

**Application of intertidal salt-marsh
foraminifera to reconstruct late Holocene
sea-level change at Kariega Estuary,
South Africa**

by

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Frontispiece



Galpin's salt marsh, Kariega Estuary, South Africa

Abstract

Unclear predictions surrounding climate change, associated sea-level rise and potential impacts upon coastal environments have placed an emphasis on the importance of sea-level change. Past sea-level fluctuations have been measured using biological and geomorphological forms of evidence. One such biological proxy is salt-marsh foraminifera, which have been used as a high-resolution indicator of past sea-level change, based on the assumption that surface foraminiferal assemblages are similar in composition to buried fossil foraminifera. In South Africa, there is ongoing research seeking to produce high-resolution records of sea-level change, however foraminifera remain an underutilized source of proxy evidence. This research applies salt-marsh foraminifera as precise indicators of relative sea-level change at Kariega Estuary on the Eastern Cape coastline of South Africa. Distributions of modern foraminiferal assemblages were investigated, revealing vertical zonation across the intertidal zone. The foraminiferal and marsh vegetation zones were in part similar and overlapped to a certain extent, identifying three zones; high, low and tidal flats. This suggested foraminiferal distribution is a direct function of elevation relative to tidal fluctuation. A 94 cm core consisting of peat, sand and clay sediments was extracted from the salt marsh. A chronological framework for the core was based on five AMS radiocarbon age determinations of both bulk sediment and shell fragment samples placing the record within the last 1500 years Before Present (BP). The basal shell age was a clear outlier to all bulk sediment ages, possibly as a result of shell recrystallisation. The bulk sediment age determinations suggested two possible age reversals, potentially linked to sedimentary hiatus or contamination. These inconsistencies in the chronology were best viewed as separate age models. The core was analysed at a high resolution, whereby fossil foraminifera were extracted every 2 cm's down the core. A transfer function was applied to calculate the former elevation at which each core sample once existed, to produce a relative sea-level reconstruction. The reconstruction was related to the age models to produce two possible sea-level curve scenarios. Reconstructed curves from both scenarios depict a 0.5 m (± 0.16 m) sea-level highstand at 1500 cal years BP followed by a lowstand of -0.6 m (± 0.03 m). Scenario One reached its lowest recorded sea-level between 600 cal years BP and 500 cal years BP and then fluctuated below present day levels. Scenario Two reached its lowest recorded sea-level around 1200 cal years BP, followed by low amplitude fluctuations and a relatively stable period from 100 cal years BP till the present day. The 1500 cal years BP highstand recorded for both scenarios correlates well with existing palaeoenvironmental literature from the southern African coastline. Chronological limitations associated with the remainder of the record hinder inter-comparison with previous studies. The outcomes of this research suggest that intertidal salt-marsh foraminifera demonstrate enormous potential for the high-resolution reconstruction of relative sea-level change in the South African context.

Preface

The experimental work described in this dissertation was carried out in the School of Agricultural, Earth and Environmental Sciences, University of KwaZulu-Natal, Pietermaritzburg, from January 2010 to February 2013, under the supervision of Prof T.R. Hill and Dr J.M. Finch.

These studies represent original work by the author and have not otherwise been submitted in any form of degree or diploma to any University. Where use has been made of the work of others it is duly acknowledged in the text.

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CHAPTER ONE: INTRODUCTION

The fundamental characteristic of the earth is that it has undergone constant change, varying in size and magnitude, through time (Hidore, 1996; Oldfield, 2005). The earth has been warmer and cooler, wetter and drier, species have disappeared and evolved (Oldfield, 2005), and sea levels have fluctuated through time, inundating and draining coastal environments (Lambeck *et al.*, 2004). The drivers of relative sea-level change include tectonic (uplift and subsidence), eustatic (fluctuations in global ice volume or ocean basin volume), hydro-isostatic processes and fluctuations in sediment supply (Carr and Botha, 2012).

As a consequence of recent advances in man's understanding of global environmental fluctuations over the last two million years and species response to past environmental changes, detailed reconstructions of past environments, including relative sea-level, are now possible (Roberts, 1998). Improved understanding and knowledge of past environments can provide insight into predictions of future environmental climatic changes (Williams *et al.*, 1993).

During the 20th century, sea level has risen significantly faster than in past centuries (Bindoff *et al.*, 2007) and has been a major threat to society especially to those living in low-lying coastal areas (Church *et al.*, 2001; Nicholls and Cazenave, 2010; IPCC, 2011). There are approximately 200 million people living in coastal regions that are vulnerable to flooding by extreme sea levels and the erosion of shorelines (Nicholls and Cazenave, 2010). The displacement of coastal ecosystems such as salt marshes or intertidal zones as a result of sea-level rise could cause significant economic and ecological losses. Such ecosystems are highly productive and provide a number of important ecosystem services; flood protection, storm protection, nurseries for fish species, waste assimilation and are important to nature conservation (Nicholls and Cazenave, 2010).

Evidence of sea-level change is derived from archaeological sites, geomorphological features, tidal gauge records, satellite records, geophysical modeling and salt-marsh sediments that accumulated during the Holocene (Cronin, 2012). In the southern African context, information related to past environmental change has been limited due to the lack of researchers, funding and environmental limitations regarding the recovery of palaeoenvironmental data (Chase and Meadows, 2007). Situated away from the large ice sheets in the northern hemisphere, southern Africa has been somewhat less affected by the melting of the ice sheets during the post-glacial period (Norstrom *et al.*, 2012). Southern African sites are useful to understanding eustatic sea-level change as they are generally tectonically stable, however, other factors such as the loading of estuaries by fluvial sediments and water table fluctuations caused by human activity need to be acknowledged (Nicholls and Cazenave, 2010). Several recent reviews of sea-level change exist (Church and White, 2011; Gehrels *et al.*, 2011; Nicholls *et al.*, 2011; Siddall and Milne, 2012; Gehrels and Woodworth, 2013) although these have a tendency to focus on the northern hemisphere.

Much geomorphological evidence lies in erosive features; such as benches, notches, pools, terraces and sea caves, which can be difficult to date (Isla, 1989). Marine deposits are one of the most commonly used indicators as their shape and stratigraphy are easily identified as coastal features (Pirazzoli and Pluet, 1991). However, shell or coral fragments can add uncertainty, as they may be older than the deposits where they are situated due to tidal range and storm events. It is also extremely difficult to quantify how close the date of organic material is to specific palaeolevels (Pirazzoli and Pluet, 1991), as one is unable to attach a specific date without supporting evidence from other related sea-level indicators.

In addition to geomorphological evidence a number of biological proxies have been used to elucidate sea-level change. Coral reefs and encrusting marine organisms collected at different growth stages are one of the most reliable measures of the age of past sea levels, especially for coral reefs and coralline algae and those species that inhabit areas close to the low tide level (Pirazzoli and Montaggioni, 1988; Hopley *et al.*, 2007). Coastal plant remains can be used for sea-level studies especially when they are found layered between marine deposits, the most widely used material for such work is marshy peat. However, it is important to remember that compaction can occur and to minimize the error, the preferred peat layer to use is that which sits above the hard older stratum. This is a good indicator of when water first flooded previously open ground along the coastline (Pirazzoli and Pluet, 1991).

Salt marshes display distinct vegetation zonation, created by different types of plants being able to withstand tidal inundation and inter-specific competition. The relationship between elevation and salt marsh vegetation has allowed researchers to date plant fragments that are embedded in salt-marsh sediments, providing them with accurate ages of sea-level markers (Gehrels, 1994). The disadvantage of this method is the large and changeable vertical distribution of plants, however plant fragments have the advantage of being visible to the naked eye (Gehrels, 1994).

Submerged forests and mangroves are also good indicators of past sea level (Pirazzoli and Montaggioni, 1988). This is especially true in situations where tree species that inhabit terrestrial areas further inland are found at significant depths along the continental shelves (Douglas *et al.*, 2001). However, submerged forests are often buried below marine sediments and can only be studied by employing geophysical surveys and advanced coring methods (Pirazzoli and Montaggioni, 1988; Pirazzoli and Pluet, 1991). Another important consideration is whether the organism was buried *in situ* or was transported and deposited elsewhere. In instances where a rooted tree stump is discovered it is safe to assume that it could not have been transported (Douglas *et al.*, 2001).

Sediment records derived from salt marshes can be analysed using several proxies including foraminifera, diatoms and plant macrofossils, thereby providing regional sea-level curves with an accuracy of 5-20 cm (Baxter and Meadows, 1999; Cronin, 2012). Diatoms are excellent indicators of past and present

environmental conditions particularly in estuarine systems, as they have discrete and well defined habitat preferences, and are abundantly and well preserved in sediment. Thus, changes in diatom assemblages down a dated sediment core are valuable in determining past environmental conditions (Denys and De Wolf, 1999; Norstrom *et al.*, 2012).

Salt-marsh foraminifera have become an essential tool for reconstructing the characteristics and timing of palaeoenvironmental change (Scott *et al.*, 2001). Foraminifera are unicellular marine organisms which form shells, or tests, that remain fossilized in sediment following death. Their fossils are durable and easily collected and separated from the sediment, making them excellent tools for reconstructing historical environmental change (Scott *et al.*, 2001). Accurate sea-level information can be obtained as a result of the vertically zoned nature of salt-marsh foraminifera, related to tidal levels and altitude (vertical zonation concept; Scott and Medioli, 1978). In some cases, salt-marsh foraminiferal assemblage zones can define sea level to within ± 0.05 m (Leorri *et al.*, 2010). This is the result of the relationship between the assemblages of agglutinated salt-marsh foraminifera and the elevation above mean sea level (Scott and Medioli, 1978). Salt marshes experience both daily and seasonal variations in salinity, frequency of flooding and other parameters, which are linked to tidal overflow (Massey *et al.*, 2006). Salt-marsh foraminifera are controlled as much by tidal elevation as by the mean salinity of the water (Leorri *et al.*, 2010). The distribution of salt-marsh foraminifera is generally similar in all temperate areas and follows the vertical zonation concept, in that they serve as accurate proxies for past sea-level reconstruction (Scott and Medioli, 1978).

Although salt-marsh foraminifera are considered to be accurate and precise tool for reconstructing sea-level changes during the Holocene (Scott and Medioli, 1978; Gehrels, 1994; Woodroffe *et al.*, 2005), there has been limited application of this proxy in South Africa (Albani, 1965; Moura, 1965; Martin, 1981; McMillan, 1986; 1990; 1993; Cooper and McMillan, 1987; Wright *et al.*, 1990; Rocha, 1995; Lindsay *et al.*, 1996; Franceschini *et al.*, 2005). Relatively little has been published regarding the modern distribution of salt-marsh foraminifera in South Africa and their relationship with tidal levels remains inadequately quantified (Franceschini *et al.*, 2005). Monitoring and management of recent sea-level change can be achieved using salt-marsh foraminifera, which ultimately can inform remediation initiatives related to changing coastal land use (Scott *et al.*, 2001). With the possibility of large scale environmental and climatic changes predicted for the future, insight into the effects of past climatic changes will contribute to a better understanding of conditions that may emerge in the near future (Anderson *et al.*, 2007). This research investigates the application of intertidal salt-marsh foraminifera to reconstruct relative Holocene sea level along the Eastern Cape coastline of South Africa.

1.1 Aim and objectives

The aim of this research is to determine the applicability of intertidal salt-marsh foraminifera to reconstruct relative sea-level change for an estuarine system along the east coast of South Africa. Specific objectives are as follows:

- I. To map estuarine vegetation and compare with compositional changes in the surface foraminiferal assemblages;
- II. To determine whether foraminiferal assemblages in the Kariega Estuary show evidence of vertical zonations;
- III. To extract a series of sediment cores from the salt marsh and establish a stratigraphy;
- IV. To determine foraminiferal species composition and abundance at increments along a selected sediment core;
- V. To develop an age-depth model for this sediment core;
- VI. To develop transfer functions using ecological information contained within modern distributions of salt-marsh foraminifera and
- VII. To apply a transfer function to fossil assemblage data to reconstruct a relative sea-level curve.

1.2 Structure of thesis

Chapter one introduces foraminifera in terms of their palaeoenvironmental significance in the reconstruction of sea-level change, particularly in the South African context. Sea-level change is an important aspect of climate change, yet is complex and often misunderstood. The chapter defines the aim and objectives for and rationale for this research. Chapter two outlines the approaches that were carried out during this study, providing a theoretical background to the research by summarizing both advantages and limitations of the methods. It identifies the important morphological characteristics of foraminifera examined. A literature review places into perspective a combination of previous palaeoecological studies conducted on sea-level change using a variety of proxies during the late Quaternary in southern Africa. The results from such studies in southern Africa are described in chapter three. This chapter sets the context for results obtained from this research. Chapter four provides a description of the study site. It provides background to South African estuaries and salt marshes. The methodology chapter describes the methods used and justification for selection. Chapter six presents and describes the results of foraminiferal analysis, radiocarbon dating and the

transfer function. Chapter seven reconstructs past sea-level change from Kariega and places it into context with previous palaeoecological studies on sea-level change conducted in southern Africa. In the final chapter, a brief summary is provided to the change in sea-level at Kariega during the Holocene and determines to what extent the aim and objectives of this study were achieved.

CHAPTER TWO: FORAMINIFERAL ANALYSIS AND DATING

The reconstruction of coastlines has been promoted by numerous disciplines, using archaeological markers, geomorphological evidence and biological indicators. This chapter provides a theoretical background to the collection and analysis techniques of foraminifera and the reconstruction of sea-level change. Much of this chapter is dedicated to identifying important morphological characteristics of foraminifera, with particular attention given to the fact that foraminiferal assemblages form discrete vertical zones along the elevational gradient of a salt marsh. Finally the chapter outlines different methods used for interpreting the data, the associated advantages and limitations, the use of transfer functions and a brief description of radiocarbon dating.

