS tide level, core SL2 is from low-water neap (LWN) tide level with abundant *Zostera capensis* growing at its surface, and core SL3 is from bioturbated intertidal sandflats at LWS tide level (Figure 4). The base of cores SL2 and SL3 consists of bioturbated shelly sands that are overlain by a bioclastic gravel that ranges in age from 6.8 to 3.2 ka. The 2.5 ka shells in the overlying shelly sands are increasingly bleached (weathered) and less abundant up core as the sediment fines upward into a diatom-rich, noncalcareous muddy quartzose sand with abundant sand-filled burrows and *Zostera* roots in the upper 10 cm of core SL2. Sediment in cores SL1 and CH5 ranges in age from 1.3 to

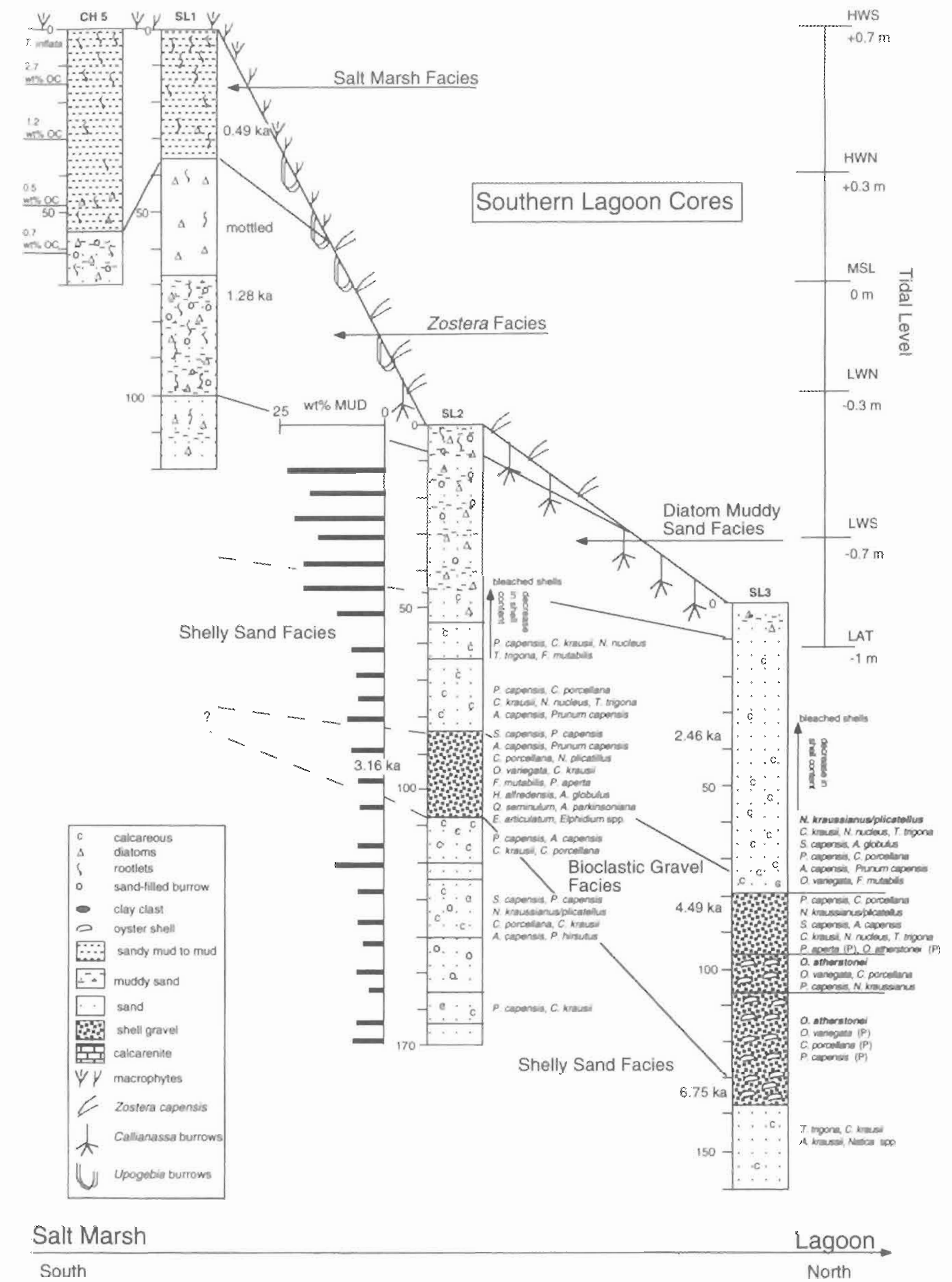

Figure 4 Stratigraphy of vibracores recovered from the salt marsh out to the intertidal sand flats of the southern Langebaan Lagoon, showing the lithofacies, tidal elevations, calibrated radiocarbon ages, organic carbon and grain-size analyses and fossil assemblages (bold indicates dominate species; P indicates fossil is present). Tidal levels are high-water spring (HWS), high-water neap (HWN), mean sea level (MSL), low-water neap (LWN), low-water spring (LWS) and lowest astronomical tide (LAT).

Table 4 Comparison of fossil (mid-Holocene) and living benthos of Langebaan Lagoon

Fossil only	Fossil and living	Living only*
Salt pan (>1 m above MSL)	*Sarscypridopsis aculeata* (A), *Gomphocythere cf capensis* (A), *Tomichia ventricosa* (LA)	
Salt marsh (*Arthrocnemum*) (0 to 1m above MSL)	Diatoms (P), *Hymeniacedon perlevis* spicules (C) *Trochammina inflata* (C), *Assiminea globulus* (A)	*Littoraria* spp. (C) *Nucella dubia* (P)