2.1 Foraminifera

Foraminifera appear in many shapes and sizes, and are found from the deepest parts of the ocean through to shallow tidal flats (Boudagher-Fadel, 2008). Foraminifera evolved during the Cambrian and continued to develop during the Phanerozoic, slowly diversifying to occupy most marine environments (Boudagher-Fadel, 2008). Although foraminifera are unicellular, they perform the life functions of a myriad of multicellular animals. As Goldstein and Watkins (1999: 37) explain “foraminifera and other protists specialized by diversifying subcellular components or organelles to perform these various functions such as eating, moving, growing and reproducing”.

2.1.1 Growth

The majority of foraminifera grow a complex, firm calcite skeleton comprised of a string of chambers known as a shell or test. These tests are resistant to decay and morphologically diverse making them geologically useful in reconstructing past sea levels (Boudagher-Fadel, 2008). Foraminifera can grow in two ways, either by a single chamber increasing in size or by sporadically adding a new chamber to the existing (Goldstein and Watkins, 1999). Single-chambered tests grow continuously, whereas multi-chambered tests grow intermittently with each growth event will be represented by a new chamber. However, when the chambers are viewed one is not able to identify any growth lines (Murray, 1979). The sub-group of benthic foraminifera known as the “larger benthic foraminifera” has a more complicated structure (Boudagher-Fadel, 2008: 07). It can take foraminifera a matter of minutes or days to form a new chamber, but it is very rare that one is able to observe the secretion of a chamber, as the pseudopodium has an important part to play in shaping and putting in place a template around which the new chambers are created. In most cases, growth is genetically controlled although in changing environments such as estuaries, it is common to obtain

environmentally-induced abnormalities. These abnormalities can include smaller chamber sizes because of food shortages or changes in the direction of the coiling of the shell (Murray, 1979).

2.1.2 Foraminiferal tests

Foraminifera are classified according to test composition into three main groups, *viz.*, porcellaneous, hyaline and arenaceous (Murray, 1979; Scott and Medioli, 1986). Porcellaneous foraminifera are calcareous; they appear white under a light source. The pseudopodia protrude out of the last formed chamber of the test instead of out of tiny perforated pores on the test (Brady *et al.*, 1870; Wetmore and Plotnick, 1992). The hyaline or vitreous test is transparent and calcareous. Many pseudopodial apertures may perforate the walls of the shell. Lastly, arenaceous foraminifera are made up of siliceous particles embedded into a chitinous matrix which is a protective substance forming part of the tests outer wall. There are also various other minerals and biogenic material such as shell debris and sponge spicules. The matrix contains a notable portion of peroxide of iron which can give it a reddish colour. The pseudopodia extend from the terminal aperture on the shell (Wetmore and Plotnick, 1992).

Foraminiferal tests will consist of a chain of chambers which are separated by sarcod partition (Figure 2.1). Each partition, or what is known as a septa, is perforated by one or more apertures (Brady *et al.*, 1870). Depending on the species, each chamber may have tiny perforations on the outer wall through which the pseudopodia extend or it may only be emitted out of the aperture on the last chamber to be formed (Wetmore and Plotnick, 1992).

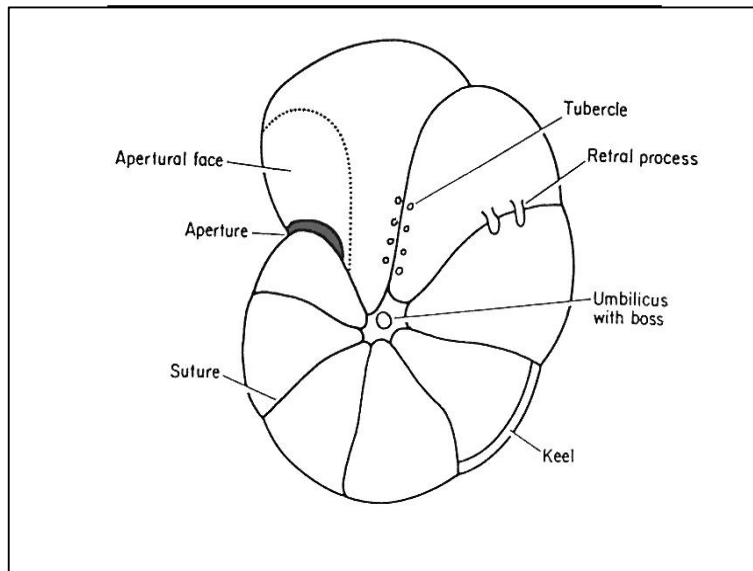


Figure 2.1: General morphological features of a foraminiferal test (after Murray, 1979)

The direction of the chamber is genera-specific with multiple variations unilocular refers to a shell comprising of a single chamber. Uniserial refers to chambers added in a single linear series, while biserial refers to chambers added in a double linear series and tiserial a triple linear series. Planispiral refers to chambers that are added in a coil within a single plane like the chambered nautilus. Trochospiral is similar to that of a snails test whereby chambers are added in a coil that forms a spiral. Milioline refers to an arrangement where each chamber stretches the full length of the test and each successive chamber is placed at an angle of up to 180 degrees from the previous, relative to the central axis of the test (Wetmore and Plotnick, 1992).

The shells of foraminifera comprise of a number of different types of material (Loeblich and Tappan, 1964) which forms the foundation for defining the higher taxonomic levels of this group. These higher groups can then be subdivided by their external morphologies (Culver, 1993). From the test composition one can determine where a range of species or their fossil remains will survive (Scott *et al.*, 2001). For foraminifera that secrete calcium carbonate (CaCO₃) to form their test, the fossilization will depend on whether or not the environment is conducive to carbonate preservation (McCrone and Schafer, 1966). Agglutinated foraminifera form their shell by cementing detrital material and are able to survive in sediments in which no carbonate is present. These areas either have low salinity or the water is much colder and therefore makes the precipitation of carbonate difficult. Thus, as saline conditions increase and the temperature of the water becomes warmer so agglutinated foraminifera are replaced by CaCO₃ secreting forms (Scott *et al.*, 2001). However, this is not the case if the pH is lowered by either a high organic matter concentration or a low oxygen concentration (Schafer, 1973). These conditions are often associated with coastal environments that have become contaminated by pollution (Scott *et al.*, 2001).

Foraminiferal tests are not only used to reconstruct past sea levels but also used in the biostratigraphy of dune deposits. Foraminiferal tests are one of the few tests that are produced by a living organism and transported completely. Therefore, fossil foraminiferal assemblages contained within the dune can be dated and a relative age assigned to strata within the dune, thereby determining when and how it was formed (Franceschini and Compton, 2007). By distinguishing between agglutinated and calcareous foraminifera, one is thus able to separate a marsh into zones of high and low marsh (Gehrels, 1994).

The chemical composition of the shell is important as this could indicate the chemical composition of the water in which it developed (Wetmore and Plotnick, 1992). The chemical composition of foraminifera tests that have accumulated in deep-sea sediments are able to provide a record of past ocean compositions. The palaeochemical record is encoded in the form of the variable isotopic and ion ratio's that exist in the test. The chemical variations between tests can be directly linked to changes in sea water composition. Therefore, the assumption is made that foraminiferal tests are in chemical equilibrium with surrounding sea water (Lea *et al.*, 1995). On a global scale scientists have been able to map past temperatures of the water by measuring

the stable oxygen isotope values of planktonic and benthic foraminifera tests found in deep-sea cores. By mapping past sea temperatures one is able to see how both currents and climate have changed which may help predict future changes (Stanley *et al.*, 1988; Wetmore and Plotnick, 1992).

2.1.2.1 Calcareous tests

Calcareous foraminifera are benthic and found in abundance in the ocean but are relatively rare in intertidal zones. Such species are found in the mudflats and channels of the intertidal zone (Horton *et al.*, 2005b) and will appear white and glassy (Gehrels, 2002). Their tests are usually considerably larger in size, will often present traces of abrasion and dissolution, and may be partially pyritized (Fiorini *et al.*, 2010; Horton *et al.*, 2005b). The dissolution of calcareous tests is common in salt marshes especially during summer, which reflects an increase in productivity within a salt marsh during the summer months (Scott and Medioli, 1980; Horton and Edwards, 2003). Calcareous foraminifera do not preserve well in marsh peats (Gehrels, 1994) and once they die, the protoplasm degrades quickly which makes it difficult to determine which were alive or dead on collection of surface samples (Horton and Edwards, 2003). Such foraminifera that build their shells from carbonate work similar to 'carbon sinks' as they absorb carbon and calcium from the seawater (Wetmore and Plotnick, 1992).

2.1.2.2 Agglutinated tests

Agglutinated foraminifera are associated with salt marshes and estuarine systems (Fiorini *et al.*, 2010; Gehrels, 2002). The majority of agglutinated tests are relatively small and in many cases pyritized or fractured (Fiorini *et al.*, 2010). Agglutinated foraminiferal tests comprise of detrital clastic material, which is a collection of separate particles that are compressed and held together by a lining. When viewed under a microscope, agglutinated foraminifera will usually appear brown or beige in colour. Unlike calcareous foraminifera they are able to survive in acidic conditions, which helps explain their presence in salt marshes (Gehrels, 2002). Agglutinated foraminifera tests are usually the only tests that preserve well in salt-marsh sediments (Gehrels, 1994), and display a parallel with mean tidal levels (Horton and Edwards, 2003; Scott and Medioli, 1986). Such foraminifera exist in distinctive biofacies within a salt marsh (Gehrels, 1994), and usually in vegetated zones of the marsh (Horton *et al.*, 2005a). All foraminifera referred to as intertidal are benthic, will inhabit the sediment surface or just below the surface (Gehrels, 2002).

2.1.3 Habitat preference

Foraminifera are found throughout the ocean with planktonic foraminifera found suspended in the water and benthic living on the sea floors and in brackish and near shore zones (Murray, 1979; Boudagher-Fadel, 2008), they however have very specific habitat preferences (Murray, 1979). As an environment becomes more stable so an organisms population will increase, this is the same for foraminiferal populations, the less variability and the warmer it is so the population increases (Scott *et al.*, 2001).

Areas of high productivity and environmental variability lie between coastal marshes and approximately the middle of coastal shelves, which are known as marginal marine habitats (Sen Gupta, 1999). The greatest diversity of foraminifera occurs in reef environments (Murray, 1991). However, salt-marsh foraminifera are able to endure a much wider range of both daily and seasonal environmental changes, which is reflected by the low variety and changeable composition of foraminiferal assemblages (Leorri *et al.*, 2010). Benthic foraminifera that reside in shallow waters such as salt marshes will attach themselves to a substrate or will live within the first few centimeters of sediment. They will not usually be found deeper than 5 cm and the sediment will need to be oxygenated (Murray, 1979; Gehrels, 2002). Other benthic foraminiferal species will prefer to cling to a solid surface such as rocks, shells or plants using their pseudopodia, and some will permanently cement themselves to a surface (Murray, 1979; Scott *et al.*, 2001).

2.1.4 Patterns of distribution

A number of environmental variables influence the distribution patterns of foraminifera. Temperature had been identified as the main controlling factor for foraminiferal global distribution; however, Cushman (1959) proposed that historical factors such as tertiary migrations could influence modern distributions (Culver and Buzas, 1999).

The majority of foraminifera are planktonic and are globally distributed in the latitudinal temperate belts. They are suspended near the surface of the water along with other marine zooplankton. Major environmental variables controlling planktonic foraminifera are well understood, and include temperature, salinity and the productivity of the upper 100 m of surface seawater. Planktonic species are restricted to the high northern and southern latitudes and absent in between and therefore are said to be distributed along the latitudes (Sen Gupta, 1999). This will be explained further in section 2.2.1.

The second group of foraminifera, benthic, are found at greater depths in the ocean and are attached either to a surface or in the sediment (BouDagher-Fadel, 2008). The distribution of foraminifera is limited when they are exposed to the atmosphere as they are not able to survive outside of water. On a local scale, the distribution of foraminifera is impacted by the availability of different surfaces substrates (Murray, 1979) and on a regional scale, temperature will affect the distribution of species (Cushman, 1959; Culver and Buzas, 1999). However, numbers of benthic foraminifera species appear greater in regions with warmer water. For example, studies conducted on the continental margins of North and Central America has shown that their distributional limits are similar to the boundaries of major water masses in the region (Culver and Buzas, 1999).

Although there are a number of different variables affecting foraminiferal zonations, the distribution of marsh surface foraminifera is similar in all temperate environments and follows the vertical zonations concept (Leorri *et al.*, 2010). The highest tidal flats are dominated by *Jadammina macrescens* and

Trochammina inflata and as the salinity and elevation decreases; *Haplophragmoides* replace the species that were found in the higher tidal levels. However, at the mid-tidal elevations calcareous species become scarcer depending on the accessibility of calcium carbonate within the environment (Leorri *et al.*, 2010). Estuaries are characterized by *Miliammina fusca* and *Ammonia beccarii*, whilst *Reophax moniliformis*, *Spiroloculina excavata* and *Quinqueloculina bicornis* are found in estuary mouths and near shore areas where salinity approaches an average of >10 ppt (Murray, 1979). During the tidal exchange of water, small species and juveniles will be transported from the coastal shelf into deeper coastal waters (Murray, 1979; Leorri *et al.*, 2010). While small benthic foraminifera are found in sediments in salt marshes, larger benthic foraminifera seem to be restricted to more tropical waters with a depth of less than 60 m (Culver and Buzas, 1999).

Benthic foraminifera predominantly occur at the sediment-water interface however, they have been found at depths of up to 10 cm in marine sediments. Even within such a small space, large changes can take place within this microhabitat. Notably, sediment grain sizes will shift from finer sediments near the surface to coarser further down. The sediment-water interface will have a high concentration of oxygen and be rich in nutrients, while at deeper depths there will be little or no oxygen and the organic matter that does exist will have a low nutritional value (Jorissen and Whittling, 1999).

2.2 Application

The following section will discuss the techniques that are applicable to the collection and analysis of salt-marsh foraminifera. Intertidal foraminifera have been used to reconstruct sea-level changes from the Holocene period (Horton *et al.*, 2007), due to the relationship between foraminiferal assemblages, the vertical relationship of the sediment core and past tidal levels, all of which can be quantified (Gehrels *et al.*, 1996; Horton, 1997; Edwards and Horton, 2000). This is achieved by combining information obtained from the foraminiferal stratigraphy and the chronology as determined from AMS radiocarbon dating, to provide a high-resolution sea-level record (Edwards and Horton, 2000). The chronology can be determined by a range of dating methods *viz.*, Pb-210, Cs-137, Am-241, pollen and tephra (Gehrels and Woodworth, 2013).

To predict the impact of future sea-level changes it may be necessary to place the predicted scenarios into a geological framework that has been developed as a result of late Holocene high-resolution sea-level reconstructions (Gehrels *et al.*, 2006; Kemp *et al.*, 2009). Such records are of importance when understanding the mechanisms and climate unpredictability responsible for changes in relative sea level.

2.2.1 Vertical zonation

Foraminiferal assemblages form thin vertical zones along the height gradient of a salt marsh, with the relative abundance of different species correlating with tidal levels in salt marshes (Scott *et al.*, 2001; Gehrels and Newman, 2004). Through an understanding of the vertical distribution of modern foraminifera it

becomes possible to reconstruct former sea levels based on the identification and interpretation of fossil foraminifera (Horton and Edwards, 2003). Benthic foraminifera are considered to be one of the most valuable groups of marine organisms that can be used as sea-level indicators. Foraminiferal environmental distribution shows a discrete vertical zonation which is related to sea level, vegetation cover and salinity of the salt marsh (Franceschini *et al.*, 2005). Salt-marsh foraminifera may be vertically zoned in a similar way to salt-marsh vegetation; however foraminiferal zonations are more compacted (Horton and Edwards, 2003). However, there are differing opinions with regards to foraminiferal distributions being influenced by elevation. Scott and Medioli (1980), Scott and Leckie (1990), and Horton and Edwards (2003; 2006) suggest that foraminiferal assemblages are vertically zoned and are strongly related to elevation. However, according to de Rijk (1995) and de Rijk and Troelstra (1997), foraminiferal distributions are influenced by salinity rather than elevation. For example, the Great Marshes (Massachusetts) display an irregular high marsh surface, where salinity is not controlled by elevation and there is no evidence of vertical zonation (Horton and Edwards, 2006). Studies conducted on salt-marsh foraminifera along the Atlantic coastline of both Canada and the United States of America show evidence of vertical zonation closely linked to tidal levels (Edwards and Horton, 2000), though variations in salinity conditions could complicate the vertical subdivision of foraminiferal assemblages (Edwards and Horton, 2000).

The use of foraminifera as sea-level indicators are justified by the idea “that characteristic foraminiferal assemblages are vertically zoned with respect to the tidal frame, that these modern relationships can be quantified, and that they are representative of past conditions encountered in Holocene salt marsh sedimentary sequences” (Horton and Edwards, 2006: 4). Thus, foraminifera found enclosed in a preserved salt marsh core can be used to determine former sea levels (Horton and Edwards, 2006; Gehrels, 1994). By identifying the modern relationship of surface assemblages of living and dead foraminifera, one can conclude that dead assemblages on the surface will not differ significantly to living assemblages and can be used for palaeoenvironmental reconstructions (Horton *et al.*, 2005b).

2.2.2 Limitations

There are a number of limitations associated with the use of salt-marsh foraminifera as sea-level indicators (Scott and Medioli, 1980; Gehrels, 1999; Horton and Edwards, 2003; Franceschini *et al.*, 2005 and Berkeley *et al.*, 2007). For example, auto-compaction of the older salt-marsh peat can limit the accuracy of dating the sediment core. Older peat can possibly collapse under its own weight, due to decay and dewatering the sediment becomes lighter and starts to sink to a depth much lower than the depth at which it was formed in relation to former sea levels (Douglas *et al.*, 2001).

Post-sampling degradation is possible, in particular with calcareous tests (Scott and Medioli, 1980; Horton and Edwards, 2003). This occurs when sediment samples are exposed to an oxidising atmosphere, causing organic material to oxidate and sulphides to reduce and, ultimately, lead to a change in pH (Berkeley *et al.*,

2008). This can affect the preservation of agglutinated tests as their organic binding cements start to degrade (Berkeley *et al.*, 2007).

A robust training set is important as it reduces the chance of error in the reconstruction. There may be a restriction on the range of the modern foraminiferal analogues and the spatial variability of a small sample size as it is likely to be under-represented (Griffith and Amrhein, 1997; Horton and Edwards, 2006). There has only been a limited foraminiferal distribution database published for South Africa (Franceschini *et al.*, 2005). It is therefore important to create one's own by assessing the distribution of modern foraminifera.

Another limitation may arise when the sampling of modern assemblages is conducted once during the year, as the samples taken on this one occasion may not be in equilibrium with the environment. Furthermore, these may not represent the typical assemblages found at such a point throughout the year or over a longer period of time, thus the reconstruction of past sea levels may not be consistent (Horton and Edwards, 2003).

Under optimum environmental conditions there will be vertical errors linked to surveying, the coring process and the datum selection. It is unlikely that these vertical errors will amount to more than 0.1 m. Age errors are common as many are reported as being twice the standard deviation and therefore once they are calibrated the error can be significant (Gehrels, 1999), however this will depend on the dating technique.

2.2.3 Methodological consideration

2.2.3.1 Field sampling

When considering the development of fossil foraminiferal assemblages it is advantageous to produce a training data set. Living foraminifera represent *in situ* production thereby providing one with an approximation of the assemblage composition of an intertidal environment (Berkeley *et al.*, 2008). Living intertidal foraminifera do not reside solely in the surface layer but are able to survive up to ± 5 cm below the surface (Berkeley *et al.*, 2008), but in order to get a good representation of modern assemblages the top 1 cm is extracted and analysed. Surface samples are extracted using a hand-held corer, which is sharpened with serrated edges to penetrate through the surface material. The hand-held serrated corer (Scott and Medioli, 1980) is inserted into the salt marsh by turning it into the ground and then retrieved by pulling it out directly, thereby avoiding compaction (Jennings *et al.*, 1995; Horton and Edwards, 2006). The altitude of each sample is determined by leveling to the nearest geodetic datum (Edwards *et al.*, 2004).

The effects of environmental factors on the distribution of foraminifera are often not taken into consideration as a result of inadequate field data. The only measurements that are frequently taken are the temperature and perhaps salinity (Bradshaw, 1968). The following environmental factors are important to consider especially when explaining those distributions of foraminifera that are less clear. These factors include; air and water

temperature, light intensity, tidal elevations, oxygen concentrations and pH. Horizontal differences are often due to the varying amounts of vegetation along the surface and therefore it is important to take into consideration the relationship between foraminiferal distributions and vegetation zones (Bradshaw, 1968).

2.2.3.2 Laboratory processing

Foraminiferal samples are stored in a solution of ethanol and protein stain Rose Bengal to identify organisms that were living at the time of collection (Scott and Medioli, 1980; Gehrels, 2002). The solution contains 70% distilled water, 30% ethanol, 1.5 g Rose Bengal stain and 0.5 g sodium carbonate (Scott and Medioli, 1980; Horton and Edwards, 2003; Berkeley *et al.*, 2008). Rose Bengal stains the protoplasm bright red leaving the test walls and organic lining either unstained or lightly stained (Scott and Medioli, 1980; Murray and Bowser, 2000; Horton and Edwards, 2003; Horton and Edwards, 2006). Those foraminifera with stained protoplasm in the last few chambers are assumed to be alive at the time of collection (Horton and Edwards, 2003). The ethanol preserves the organic material and the foraminiferal assemblages, while the sodium carbonate acts as a buffer solution against any pH change that may affect the calcareous test post-sampling (Berkeley *et al.*, 2008).

Before counting, samples are rinsed in water and washed through 63 μm and 500 μm sieves (Scott and Medioli, 1980; Scott and Hermelin, 1993; Edwards and Horton, 2000). The larger sieve is used to remove any organic matter and coarse detritus; the remaining sediment on the smaller sieve will be retained for foraminiferal analysis (Culver and Horton, 2005; Berkley *et al.*, 2008). The retained sediment sample is subdivided into eight aliquots using a wet splitter (Gehrels, 2002).

The use of a wet splitter is a dependable and consistent method for the separation of foraminiferal assemblages (Scott and Hermelin, 1993; Gehrels, 2002). Such a process can split a sample that contains thousands of individuals into smaller equivalent subsamples (Scott *et al.*, 2001; Horton and Edwards, 2006). There is a trade off between the number of samples that can be analysed and the level detail at which each sample is analysed. Thus, the use of the wet splitter allows more subsamples to be analysed in an equivalent time.

2.2.3.3 Counting and identifying

Most foraminiferal fossils fall between 63 and 500 μm in size, thus can be observed under a low-power stereomicroscope. Foraminifera are counted wet under a microscope, thus preventing aggregation of organic particles and ensuring that stained tests are recognizable (Edwards and Horton, 2006; Berkeley *et al.*, 2008). Counting foraminifera while wet avoids preservation concerns, as calcareous foraminifera often disintegrate easily after death (Murray and Alve, 2000) due to desiccation.

Some researchers advocate counting the total assemblage (living and dead) to provide the best indication of environmental conditions as both season and temporal fluctuations are integrated (Scott *et al.*, 2001; Horton *et al.*, 2005a). According to Horton *et al.* (2005a), in a temperate environment it is important to exclusively use dead assemblages, as the dead assemblages are not susceptible to seasonal variations and will more accurately resemble subsurface assemblages. Dead assemblages represent a time averaged crop and are not affected by seasonal changes (Murray, 1979; Horton and Murray, 2007).

Approximately 250 individual foraminifera per sample are counted, providing a good indication of foraminifera present in the sample (Horton *et al.*, 2007). From a statistical perspective a count of 250 individuals is necessary especially if significant indicator species are present but in low numbers (Gehrels, 2002). A low count may suffice if indicator species are abundant but species diversity is low. However, in the upper reaches of a salt marsh, foraminifera may be absent, therefore the desired number may not be obtainable (Gehrels, 2002).

2.2.4 Data Analysis

Where changes in foraminiferal zonation are identified it becomes possible to reconstruct the changes that occur in the elevation of the marsh surface relative to sea level. The distribution of present day foraminifera on the surface of the marsh are quantified and the information obtained is applied to the core. This will result in foraminifera samples found in the core being given a former surface height at which it once would have existed (Gehrels, 2002). To construct past sea levels from fossil foraminifera there is a two step procedure. The first step is to establish the relationship between variables, whereby a regression analysis is used to create a model from the training set of surface foraminiferal samples. The second step is the calibration, which is used to infer past heights of salt marshes surface from the fossils. The calibration is conducted using a set of mathematical equations, which form a transfer function (Gehrels, 2000; 2002). To determine the most appropriate regression and calibration technique, a multivariate statistical analysis is used. This determines the relationship between total surface assemblages and mean tidal levels or elevation. This is achieved by performing a Detrended Canonical Correspondence Analysis (DCCA) using CANOCO (Ter Braak, 1995); which estimates the gradient lengths in standard deviation (SD) units (Birks, 1995). If the gradient lengths are >2 SD units then a unimodal response model is appropriate, whereas if they are <2 SD units a linear response model is used (Birks, 1995; Massey *et al.*, 2006).

A transfer function is an acceptable depiction of a linear time invariant dynamical system (Birks, 1995). Mathematically, a transfer function would be expressed as a function of complex variables, whereby the input-output relation for a linear system is described. For example, a transfer function quantifies the relationship between the chosen environmental variable and the environmental proxy so that the environmental variable can be expressed as a function of the environmental proxy. A number of transfer functions have been developed to quantitatively reconstruct palaeoenvironmental variables (Horton and

Edwards, 2006; Leorri *et al.*, 2008), and therefore a foraminiferal transfer function is used to reconstruct past tidal levels through the use of fossil assemblages (Horton, 1997). Such a transfer function is a statistical method that is capable of using ecological information, which is contained in modern distributions of foraminifera within an intertidal environment, and then using this information to acquire palaeoecological data from the fossil assemblages (Edwards and Horton, 2000).

The fossil foraminiferal data that were extracted from the core is calibrated using the transfer function, to predict water level values for each sample. In addition to this, a matching analogue test is performed for each of the samples. This is conducted to identify the similarities between the fossil assemblages and the modern assemblages that form the training set (Edwards and Horton, 2000). Such an approach provides a quantitative and strong methodology for reconstructing past changes in sea level from intertidal environmental deposits and using intertidal foraminifera as indicators (Gehrels, 2000).