Intertidal muddy *Zostera* sands (0 to 0.7 m below MSL)	Diatoms (A), *Hymeniacedon perlevis* spicules (C) *Gibbula cicer*, *Gibbula beckeri* (P) *Nassarius kraussianus* (LC) *Haminoea alfredensis* (P)	*Cylindroleberis* spp. (C), *Upogebia africana* (C), *Tellina trilatera* (LC), *T. gilchristi* (P), *Venerupis corrugata* (C), *Siphonaria compressa* (LC)
Intertidal sandflats (0.7 to 1 m below MSL)	Diatoms (LC), *Quinqueloculina* (cf.) *seminulum* (C), *Ammonia parkinsoniana* (C), *Elphidium articulatum* (P), *Elphidium* spp. (P) *Callianassa kraussi* (A)	*Cycloleberis galatheae* (C) *Clionella sinuata* (A)
Subtidal sandbanks (>1 m below MSL) *Solen capensis* (LA) (Abundant)	*Carditella rugosa* (C), *Carditella capensis* (P) *Nucula nucleus*, *Tellimya trigona* (C) *Fissurella mutabilis* (C) *Nassarius plicatellus* .... *Philine aperta* (P), *Pilumnus hirsutus* (P) *Protomella capensis* (A), *Prunum capensis* (A)	*Macoma crawfordi* (LC) *Mactra glabrata* (LA) *Venus verrucosa* (P) (Present), *Nassarius speciosus* (C) *Natica tecta* (C)
Rocky substrate *Ostrea atherstonei* (A) (subtidal) *Arcuatula capensis* (LA) (*Zostera* facies/intertidal) *Oxystele variegata* (A) .... (intertidal)	*Anachis kraussii* (P) *Crepidula procellana* (A) *Dendrofissurella scutellum* (C) Restricted to Saldanha Bay	*Austromegabalanus algicola* *Balanus amphitrite* (C) *Choromytilus meridionalis* (P) *Burnupena* spp. (C) *Nodilittorina africana* (C) *Siphonaria serrata* (A)

A = abundant, C = common, P = present, L = locally.
*(Day, 1959; Carr, 1976; Christie and Moldan, 1977; Puttick, 1977).
Identification and habitats of molluscs from Branch and Branch (1981), Kilburn and Rippey (1982) and Branch *et al.* (1994).