Given that intertidal foraminiferal distribution is influenced by surface elevation, one is able to use foraminifera as proxies for elevation. This provides a means whereby faunal data are converted into environmental data, in this instance, elevation. There are three stages to constructing a foraminiferal transfer function; compiling a training set, fauna environmental regression and the calibration of the fossil samples (Horton and Edwards, 2006). For the purpose of this research, C2 (Juggins, 2003) was used to calculate regression statistics.

For species rich data whereby the gradients are >2 SD units, unimodal regression models are used; weighted averaging (WA), weighted averaging with tolerance down weighting (WA-Tol), weighted averaging partial least squares (WA-PLS) (Massey *et al.*, 2006). Species that respond unimodally to the environment or the samples that are not perfectly distributed along the environmental gradients then the WA is most suitable (Gehrels, 2000; Massey *et al.*, 2006). WA-Tol attempts to give more weight to those species that have a relatively narrow range as they are considered to be better indicators (Gehrels, 2000). WA-PLS is an adapted version of WA. It uses the structure of its WA residuals to better predictions (Gehrels, 2000; Massey *et al.*, 2006). Regression statistics are calculated using WA and WA-PLS models. As Massey *et al.* (2006: 218) explains, the “regression statistics produced by C2 include the root mean square error prediction, the coefficient of determination (r^2) and values for the maximum bias which is the difference between mean observed and predicted values along the gradient.

2.3 Radiocarbon analysis

Radiocarbon dating is only applicable to the last 50,000 years; however it was one of the first radiometric techniques and is the most widely used techniques. After World War Two, Willard Libby started investigating whether or not radiocarbon existed in biological materials. Radiocarbon dating was born and

has revolutionized researcher's views on the late Quaternary (Walker, 2005). The carbon atom may be the building block for life, but also provides us with a means of dating life (Walker, 2005). There are three isotopes of carbon, carbon 12 and carbon 13 and carbon 14, which is radioactive. Carbon 12 and carbon 13 are both stable isotopes whereas radiocarbon (^{14}C) will decay becoming a stable form of nitrogen, ^{14}N by the emission of beta particles. For every one radiocarbon atom that decays a beta particle is released from the nucleus (Broecker *et al.*, 1984).

Within all living organisms, the radiocarbon isotope exists (Pilcher, 1991), even in the ocean as it is absorbed as dissolved carbonate. This means that organisms that reside in the ocean are able to take up radiocarbon during their lifecycle (Broecker *et al.*, 1984; Pilcher, 1991). As Walker (2005) explains once an organism is dead it becomes isolated and is no longer able to take up radiocarbon and the 'radiocarbon clock' starts to decrease as a result of radioactive decay taking place at a constant rate. Thus, the age of an organism can be determined by measuring the total amount of radiocarbon that remains in the fossil material and then comparing this to the amount of modern radiocarbon that is present in standard material. Experiments conducted on the rate of decay of radiocarbon have shown that the decay rate is 1% every 83 years (Williams *et al.*, 1993; Walker, 2005). Using the rate of radiocarbon decay by calculating the current radiocarbon content with the exponential rate of decay of radiocarbon taking into consideration the maximum age of 50,000 cal years Before Present (BP), fossil organism's ages can be determined (Fairbanks *et al.*, 2005; Ascough *et al.*, 2009). Radiocarbon, like all radiocarbon isotopes, does not have a linear decay curve but an exponential curve (Figure 2.2; Walker, 2005).

There are two approaches to measuring radiocarbon activity in samples, the first being beta counting which involves detecting and counting the beta emissions from radiocarbon atoms. This is based on the principle that the emission level of beta will reflect the remaining level of radiocarbon activity from a specific sample (Walker, 2005). The second approach is accelerator mass spectrometry (AMS), this employs particle accelerators as mass spectrometers, which are able to count the number of radiocarbon atoms contained in a sample (Libby, 1955; Gehrels, 1999).

The advantage of using AMS dating over beta counting is the small amount of sample required and the fact that the sample can be processed effectively and speedily (Walker, 2005). However, with radiocarbon dating there are limitations, which include a limited timescale, the calibration curve and the possibility of contamination (Barber and Charman, 2005). The radiocarbon timescale is restricted to the last 50,000 years, making this the biggest limitation for Quaternary modeling (Huntley, 1996). Contamination refers to the addition of either young or old carbon to a sample (Walker, 2005). Sources of contamination may include; the introduction of carbon from roots growing down through a peat or soil profile introducing younger materials to older horizons, carbon being exchanged between ground water and relic shells and finally isotopic fractionation resulting in age determination errors (Walker, 2005; Blockley *et al.*, 2007).

Contamination may also occur post field sampling during storage in the laboratory whereby there is either fungal growth, dust or skin contamination (Walker, 2005).

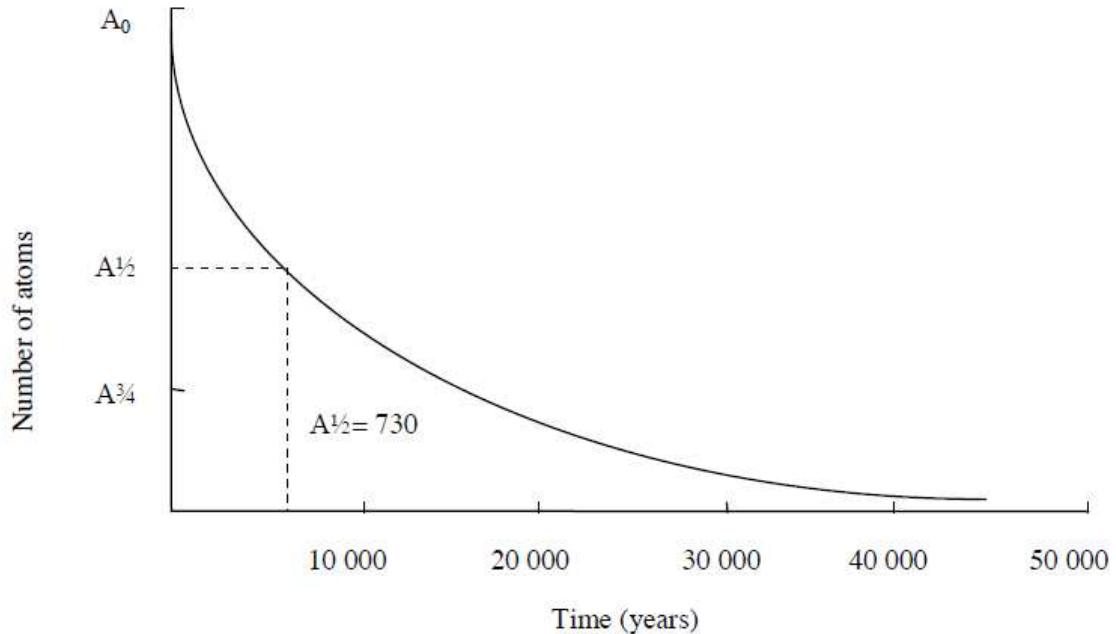


Figure 2.2: Radiocarbon decay curve is exponential, therefore the percentage decrease in number of atoms in a given unit of time is constant. After each half life the number of atoms remaining are halved. If there are A_0 atoms of radiocarbon at the beginning of the decay process, then after one half-life there will be $A_0/2$ atoms remaining (Adapted from Walker, 2005)

Over time there have been many improvements made to the sampling, dating and calibration techniques, ensuring that results are more accurate and reliable. AMS dating in itself has improved the accuracy in dating samples with smaller errors compared to those of radiometric dating (Andrews, 1998; Walker, 2005).

Calibration is a critical part of the analysis and needs to be undertaken in the early stages of the radiocarbon analysis. The calibration process allows the fluctuations in the production of radiocarbon to be corrected (Pilcher, 2005), by comparing the radiocarbon dates with dates derived from alternative methods. There are separate radiocarbon calibrations for each hemisphere as the atmospheric concentrations of radiocarbon in the southern and northern hemisphere are different; much effort has been made to lengthen the radiocarbon calibration curves to cover the complete radiocarbon timescale (Pilcher, 2005). Both the international calibration model (IntCal09) and the southern Hemisphere calibration model (SHCal04) were developed using proxy data (McCormac *et al.*, 2004; Reimer *et al.*, 2009; Blaauw, 2010). The southern Hemisphere calibration model, due to a lack of records and the changes in the carbon reservoirs during the Holocene, should not be used to date fossils beyond 11,000 cal yr BP (McCormac *et al.*, 2004). The datasets that are

incorporated into these calibration models result from dendrochronological and marine records (Walker, 2005).

When marine samples such as shells or marine mammal residues are dated the radiocarbon ages are substantially older than terrestrial samples (Dumond and Griffin, 2002). Theoretically, the concentration of radiocarbon in the atmosphere is equivalent to that in the ocean and biosphere. Terrestrial organisms take in carbon dioxide by consuming plants, which have absorbed CO₂ via photosynthesis (Mooney *et al.*, 1987), whereas marine organisms take in radiocarbon through an exchange process of radiocarbon in the atmosphere and the ocean and from the radioactive decay that is taking place at different levels in the ocean. Therefore the concentration present in all marine organisms is not the same (Oeschger, 1985; Dewar *et al.*, 2012). As a result, marine organisms living in different water masses will have different ages to those of terrestrial organisms (Dewar *et al.*, 2012). Therefore, if a marine organism and a terrestrial organism of the same age were compared they would not have the same concentration of radiocarbon and would appear to have very different radiocarbon ages, varying by up to 400 years (Dumond and Griffin, 2002). This marine radiocarbon offset that is present is even more obvious along the west coast of countries where upwelling is greater (Dewar *et al.*, 2012). The differing age of sea water is predominantly larger due to different patterns and movement of the water; surface currents and upwelling (Walker, 2005). The shape of the coastline, ocean topography and the local climate will also affect the age (Dumond and Griffin, 2002). Though these changes do not only vary in space but also in time, thus making it difficult to determine correction factors for fossil material found at different localities. Fortunately this offset can be corrected through formulated methods that better calibrate radiocarbon ages that have been obtained from marine organisms (Dewar *et al.*, 2012), though this problem has not yet been fully resolved (Walker, 2005).

To determine the age of marine organisms that have built their skeletal structures out of dissolved carbonate such as foraminifera and mollusks, the determined date needs to be corrected to the age of the sea water, in which they lived (Dumond and Griffin, 2002; Walker, 2005). There are many limitations associated with radiocarbon dating of shell carbonates, as carbonates are fairly soluble and chemically interact with the environment, with the implication that dating accuracy cannot be guaranteed (Dumond and Griffin, 2002).

In order to calibrate the ages from marine samples the global marine reservoir effect $R(t)$ and the local marine reservoir correction (ΔR) need to be used. The difference between the concentration of radiocarbon between the global and regional surface ocean will vary over time (Stuiver *et al.*, 1986). Off the west coast of southern Africa, oceanic conditions are complex as there is a mixing of two water bodies, which will have an influence, thus determine the value of ΔR (Dewar *et al.*, 2012).

CHAPTER THREE: LITERATURE REVIEW

With sea-level change a major and evident consequence of climate change, affecting both the physical environment and societies, this chapter discusses the physical causes of sea-level change, past regional and global sea-level changes and predictions of changes. It seeks to clarify the relationship between change in sea levels and the climate and draws on other palaeoecological studies conducted on global sea-level change. Finally, this chapter reviews studies of sea-level change in southern Africa during the Holocene, which will place the results from this study into a local and global context.

3.1 Introduction

According to Church *et al.* (2001: 643), mean sea level is defined “as the height of the sea with respect to a local land benchmark, averaged over a period of time, such as a month or a year, long enough that fluctuations caused by waves and tides are largely removed”. The rise and fall of present day sea levels is of significant interest due to the impact it could have on human populations residing along coastal regions and islands (IPCC, 2011). It has been estimated that there are approximately more than 200 million people living in coastal regions that are vulnerable to flooding by extreme sea levels (Nicholls and Cazenave, 2010). The IPCC report has indicated a global increase of *c.*3 mm/year. This trend is already evident in South Africa, with an increase of 2 mm/year record from Cape Town (Cartwright, 2008). Sea levels have been changing for centuries, some changes have been rapid while others much slower (Pugh, 2004; IPCC, 2011), while some have occurred on a local scale others a global (Pugh, 2004). Future predictions by the IPCC for the next 80 years range from a minimum of 0.18 m to a maximum of 0.59 m (IPCC, 2011).

The global climate system is dynamic and it's affect on sea-level rise is complex (Oldfield, 2005; Aral *et al.*, 2011). Understanding this complex climatic system and how it might behave in the future is of great importance if future sea levels are to be predicted (Oldfield, 2005). A century long record of global sea-level fluctuations has been documented by the IPCC (IPCC, 2011). The changes in global sea levels, is the result of two major processes which are related to the climatic changes that have recently occurred causing the water levels in the ocean to alter (IPCC, 2011). These processes operate at different timescales from a matter of seconds to millions of years (Church *et al.*, 2001; Lambeck and Chappell, 2001). The first process is thermal expansion and the second, the exchange of water between the ocean and polar ice caps, melting glaciers (Church *et al.*, 2001; IPCC, 2011), and anthropogenic changes that have taken place resulting in both land and atmospheric hydrological changes (Church *et al.*, 2001; Goudie, 2006; Nicholls and Cazenave, 2010; IPCC, 2011). There are predictions that sea levels in 2100 may be similar to those of 130,000 years ago, which happen to be several meters higher than present sea level (Overpeck *et al.*, 2006).

3.2 Late Quaternary sea-level change

Historical records indicate a rise in atmospheric temperatures along with rising sea levels, bringing the temptation to directly link the increase in sea levels with climatic change and particularly in recent times the warming of the ocean (thermal expansion). However, the reliability of this relationship is questionable as the overlap between temperature observations and tidal gauge recordings is relatively short. To determine more reliable trends between climate and sea-level changes it is important to look at this relationship over a longer period of time, using palaeorecords (Oldfield, 2005). Relative sea-level changes are partially caused by climatic factors, though there are other contributing factors such as tectonic forces and sediment loading (Anderson *et al.*, 2007). Palaeorecords of historical sea levels are therefore the basis of our understanding the relationship between sea level and climatic processes (IPCC, 2011). For example, Siddall *et al.* (2003) present a detailed reconstruction, using 470,000 year old sediment records that were taken from the Red Sea. Siddall *et al.* (2008) used isotopic records in conjunction with fossil foraminifera preserved in the sediment to determine at what depths the Red Sea was opened to ocean, from this sea-level changes can be inferred (Siddall *et al.*, 2003; Siddall *et al.*, 2008; Oldfield, 2005). This sequence tracks changes in sea level between 70,000 - 20,000 cal years BP, which are closely related to temperature records from the Antarctica. Palaeorecords have assisted in improving the error in estimations of past global sea levels that occurred during the transition between the Last Glacial Maximum (LGM) and the Holocene (IPCC, 2011). The Holocene has a much milder climate compared to that of the former glacial period (Lambeck *et al.*, 2002).

The LGM occurred between 22,000 to 19,000 years ago (Anderson *et al.*, 2007), during which time thick ice sheets extended across the high latitudes of Europe and North America and the ice sheets in the Antarctic were more widespread than they are presently (Lambeck *et al.*, 2002). Sea levels during this period were at their lowest, approximately 120-135 meters lower than modern sea levels. There is a general consensus that during the LGM continental shelves were elevated especially in glacial regions (Clark and Mix, 2002). Some evidence of the LGM is obtained from deep sea sediment cores (Anderson *et al.*, 2007). After the last glacial, global ice volumes decreased by approximately 10% within a few hundred years (Yokoyama *et al.*, 2000; Anderson *et al.*, 2007). At this point sea levels continued to rise 15 mm per year until 9000 years ago, with modern sea levels being reached approximately 7000 years ago (Lambeck *et al.*, 2002).

It was first thought that at the end of the LGM there were rapid and distinct shifts from higher to lower global ice volumes. However, according to high-resolution proxy data, it was in fact far more complicated. Over the last glacial transition period there were oscillations in the climate and glaciers did not retreat at a constant rate. Both warm and cool phases interrupted the rate at which glaciers retreated (Anderson *et al.*, 2007). According to Lambeck *et al.* (2002: 343), “sea-level change caused by the growth or decay of ice sheets is spatially variable because of the adjustment of the earth's surface to the time-dependent ice-water load and because of the changing gravitational potential of the earth-ocean-ice system”. Both environmental

and climatic changes during the LGM have been the focus of most recent research as high-resolution records become more readily available and dating methods improve (Anderson *et al.*, 2007).

Presently, ice caps cover parts of the Antarctic, Greenland, parts of North America and north Europe and other areas in higher latitudes and altitude (Lambeck, 2002; Oldfield, 2005). There can often be delays of several decades between changes in the local climate and glacier responses (Oldfield, 2005), therefore past climatic changes are influencing the present day melting of polar ice caps in Greenland and the Antarctica (IPCC, 2011). Evidence suggests that over the last few thousand years sea level has changed significantly by a few meters even though the majority of the ice sheets 7000 years ago had completely melted (Lambeck, 2002). With the sea-level rising, previously exposed continental shelves are now covered (Petit-Maire, 1986).

Siddall *et al.* (2003) provided a reconstruction of sea-level changes over the last 6000 years. This included results from the melting of major ice sheets and studies on coral reefs (Siddall *et al.*, 2003). The records from the past 1000 years was of interest as around 700 years ago, the records show a peak in levels (Oldfield, 2005).

Geological evidence indicates that there have been major temporal and spatial variations over the last 10,000 to 20,000 years concerning relative sea-level change (Church *et al.*, 2001; Lambeck and Chappell, 2001). Not surprisingly, there are many excellent case studies of sea-level change, including Nunn (1998), Church *et al.* (2001), Gehrels (2002), Gehrels *et al.* (2002), Lambeck (2002), Goodwin (2003), Siddall *et al.* (2003) and Lambeck *et al.* (2004). These studies reflect the importance of the use of proxies as indicators of sea-level change. These case studies collectively summarize evidence for sea-level changes in the northern and southern hemisphere over the last 200,000 years.

Both temporal and spatial variations in relative sea-level change may be the result of melting ice sheets from the Pleistocene or continental drift (Nakada and Lambeck, 1989). However, in the late 17th century some of the coldest years since the last glacial period were experienced. This period brought about the introduction of instrumental records for recording changes in the climate. Such records documented severe cold periods in places such as Holland whereby canals were frozen for long periods of time (Douglas *et al.*, 2001).

Using dated microatolls from the Southern Cook Islands, which are found only in corals that grow in lower intertidal zones along with indicators of sea level from salt marshes and rocky coastlines, Goodwin (2003) provides evidence that links sea level and climate for the last 2000 years (Oldfield, 2005). Goodwin's (2003) work correlated with Nunn's (1998) reconstruction of sea level from the last 1300 years for the southwest Pacific. Nunn's (1998) study showed that during the last 1000 years, the sea level in the Pacific fluctuated, mimicking sea-level fluctuations of the Holocene. Using collected data from the southern African coastline

and taking into considerations, statistical uncertainties, calibrating radiocarbon dates, Nunn's (1998) sequence suggested changes in relative sea level up to 2 m with a maximum high around and prior AD 1300 and a low around AD 1800 (Oldfield, 2005). From this study it is suggested that sea level in the Pacific would continue to rise through to the end of the 21st century (Nunn, 1998).

Studies conducted along the east coast of North America indicate that in the last 200 years there has been a rise in sea level of approximately 0.2–0.3 m. A study in the Gulf of Maine using salt-marsh foraminifera shows parallels with the Pacific Ocean sea level study, as it shows a peak in sea level before AD 1300 followed by a low until approximately AD 1800 and thereafter a steep rise (Gehrels *et al.*, 2002; Oldfield, 2005). Evidence suggests that sea level in both the north and southern hemispheres was 0.2-0.5 m lower between AD 1400 and 1850 than is in the 21st century, with a highstand of approximately 0.5-1.0 m between AD 1000 and 1300 years (Goodwin, 2003; Oldfield, 2005).

Australia as a continent is relatively stable making it a suitable study site for late Pleistocene and Holocene sea-level change studies (Lambeck, 2002). Sites selected in Australia had little tidal change, no barriers or features along the coastline that may influence local conditions and they needed to possess biological indicators that have a consistent relationship with the ocean (Church *et al.*, 2001). Studies conducted along the Australian coastline have indicated that sea level over approximately the last 6000 years has remained constant with only a variation of a few meters (Lambeck and Nakada, 1990, Lambeck, 2002).

A study conducted in Tasmania using the vertical zonations of salt-marsh foraminifera as a proxy indicated that sea level during the 20th Century had risen faster compared to earlier records. This work conducted by Callard *et al.* (2011), was the first in the southern hemisphere to demonstrate a faster sea-level rise than previous centuries. Tasmania is of interest to sea-level researchers as the oldest tidal benchmark in the world is located here and some of the southern hemispheres first sea-level observations were collected here (Callard *et al.*, 2011). Using observations from this site the rate at which sea level is rising was reconstructed and it was found that sea level was rising ± 0.2 mm per year between AD 1841 and 2003 (Pugh, 2004; Callard *et al.*, 2011). This rate was found to be in accordance with tide gauge records from New Zealand and Australia. From the study in Tasmania using foraminifera, sea level was reconstructed with approximately a 0.10 m precision to those reconstructions using tide gauge records (Callard *et al.*, 2011).

Sea level is expected to rise beyond AD 2100, however the effect global sea level rise will have on coastlines will depend on the locality and geomorphological factors (Oldfield, 2005). These factors include; land movement, lithology, supply and transportation of sedimentation, tidal regime, atmospheric pressure and wind and wave conditions (Zhang *et al.*, 2003; Oldfield, 2005). Although sea level poses many threats to coastal populations, there is no way to assess the likely impacts on the regions (Oldfield, 2005), the biggest concern is the future of low lying islands such as the Maldives as they are highly vulnerable (Oldfield, 2005).

Table 3.1: Summary of past rates of sea-level change

Period	Period of time	Study region	Source of data	Total change (cm)	Sea-level rise rate (mm/yr)	Reference
Late Holocene	50 - 2000 BP	Mediterranean	Archaeological/GIA modeling	13	0.07	Lambeck <i>et al.</i> (2004)
Late Holocene	1500-1600 BP	ME coastal marsh	Coastal marshes	15	1.5	Gehrels <i>et al.</i> (2002; 2005)
Historical	1950-2009	Global	Tide gauge-satellite modeling	10	1.7	Nicholls and Cazenave (2010)
Interstadial-MIS3	60000-25000BP	Red sea, deep sea	Σ 18 O	3000-4000	15.5	Siddall <i>et al.</i> (2008)
20th Century	1841-2003	Tasmanian	Salt marsh foraminifera	-	0.2	Callard <i>et al.</i> (2011)
1000 years BP	last 1000 years	Pacific islands	Various (archaeological)	2000	-	Nunn (1998)

(Adapted from Nunn, 1998)

3.3 Holocene sea-level change in South Africa

During the last 7000 years the sea level along the southern African coast has fluctuated no more than 3 m (Miller *et al.*, 1995; Ramsay, 1995; Baxter and Meadows, 1999; Compton, 2001). As Woodroffe *et al.* (2005: 29) explain, “Rapid changes in sea level are part of a complex pattern of interaction among the oceans, ice sheets and solid earth, all of which have different response time scales”. In the last 350 years, South Africa has only experienced six minor earthquakes with a magnitude range of 5.5 - 6.3, therefore the southern African coast is not considered to be an active margin (Compton, 2001). Tectonism was therefore unlikely to have played an active role in sea-level changes and it is possible that regional isostasy would have more of an effect on the southern Africa coastline (Miller *et al.*, 1995). With the southern African coastline being tectonically stable throughout the late Quaternary, Miller *et al.* (1995) suggests that sea-level change of the coast of South Africa would have been marginally affected by post-glacial eustatic rise at the end of the Pleistocene and the early parts of the Holocene. It has been documented that during the early parts of the Holocene, sea-levels globally rose rapidly in response to melting ice sheets (Carr *et al.*, 2010; Norstrom *et al.*, 2012).

Isostatic processes might explain why sea-level data for South Africa shows such variations along the coastline (Compton, 2001; Miller *et al.*, 1995). Much evidence points to the fact that globally the sea-level rose significantly during the mid-Holocene (Miller *et al.*, 1995). Miller *et al.* (1995) and Ramsay (1995) have reported that sea level during the Holocene was higher than present along the South African coastline,

however very few of their records were accurately dated. Relative sea-level curves have been constructed for a large portion of the coastlines in the northern hemisphere but very few have been constructed in the southern hemisphere (Pirazzoli and Pluet, 1991; Goodwin, 1998). With knowledge of fluctuations having taken place, it is important to look at local shifts in sea-level changes. Researchers are now focused on regional shifts using a variety of methods (Table 3.1) to determine past sea levels, though some vastly contradicting results have been presented. Such evidence from these various locations suggests that South Africa's sea level follows that of the Caribbean, however, deviations do occur at times when the sea level is at its highest (Ramsay and Cooper, 2002). A range of proxies and palaeoenvironmental indicators (Table 3.2) have been used in southern Africa to reconstruct Quaternary sea-level changes (Figure 3.1) (Ramsay and Cooper, 2002; Norstrom *et al.*, 2012). Quaternary sea-level indicators radiocarbon dates are tabulated for South Africa in Ramsay and Cooper (2002).

Sea-level curves based on observational data for southern Africa indicate that the Holocene sea-level maximum took place between 6500 cal years BP (Miller *et al.*, 1993; Compton, 2006) and 5000 cal years BP (Ramsay, 1995; Baxter and Meadows, 1999; Ramsay and Cooper 2002). Observed fluctuations during the late Holocene in southern Africa were relatively small (approximately 1-2 m); however they still had a major impact on past coastal environments (Miller *et al.*, 1993; Ramsay, 1995; Baxter and Meadows, 1999 and Compton, 2001).

Radiocarbon dating of beachrock fossils has been the main source for studies in South Africa on the topic of Holocene sea-level change and is said to be the most accurate (Ramsay and Cooper, 2002). However Ramsay and Cooper (2002) are not supported by Marker and Miller (1993) whose study was conducted at Knysna. This study involved dating shells that were deposited on top of sandy silts that were of marine origin. From this it was identified that 6000 cal years BP the relative high sea level along the south coast ranged between +2.8 meters and +3.8 meters (Marker and Miller, 1993). This corresponds with findings at Keurboom estuary which indicates a +1.7 to +2.7 m sea level 5900 cal years BP (Ramsay, 1995). On the other hand, studies conducted along the west coast of South Africa (Compton, 2001), which involved the dating of salt-marsh peats, agree with previous studies along the west and south coast, including salt marsh deposits at Verlorenvlei (Baxter and Meadows, 1999), tree stumps at Knysna (Marker, 1997) and estuarine shells at Knysna (Marker and Miller, 1993 and 1995). According to Compton (2001), there is evidence from both the west and south coast of South Africa that supports the mid-Holocene sea-level maximum and appears to have occurred approximately 1000 cal years BP before it occurred on the east coast of South Africa. By 5700 cal years BP at Verlorenvlei and 4900 cal years BP at Langebaan the maximum mid-Holocene sea levels on the west coast had returned to ± 1 m of present day levels (Baxter, 1997; Compton, 2001), though tidal gauge records over the last decade indicate that sea levels have risen approximately 1.2 mm/year in South Africa (Brundrit, 1995). Thus, with the current trends taking place sea-level rise is

expected to increase, with estimated rises of 12.3 cm by 2020, 24.5 cm rise by 2050 and a 40.7 cm rise by 2080 (Nicholls *et al.*, 1999; Kirwan and Murray, 2008).