N. plicatellus appear to have been largely displaced by the living molluscs *M. glabrata*, *C. sinuata* and *N. speciosus*. This change in mollusc assemblage indicates a loss of subtidal sandbank environments after 1.3 ka in the southern lagoon.

The subtidal shelly sand facies grades upcore into the intertidal diatom-rich muddy sand facies at higher elevations with a lower slope and tidal flow velocities. The absence of large filter-feeding molluscs from the muddy sands of the intertidal flats may result from unsuccessful competition with the frenetic burrowing activity of the sand prawn *Callianassa kraussi* (Branch and Branch, 1981). Abundant fossil pincers up to 30 mm long and sand-filled burrows indicate that the intertidal flats were intensively bioturbated by *C. kraussi* which is capable of turning over 60% of the sediment to a depth of 30 cm in a one-month period (Branch and Pringle, 1987). Rapid sediment turnover results in living benthic diatoms to depths of 30 cm with the greatest diatom biomass found in the intertidal sands of the southern Langebaan Lagoon (Fielding *et al.*, 1988). Fossil diatom abundance in this facies is highly variable, but can make up as much as half of the mud fraction along with sponge spicules, quartz silt, organic matter and clay minerals. Sponge spicules are derived from the sponge *Hymeniacedon perlevis*, which is locally common in salt marsh and *Zostera* vegetation of the lagoon (Day, 1959). The benthic forams *Quinqueloculina cf. seminulum* and *Ammonia parkinsoniana* are common and *Elphidium articulatum* and *Elphidium* spp. are present as fossils and probably live in the lagoon today (I. McMillan, personal communication, 1999).

There is a distinct loss of calcareous fossils in the transition from subtidal channel and sandbank environments upslope into the diatom-rich muddy sand, *Zostera* and salt marsh facies, where siliceous fossils are common to abundant, and well preserved. Preservation of calcareous fossils in subtidal environments may result from less bioturbation, less frequent subaerial exposure, and lower organic matter contents than the muddier intertidal to supratidal environments. The diatom-rich, muddy sand facies grades into the *Zostera* facies which is also diatom-bearing (including epiphyte pennate diatoms) and is heavily bioturbated by the mud prawn *Upogebia africana*. *U. africana* is not observed as a fossil possibly because of its far smaller and thinner carapace compared to *C. kraussi*. The *Zostera* facies has higher mud, root and organic matter contents than the intertidal sand flats, reflecting the impact of *Zostera capensis* plants that grow between LWS and MSL (Day, 1981b). The upcore transition from the *Zostera* facies to salt marsh facies is marked by a further increase in mud, root and organic matter contents. Diatoms and spicules are present, but are not as abundant as in the *Zostera* and sandflat sediments. The salt marsh facies is flushed daily by the tides with near seawater salinities and generally lacks calcareous fossils. The salt marsh contains relatively few macrobenthos. The agglutinated foram *Trochammina inflata* occurs in the salt marsh sediment, but the locally abundant gastropod *Assiminea globulus* is not preserved as a fossil except when transported to subtidal sands.

The abundant fossil ostracods *Sarscypridopsis aculeata* and *Gomphocythere cf capensis* (K. Martens, personal communication, 1998) recovered from the salt pan cores (Figure 3) are typical of temporary freshwater pools and can tolerate salinities perhaps as high as 11‰ (Dingle and Honigstein, 1994; Martens *et al.*, 1996). These ostracods, and the locally abundant fossil gastropod *Tomichia ventricosa*, probably thrive after the pans fill with brackish waters during periods of high rainfall or lowered sea level. In addition to biogenic carbonate, the salt pan facies contains abundant calcite-cemented grains that probably precipitate directly from evaporated pan waters. Organic carbon contents are similar (1.9 to 2.7 wt %), but the salt pan facies has a higher postdates the proposed increase in Antarctic ice volume and approximate 1 m sea-level lowering between 4 and 2.5 ka (Goodwin, 1998).