Table 3.2: Proxies used to reconstruct sea-level fluctuations along the southern African coastline

Reference	Proxy	Site
Ramsay, 1991	Coral in beach rock	Kosi Bay
Marker and Miller, 1993	Estuarine shells	Knysna
Marker and Miller, 1995	Lagoonal and estuarine desposits	South Africa
Brundrit, 1995	Tide gauge	South Africa
Ramsay, 1995	Beach rock	East coast
Marker, 1995	Tree Stumps	Knysna
Baxter and Meadows, 1999	Estuarine deposits	Verlorenvlei
Compton, 2001	Salt marsh sediment	Langebaan Lagoon
Ramsay and Cooper, 2002	Beach rock	South Africa
Franceschini <i>et al.</i> , 2005	Foraminifera	Langebaan Lagoon
Compton, 2006	Mollusc and shell fragments	Bogenfel Pan (Namibia)
Nortrom <i>et al.</i> , 2012	Diatoms	Macassa Bay (Mozambique)

The study conducted by Baxter and Meadows (1999) has suggested that over the last 20,000 years the rise and fall in sea level has had a major impact on the geomorphology, ecology and archaeology in the area, therefore these coastal environments may have endured rapid changes (Baxter and Meadows, 1999; Miller *et al.*, 1995).

From studies conducted along the west coast (Miller *et al.*, 1995; Baxter and Meadows, 1999), it is evident that the coastal geography during the mid-Holocene was considerably different especially during times of high sea levels when the topography would have been subject to strong wave action. The sandy southwestern coastline of South Africa would have been reshaped during the late Holocene and only recently has its present day form been achieved (Miller *et al.*, 1995). Reshaping of the coastline is in reaction to the drop in mean sea level during the late Holocene; however this is still taking place at St. Helena Bay along the west coast (Miller *et al.*, 1995). It is apparent that environments found along South Africa's coastline have been influenced significantly by the rise and fall of the sea level during the Holocene (Baxter and Meadows, 1999) which rather modest change in sea level of approximately 2-3 meters dramatically remodeling local coastlines (Miller *et al.*, 1995).

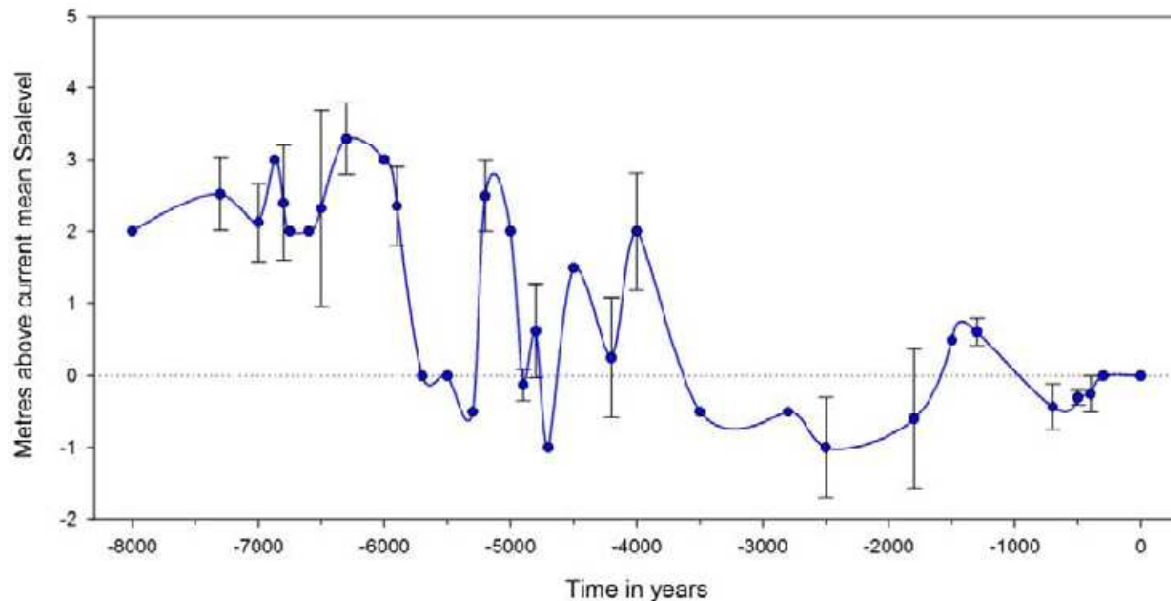


Figure 3.1: Sea-level curve compiled from studies along the South African coast dating back the last eight thousand years. The error bars indicate the variation in the data from the various studies (Miller *et al.*, 1995; Ramsay, 1995; Baxter and Meadows, 1999; Compton, 2001; Ramsay and Cooper, 2002; Compton and Franceschini, 2005; Compton, 2006) (after Wilken, ud)

A recent study from an estuary at Macassa Bay, southern Mozambique using multiple proxies, reflects shifts between marine and freshwater influences during the last 6600 cal years BP. The proxies used include an analysis of fossil diatoms, isotope composition of the organic component, C/N ratio and mineral magnetic properties. The fluctuations between freshwater and marine water can indirectly be linked to changes in relative sea level (Norstrom *et al.*, 2012). During the last millennium, Macassa Bay has become a terrestrial biome with little marine influence. The records for this site reveal two phases associated with sea-level highstands and greater marine influence, which occurred between 6600 and 6300 cal years BP; and 4600 and 1000 cal years BP (Norstrom *et al.*, 2012). According to the sedimentation analysis between the two highstands, the sea level was 1.1 m lower relative to present levels. The results from this study generally support those from the western and southern coastlines (Compton, 2006; Norstrom *et al.*, 2012), but do not concur with those of Ramsay and Cooper (2002) from the northeastern coast of South Africa (Norstrom *et al.*, 2012).

Studies along the South African coastline suggest that sea levels along the coast have shifted approximately 3 m from the current level over the last 8000 years. Generally, evidence of sea-level rise from along the coast of South Africa supports a highstand from 6000 to 4000 cal years BP and a lowstand from 3000 to 2000 cal years BP (Miller *et al.*, 1995; Compton, 2001). Evidence from the west coast suggests that there was a highstand from 1500 to 1300 cal years BP of approximately +0.5 m followed by a lowstand of -0.5 m from

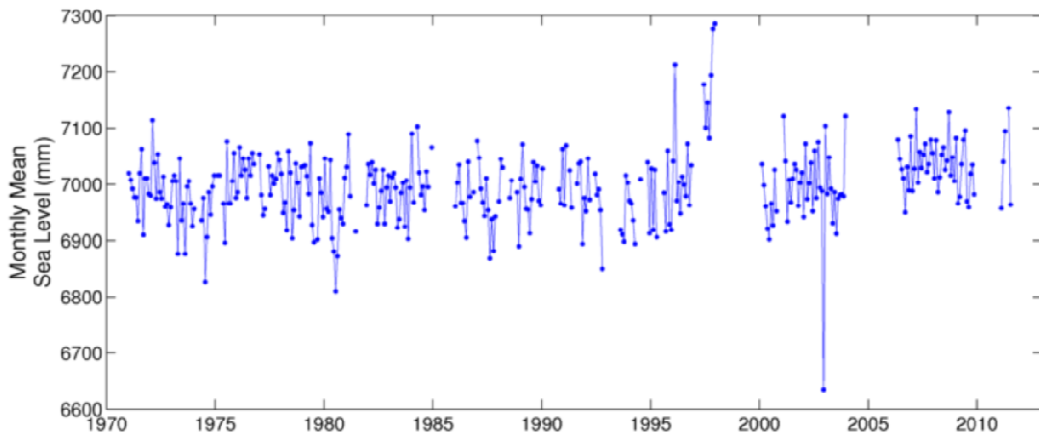
700 to 400 cal years BP (Compton, 2001). A lowstand on the south coast at 700 cal years BP was indicated from *in-situ* tree stumps exposed at low tide at Knysna (Marker, 1997). Proxy evidence presented by Miller *et al.* (1995) for the southwest coast is consistent with previous sea-level history, post the early Holocene sea-level rise, approximately 6000 cal years BP there was a rapid drop in sea level followed by a 2 m rise approximately 4000 cal years BP. Thereafter, the sea level dropped once again just below the present level between 3500 cal years BP and 2800 cal years BP. Thereafter sea level rose, reaching a peak perhaps approximately 1800 cal years BP before falling again (Miller *et al.*, 1995).

South African researchers tend to have relied on ‘global records’ as a comparison in dating shorelines, which are compared with particular elevations (Ramsay and Cooper, 2002). The Holocene sea-level records for the eastern and western coastlines, show partial but not a complete overlap at the current resolution (Miller *et al.*, 1995), for this reason there is a need for higher resolution studies, as a result the exact changes that have taken place along the coastline may become more apparent. There is a need to date more precise sea-level indicators to construct sea-level curves in South Africa (Compton, 2001).

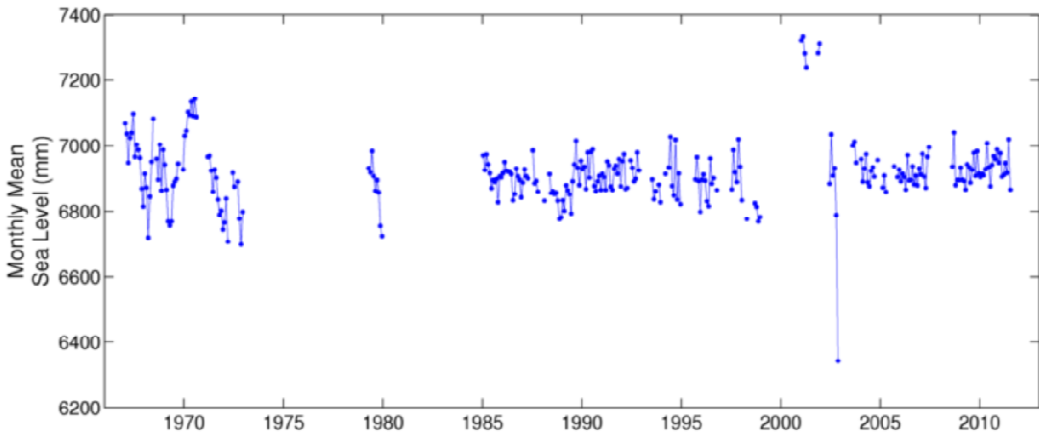
3.4 South African instrumental record

Instrumental evidence of recent sea-level change was confined to tide gauge data before 1993, when satellite altimetry was introduced (Mather *et al.*, 2009). In many ways satellite data has improved the geographical coverage of sea-level records, however accurate tidal gauge data are still needed to calibrate the satellite altimeter results (Mitchum, 1998; Mather *et al.*, 2009). Unfortunately, not many of the tide gauge records for the southern hemisphere extend further back than 50 years. In South Africa there has been little research on the rate of sea-level change calculated from tide gauge data (Mather, 2007). Brundrit (1984) focused on sea-level changes using tidal gauge information along the west coast of South Africa and since then very little research has been conducted (Mather *et al.*, 2009).

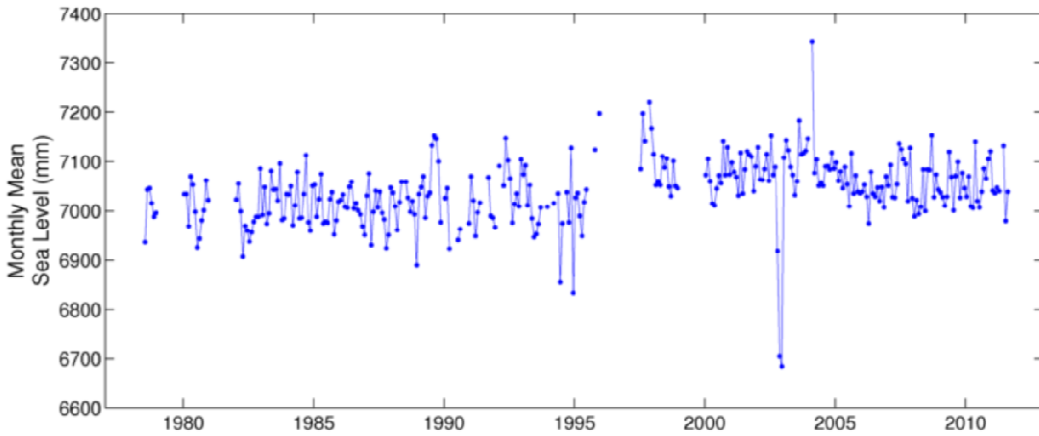
The South African coastline can be divided into three regions, east, south and west. The southern region is subject to variable temperatures due to the mixing of the Benguela and Agulhas Currents (Bindoff *et al.*, 2007; Mather *et al.*, 2009). Tidal gauge data indicates that Simon’s Town and Knysna’s eustatic sea-level trends were +1.97 mm year and +2.00 mm year, taking into consideration vertical crustal movements. Sea-level change appears to be greater in the southern region where it is warmer offshore compared to the cooler west part of the southern region (Mather *et al.*, 2009).



Durban



East London



Port Elizabeth

**Figure 3.2: Tide gauge time series for three stations along the east coast (PSMSL, 2012).
Note: Monthly mean sea level axes vary in scale.**

The western region is dominated by the cold upwelling waters of the Benguela, which are trapped by warm waters from the north and south. An average sea-level trend for this region was +0.80 mm year, once all corrections were applied and vertical crustal movements introduced. This is found to be in conjunction with global sea-level change trends in the IPCC assessment (Mather *et al.*, 2009). The eastern region is affected by the Agulhas Current, with warm water near the top and therefore exposed to rising air temperatures in the equatorial zone. There are four tide gauge stations in this region, however only two were used by Mather *et al.* (2009) to calculate the average sea-level rate of change as for both Richards Bay and East London (Figure 3.2) there was insufficient data. The average rate of sea-level change for this region was +3.03 mm year; this was the highest estimated rate of change for the South African coastline (Mather, 2007; Mather *et al.*, 2009). This figure is slightly higher than the global average of +3.0 mm year, though is expected because of the warmer water being fed down from the equator (Bindoff *et al.*, 2007; Mather *et al.*, 2009).

From tidal gauge calculations along the South African coastline over the past 50 years sea-level rise has been relatively constant. However, there are variations in sea-level changes depending on the location. These differences are influenced by the interactions of the Agulhas and Benguela currents, water temperature and vertical crustal movements (Mather *et al.*, 2009). The South African coastline and shelf comprises two trailing edge margins and has been tectonically stable since the late Quaternary (Miller *et al.*, 1995; Ramsay and Cooper, 2002).

3.5 Conclusion

During the LGM, thick ice sheets extended across the high latitudes and sea levels during this period were at their lowest. It was during this period according to sea-level records as the most recent period in the earth's history when ice-sheets globally were at their maximum. The Holocene has a much milder climate compared to that of the former glacial period (Lambeck *et al.*, 2002). The rate of sea-level rise has increased during the 20th century (Leorri *et al.*, 2010). However, the analyses of instrumental sea-level observations are unable to resolve the timing of sea-level acceleration, or variation (Woodworth, 1999; Leorri *et al.*, 2010). Some studies acknowledge acceleration as having occurred round the end of the 19th century and approximately 1930, whereas others have suggested a deceleration of the sea level within the 20th century (Woodworth, 1999; Gehrels *et al.*, 2008). Possibly, instrumental records showing rates of sea-level change are too short and therefore unable to provide adequate spatial coverage to provide a definite answer for the timing of sea-level acceleration (Gehrels *et al.*, 2008).

CHAPTER FOUR: STUDY SITE

This chapter introduces the study site, consisting of a physical description and historical account. It provides background to estuaries and intertidal environments in South Africa, pertinent to the selection of an intertidal environment as a study site for this research. It will then more specifically review estuarine systems and salt marshes in South Africa and finally discuss the chosen salt marsh for this research.

4.1 Estuarine systems

Day (1980: 3) defines estuaries as “that portion on the earth’s coastal zone where there is interaction of the ocean water, fresh water, land and atmosphere.” An estuary is a partially enclosed coastal body of water, which has a direct link with the ocean, and the water within the estuary is diluted by fresh water that is the result of runoff from the surrounding land. Therefore, if one were to look at an estuarine system, one would see that an estuary is a multifaceted, dynamic and biologically rich environment, which is subject to physical forces (Day, 1980).

Estuarine environments are considered to be one of the most productive types of ecosystems found and thus are a rich food source for both commercially and recreational marine fish species that use estuarine systems as fish nurseries (Chuwen *et al.*, 2009). Estuaries are usually open to the sea and are classified according to the salinity and tide action. Plant species within an estuarine system show varying degrees of tolerance to saline conditions and therefore will be found in different parts of the estuary (Paterson, 1998). The dominant feature of an estuary is its intertidal channel, it is at this point at which the water is interchanging between the river and the sea, as the tide either retreats or advances. This main channel will eventually lead into a tidal creek and then continue to meander into intertidal mudflats marshes (Lubke, 1988).

4.1.1 Intertidal zonations

Intertidal zonation refers to specific conditions found between the low and high tide watermarks along the shoreline (Bennett, 1987; Peterson, 1991), whereby plants and animals form visible communities in this particular zone (Peterson, 1991; Gehrels, 1994; Lubke and de Moor, 1998; Gehrels, 2002). The plants and animals living in this zone need to be able to withstand different physiological stresses such as exposure to air, flooding during high tide, and change in temperatures as well as exposure to fresh water in the form of rain (Bennett, 1987; Peterson, 1991). The greater the elevation of the shoreline the greater the risk there is for plants and animals living there to starve (Peterson, 1991), as many of them will only feed when they are submerged in water (Bennett, 1987). Water coming in with the high tide provides these marine animals and plants with the nutrients they require (Gehrels, 2002). In addition, when the plants and animals living in this

zone are covered by seawater they are able to function fully (Lubke and de Moor, 1998). The concept behind intertidal zonation has a major part to play in intertidal foraminifera being appropriate indicators of past sea-level change (Gehrels, 2002).

Few marine plants and animals are able to withstand dryness resulting from the tide going out in the higher elevations of the intertidal zone (Lubke and de Moor, 1998). Competition bands of different species are very obvious across the intertidal zone (Gehrels, 2002). These bands are a reflection of both the plants and animals ability to adapt to air exposure and the wave action striking the shore (Lubke and de Moor, 1998). Conditions closer to the water will be less severe, thus there will be more competition in the lower zones and more species will appear (Lubke and de Moor, 1998). The best examples of these species bands found in intertidal zones are seaweeds found on rocky shores and salt-marsh vegetation zones (Lubke and de Moor, 1998; Gehrels, 2002). The vertical zonation patterns of plants may not be the result of just tidal inundation, but the fact that plants show varying degrees of tolerance to factors such as salinity and water currents. The consequences of plants tolerance towards salinity and water conditions will be vary biomass in different zones of the salt marsh (Lubke and de Moor, 1998).

4.1.2 Description of salt marshes

Salt marshes have considerable differences in temperature, salinity and pH across such an environment (Scott *et al.*, 2001). Situated in the higher part of the intertidal zone (Gehrels *et al.*, 2008), they are distinguished from their neighboring tidal mud flats by the halophytic vegetation covering the surface (Horton and Edwards, 2006) and are frequently flooded by the ocean (Allen, 2000). A salt marsh is an example of a true ecotone; containing organisms which have adapted to an environment that has both aquatic and terrestrial characteristics (Cattrijsse and Hampel, 2006). Cattrijsse and Hampel (2006: 293) define a salt marsh as an area “with alluvial sediments deposited on the shore by the sea and subjected to tidal or weather-effected inundation by more or less diluted seawater”. Salt marshes occur on intertidal shores in the mid to high latitude regions globally and often in the mid to lower reaches of an estuary (Gehrels *et al.*, 2008; Kennish, 2001). Salt-marsh sediments vary from being continuously waterlogged to moist for a short period of time and dominated by halophytes plant species (Cattrijsse and Hampel, 2006; Booth, 2007).

Vegetated surfaces of salt marshes in Europe will only be flooded during the highest tides, whereas in North America the salt marshes are generally flooded at every high tide (Cattrijsse and Hampel, 2006). The extent to which a salt marsh is subject to flooding will determine the distribution of vegetation and this in return the use of the salt marsh by both fish and invertebrates (Booth, 2007; Cattrijsse and Hampel, 2006). The duration and height of the tides will differ based on local hydrological conditions, tidal cycles and existing patterns (Odum, 1988; Adam, 2002).

Vegetation found in a salt marsh has specific physiological adaptations giving the vegetation the ability to survive in varying saline conditions (Booth, 2007). Saline conditions of a salt marsh usually range between 18 and 35psu, however, pools that are present during the intertidal period can reach hypersaline levels due to evaporation (Odum, 1988). Frequently flooded areas will have different vegetation compositions compared to those areas that are only flooded during the exceptionally high spring and neap tides (Odum, 1988). However, the extreme low regions of salt marshes may not have any vascular plants present, may have microphytobenthic and macroalgae, which is a vital food source for various fish species (Odum, 1988; Booth, 2007).

Estuarine fauna can be subdivided into resident and non-resident species. Resident species will usually have a greater tolerance to fluctuations in saline and water levels (Edwards and Horton, 2006; Booth, 2007). Resident macro-faunal species of a salt marsh will usually consist of crabs, isopods, amphipods and marine or freshwater molluscs (Odum, 1988). Whilst non-resident species will comprise predominantly of zooplankton, which can consist of either juvenile or larval forms of invertebrates but non-resident species can also include adult fish and water birds who feed off of the salt-marsh vegetation or the invertebrates (Booth, 2007). Within the estuarine community, salt marshes have been documented as being important habitats for both fish and invertebrates (Shenker and Dean, 1979).

4.1.3 Marsh surface elevation

Horton and Edwards (2006: 5) explain that “vertical changes in relative sea-level change are expressed in terms of their altitude relative to a geodetic datum,” in South Africa the official geodetic datum is the Hartebeesthoek94 Datum (Department of Rural Development and Land Reform, 2010). The changes that occur with relation to marsh surface altitude are not the same as the changes in marsh surface elevation (Horton and Edwards, 2006).

A typical salt marsh will grow vertically because of gradual accumulation of stone, sand, silt and clay sediments from marine environments and the accumulation of sediment derived from organic matter such as decomposing salt-marsh plants above the ground (Allen, 2000; Horton and Edwards, 2006). The contribution of these sediment sources, which are, known as minerogenic and organogenic sediments, are controlled by the extent, frequency and duration of a raising body of water because of tidal fluctuations, which ultimately is determined by the marsh surface elevation (Horton and Edwards, 2006). Subsequently, there is a feedback loop that exists, in which the elevation of a marsh controls and is controlled by tidal inundation characteristics (Allen, 2000). Tidal range, relative sea level and minerogenic sediment supplies all contribute towards a dynamic equilibrium being reached (Edwards and Horton, 2006).

Salt marshes in northeastern North American and the British Isles have provided scientists with a large amount of information pertaining to Holocene sea levels (Scott and Medioli, 1978). These marshes develop

very differently from one another; the difference is usually organogenic versus minerogenic sedimentation. Those salt marshes in which growth is dominated by organogenic sedimentation would have had a very high preservation with approximately 70% of the plant material that was buried centuries before being identifiable. Therefore, those marsh surfaces that were accumulating at a rapid rate were able to keep up with the rising relative sea level that took place during the late Holocene. If one were to take a core from such a marsh, meters of marsh peat that has accumulated over the last 3000-4000 years would be reflected in the stratigraphy (Gehrels, 2000).

Furthermore, one will find salt marshes that are inorganic especially during the late Holocene and minerogenic sediments are the marshes governing growth controls. The sediment will accumulate most rapidly in areas with a low elevation and decrease as the marsh surface elevation increases and the tidal inundation decreases (Horton and Edwards, 2006). The differences in the marsh characteristics have an impact on modern intertidal foraminiferal behaviour which has implications for reconstructing relative sea-level changes (Scott and Medioli, 1978). Not only is foraminiferal behaviour impacted upon but their abundance can also be influenced by the chemical and physical interaction of salt-marsh vegetation with environmental variables (Franceschini *et al.*, 2005).

4.2 South African estuarine systems

Estuarine systems are situated along the South African coastline, containing a great diversity of both fauna and flora habitats (Lubke and de Moor, 1998), which are often good indicators of the transition between marine and freshwater (Lubke, 1988). The majority of South African estuaries were created as a result of river valleys becoming immersed by seawater (Lubke, 1988; Lubke and de Moor, 1998). In the last 15,000 years, sea level along the South African coastline has risen by 130 m (Lubke and de Moor, 1998).

The shape of an estuary is determined by the amount of rainfall and the size of the catchment area. Whitfield and Kok (1992) suggest that there are five types of estuarine systems of which all are found along the southern and eastern coastline of South Africa, the two most prominent being permanently open or temporarily open or closed (Bornman and Adams, 2008). The state of the mouth is determined by large sand bars, which in return are controlled by the flow of the river (Bornman and Adams, 2008). According to Reddering and Rust (1990) there are only 37 true estuarine systems found in South Africa that continue to have a permanently open inlet to the ocean out of the 289 river mouths along the South African coastline.

4.2.1 Salt marshes

Salt marshes found in South Africa are situated either in estuaries or within a protected embayment along the southern, eastern and western coastlines (Paterson, 1998; Franceschini *et al.*, 2005). Salinity and tidal characteristics are determined by the interaction that occurs between the ocean and the river (Franceschini *et*

al., 2005). Seventy-five percent of South African's salt marshes occur in the Langebaan, Swartkops, Knysna, Berg and Olifants estuarine systems (Paterson, 1998). By looking at the high species diversity of vegetation, one could conclude that South African salt marshes are similar to European salt marshes and are infrequently flooded (Paterson, 1998). Although macrophytes grow in abundance on South African salt marshes, they show very little species diversity as a plant community (Lubke, 1988; Franceschini *et al.*, 2005). Macrophytes distribution within the salt marsh will be directly related to salt tolerance and elevation (Franceschini *et al.*, 2005).