Sea level appears to have remained below present-day levels until at least 1.8 ka as determined from an organic gyttja deposit at Groenvlei (Deevey *et al.*, 1959) and a salt marsh deposit from Klaarfontein Spring at Verlorenvlei (Baxter, 1997). Sea level then rose to around +0.5 m as evidenced by the 1.3 ka *Zostera* facies recovered in core SL1, the pollen record at Verlorenvlei (Baxter, 1997) and 1.5 ka oyster shells recovered from the central Langebaan Lagoon (Flemming, 1977a). The approximate +0.5 m highstand from 1.5 to 1.3 ka was followed by a lowstand of around −0.5 m from 0.7 to 0.4 ka. *In-situ* tree stumps exposed at low tide in Knysna estuary also indicate a lowstand on the south coast at 0.7 ka (Marker, 1997). A rise in sea level to its present-day position since 0.4 ka is supported by active cliff erosion of the Upper Pleistocene aeolianites along the western margin of the lagoon and erosion of the foredune along Sixteen Mile Beach. The lowstand between 0.7 and 0.4 ka corresponds to evidence for lower sea-surface temperatures along the west coast between 0.5 and 0.4 ka (Cohen *et al.*, 1992). Steric effects from colder water temperatures may have contributed to these sea-level lowstands (Jerardino, 1995). The lowstand also overlaps with the 'Little Ice Age'(Grove, 1990) and suggests that climate change may have influenced late-Holocene sea-level fluctuations.

Summary and conclusions

Facies analysis, combined with radiocarbon ages of sediment recovered along a depth profile of the southernmost salt marsh of the Langebaan Lagoon, has increased our understanding of subtle (±1 m) sea-level changes at the end of the Flandrian Transgression on the southwest coast of South Africa. Subtidal oyster-rich bioclastic gravels indicate that high-energy tidal channels extended the length of the lagoon by 6.8 ka and that sea level was at least 0–3 m above present day. Sea level was near present-day levels between 4.9 and 2.5 ka followed by a hiatus in deposition and subaerial exposure that indicates a drop in sea level of 1–2 m between 2.5 and 1.8 ka. A brief sea-level highstand of approximately +0.5 to +1.0 m is centred around 1.3 ka, followed by a lowstand of around −0.5 to −1.0 m centred around 0.7 ka. A marine transgression since 0.7 to 0.4 ka has resulted in the accumulation and progradation of the present-day salt marsh.

Holocene sea levels from Langebaan Lagoon are generally similar to those proposed previously for the west and south coasts of South Africa. The mid-Holocene sea-level maximum on the west and south coasts of South Africa appears to occur at least 1 ky before it occurs on the east coast of South Africa. The return to present-day sea levels is also much more rapid on the west and south coasts of South Africa than predicted by global geophysical models of isostatic deformation. Sea-level fluctuations of around ±1 m since 4.9 ka may correspond to changes in climate, ocean temperature, ice volume or tectonism.

The recovered mid-Holocene molluscan faunal assemblage is broadly similar to the living assemblage of the lagoon but with several notable exceptions. The abundant fossil oyster *Ostrea atherstonei*, razor clam *Solen capensis*, brown mussel *Arcuatula capensis* and the periwinkle *Oxystele variegata* no longer live in the lagoon. The change in molluscs appears to reflect a replacement of subtidal sand and hard substrate habitats by intertidal sand flats and salt marshes, particularly after the marine regression between 2.5 and 1.8 ka. The drop in sea level allowed reworked sands, eroded during the previous highstand from surrounding Upper Pleistocene aeolianites, to prograde into the lagoon, filling channels and lowering the energy gradient of the southern lagoon and replacing hard substrate with sandy to muddy sand habitats. Changes in tidal-current flow velocities related to substrate changes and colder surface water temperatures associated with intensified upwelling since 4 ka have probably also contributed to the faunal changes in the lagoon.

Acknowledgements

I thank G. Branch, C. Griffiths, J. Pether, I. McMillan, K. Martens and J. Day for help in the identification of fossils and their habitats, and the West Coast National Park for permission to sample the salt marsh. This research was funded by the South African National Research Foundation and the University of Cape Town. Comments from J. Rogers, D. Miller and I. Goodwin improved the manuscript.

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