Vegetation most commonly found in South Africa's high salt marshes is *Sarcocornia perennis* (Taylor and Allanson, 1995) and *Limonium linifolium* (sea lavender). These zones are highly productive due to the nutrient rich soil found in a salt marsh however, there are very few animals that feed off these plants and therefore the organic matter from these plants enter the food chain as litter (Lubke, 1988). The low marsh areas and mudflats are characterized by *Spartina maritima* (cordgrass) and *Zostera capensis* (eelgrass) both of which cover large areas and are abundant (Adam, 2002). *Zostera capensis* is susceptible to environmental changes and in many salt marshes in South Africa has disappeared such as the Swartkops River (Lubke, 1988). Microphytobenthic and macroalgae species are found on the salt marshes bare mud (Paterson, 1998).

The majority of salt marshes in South Africa have very distinct vegetation zones, this is apparent as one moves from the intertidal channel onto higher ground, with the *Spartina maritima* zone approximately the low water tidal mark to the high water mark (Lubke and de Moor, 1998). *Sarcocornia perennis* or *Chenolea diffusa* will stretch from the high tide mark to the spring tide mark (Allanson and Baird, 1999). However, in the majority of South African salt marshes *Sporobolus virginicus* is the most dominant and is found on higher ground (Lubke, 1988).

The zonations that are evident are a reflection of the succession within the salt marsh. *Zostera capensis* will be the first plant species to inhabit the mudflats, once salt has accumulated and the water has become shallower so *Spartina maritima* will begin to invade the area, thus stabilizing the mudflats (Lubke, 1988). According to Lubke (1988: 343) it is a long process for mesophytes or hydrophytes to colonize an area rather than halophytes. For hydrophytes to become established, the salt that has accumulated in the soil of the salt marsh will need to be washed out either by heavy rainfall or by freshwater flooding the area (Lubke and de Moor, 1998).

4.3 Site description

The South African coastline is affected by two currents, the Agulhas and the Benguela (Ramsay and Cooper, 2002). The Agulhas current, which flows along the eastern coastline of Africa, is narrow, swift and strong. It is a western boundary current and like many western boundary currents it is extremely fast, and is the largest

of the western boundary currents in the world (Gordon, 1985), whereas the Benguela current flows northward up the west coast of Africa. Unlike the Agulhas current that is narrow, the Benguela is a broad current, approximately 200-300km wide (Wedepohl *et al.*, 2000).

Found adjacent to the town Kenton-on-sea in the Eastern Cape is the Kariega River mouth. The Kariega Estuary is a permanently open system (Paterson *et al.*, 2008), situated on the east coast of South Africa (33°41'S, 26°42'E) and stretches approximately 18 km inland (Paterson and Whitfield, 2000a,b; Bate *et al.*, 2004). The channel varies between 40 m wide in the upper reaches and 100 m in the lower reaches of the estuary (Bate *et al.*, 2004) and is surrounded by salt marshes, sand flats (Grange, 1992), and steep slopes up the reach of the estuary. Kariega River is highly sinuous towards the mouth (Figure 4.1) resulting in complex estuarine sediment transport processes. The steep slopes that run up part of the length of the estuary are covered in valley-bushveld vegetation (Froneman, 2001).

The Kariega Estuary is a marine-dominated hypersaline system with very little riverine influences (Grange *et al.*, 2000). Restricted freshwater input into the system is largely due to the fact that the Eastern Cape is relatively arid which, is made worse by the low rainfall to runoff conversion (Bate *et al.*, 2004; Paterson *et al.*, 2008). In addition, the Kariega Estuary catchment is small and is extremely regulated by three dams and various weirs in the area. Even though there is no strong freshwater input, there is still sufficient scouring by tidal currents to maintain a permanent connection with the sea (Bate *et al.*, 2004). This system, like many other systems found along the Eastern Cape shoreline, has a small tidal prism. Water levels in this system change due to the semi-diurnal and spring neap tidal cycles (Taylor, 1987). Kariega Estuary is said to have been formed as a result of drowned river valley due to sea-level rise, but has since been impacted by man. Throughout colonial times the river mouth has remained open to the sea. Thus it can be classified as a ria-type estuary according to Reddering and Rust (1990). Such sea-level rise that may have resulted in the river valley being submerged may have either been eustatic or isostatic, resulting in a large estuary being formed at the mouth of what was a comparatively insignificant river (Reddering and Rust, 1990).

There are three main intertidal salt marsh systems at Kariega, namely Taylor's, Grant's and Galpin's marshes. Grant's salt marsh is dominated by *Spartina maritima* and *Sarcocoria perennis*, while Taylor's salt marsh is dominated by *Sarcocornia perennis* and has a single unvegetated intertidal creek and one opening into the estuary. Both Taylor's and Grant's creeks are shallow and narrow (Paterson and Whitfield, 2000a; 2000b). Galpin's is unlike the other two salt marshes in that it is attached to a small estuarine embayment rather than the main channel, is dominated by *Spartina maritima* and *Sarcocornia perennis*, and this site does exhibit clear vegetation zones. There are five main high marsh pools and the intertidal creek is wider and deeper than the previous two (Whitfield and Harrison, 2003), thus it was clear that this was a potential research site.

Kariega Estuary is a unique system with very little freshwater input and therefore the salinity of Kariega Estuary is extremely high. Salt-marsh plants are found along the length of the estuary system. The salt-marsh vegetation found in Galpin's salt marsh indicates noticeable zonations as one proceeds from the intertidal channel towards the terrestrial edge.

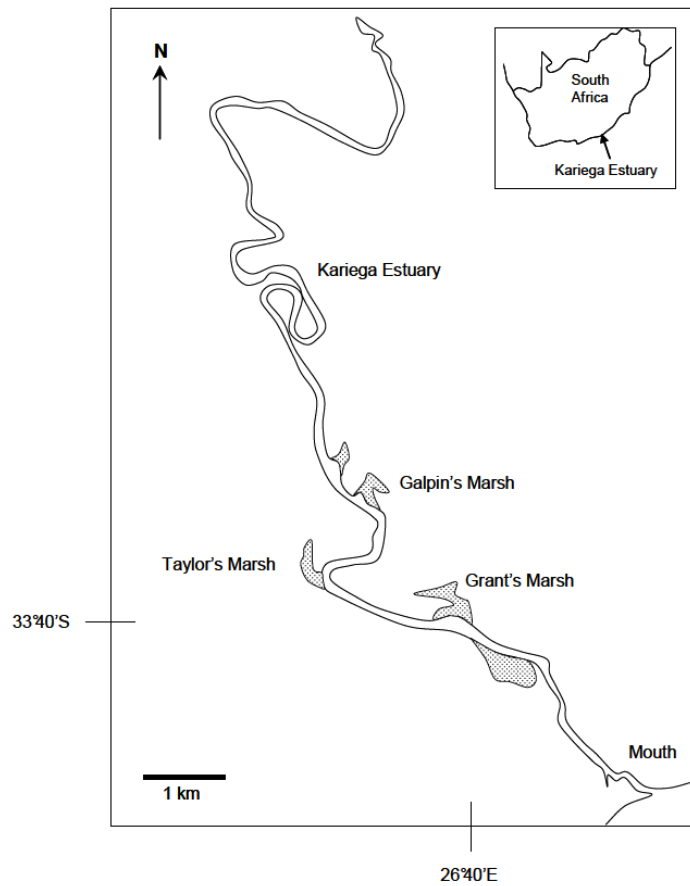


Figure 4.1: Map of Kariega Estuary indicating the position of Galpin's salt marsh (Paterson and Whitfield, 2000a)

CHAPTER FIVE: METHODS

This chapter describes the research design, and data collection techniques both in the field and laboratory. Field work consisted of surface sampling, core extraction and a stratigraphic transect. Laboratory work included sub-sampling of sediment cores, extraction of foraminifera and radiocarbon analysis.

5.1 Field component

The fieldwork component of this research was conducted in April 2011 at Galpin's salt marsh, Kariega Estuary, South Africa. The purpose of the fieldwork was to map vegetation zonation, collect surface sediment samples that would be analysed for modern foraminifera assemblages, establish a detailed stratigraphy of the salt marsh, and to collect fossil sediment cores for palaeoenvironmental analysis and for establishing a suitable chronology.

5.1.1 Vegetation mapping

Vegetation cover was identified (Lubke and de Moor, 1998) and abundance recorded using a Braun-Blanquet scale (Poore, 1955) along each of the surface transects. Distribution, manifested as zonation and species abundance were plotted along each transects.

5.1.2 Surface samples

Three sampling transects were positioned along an elevation gradient from high marsh through to the low marsh and tidal mud flats to the low tide mark (Figure 5.1). Tidal pools were not sampled for foraminifera. However, to determine the vertical zonation of foraminifera for the training set, Transect One and three were joined together. This was possible as Transect One ran from the terrestrial edge to the high water mark, while Transect Three ran from the high water mark to the low water mark. For the purpose of this research transect One and Three will be identified as Transect Four. Thirty-nine surface foraminiferal samples (Plate 5.2) were collected using a hand-held surface sampling device (Plate 5.1) at vertical intervals of 5 cm along each transect. The surface samples can be root bound and difficult to extract, therefore the hand-held sampling device has a sharpened edge. The sampling device is turned as it goes into the ground to avoid compaction.

Samples elevation is determined by leveling to a common benchmark. This benchmark was determined by hammering a ground peg into an area of stability (above highest astronomical tide level). The top of the peg was located accurately in terms of X, Y and Z coordinates and all surveyed data was leveled back to this height. Field measurements of elevations of all samples were surveyed relative to the South African TrigNet

reference station at Grahamstown, heights above ellipsoid and then linked to local tidal datums (Table 5.1) (SA Navy, 2011).

It is important to take into consideration the effects of environmental factors on the distribution of foraminifera such as the following environmental; air and water temperature, light intensity, tidal elevations, oxygen concentrations and pH. However, for the purpose of this thesis only sample elevations were determined.

Table 5.1: Tide levels at Port Elizabeth relative to chart datum (CD) (SA Navy, 2011)

Tide Level	Height (m CD)
Lowest Astronomical Tide	0
Mean low Water of Springs	0.21
Mean low water of Neaps	0.79
Mean Level	1.04
Mean High Water of Neaps	1.29
Mean High Water of Springs	1.86
Highest Astronomical Tide	2.12



Plate 5.1: Surface sampling device

5.1.3 Stratigraphic description

An extensive program of coring using a 20 mm Ø gauge auger (Plate 5.4) was conducted to establish the stratigraphy of the salt marsh along the transect (Plate 5.3). The stratigraphy was classified using the Troels-Smith system of sediment description and portrayal (Troels-Smith, 1955; Kershaw and Bulman, 1996), which is suited to unconsolidated organic sediments. Classification was performed for each of the seven

transect cores (See appendix A), which was accompanied by Munsell colour description (Plate 5.4). This process was used to determine where along the transect the deepest salt-marsh peat occurred. Cores were taken at the start and ends points along the transect and at regular intervals of 3 or 5 m.



Plate 5.2: Surface sample extracted



Plate 5.3: Stratigraphy points along Transect Two, Galpin's salt marsh, Kariega Estuary

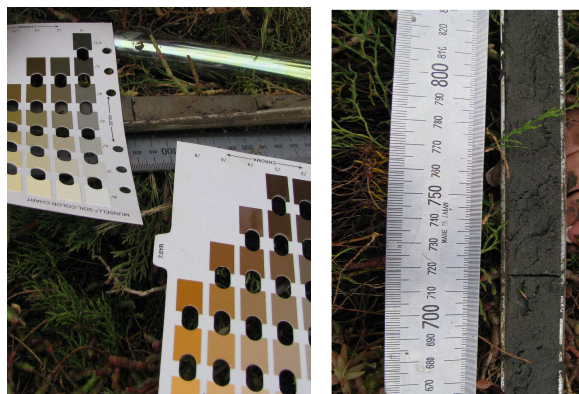


Plate 5.4: Munsell colour description charts and 20 mm Ø gauge auger

5.1.4 Core extraction

Five one meter continuous cores were extracted along Transect Two in the vicinity of the deepest salt-marsh peat profile (33°39'04"S 26°39'74"E). A 50 mm Ø gauge auger was used for core recovery. The extraction of cores was carried out carefully so as not to disturb the profile or contaminate samples. Sediment cores recovered from the high marsh zone were then placed into labeled PVC piping, wrapped in cling wrap and heavy duty plastic and transported back to the laboratory where they were refrigerated for later analysis.

5.2 Laboratory Component

Analysis in the University of KwaZulu-Natal, Pietermaritzburg laboratories took place between April 2011 and August 2012. Modern and fossil sediments were subsampled for foraminiferal assemblages. Bulk sediment samples from the fossil sediments were also used to provide material for dating.

5.2.1 Subsampling

A subsample of four cm³ was extracted from the top five cm of each surface sample, preserved in a solution of Rose Bengal and ethanol and buffered by sodium bicarbonate. Samples were stored in the refrigerator at a temperature of 2 - 5°C and shaken regularly to ensure that the living foraminifera took up the Rose Bengal stain. The samples were left to soak for a minimum 36 hours (Scott and Medioli, 1980; Gehrels, 2002).

Two of the five cores were used for further analysis, Core Four was subsampled at selected intervals for AMS ¹⁴C and the other sampled along its entirety for foraminifera (Table 5.2). The description of each sediment core (Appendix B and Appendix C) was determined prior to sub-sampling using Troels-Smith Sediment Classification Scheme (Troels-Smith, 1955; Kershaw and Bulman, 1996) and Munsell Colour Description (Figure: 6.10).

Table 5.2: Table of cores

Core Name	Depth (cm)	Use
Core 1	66	Archiving
Core 2	94	Foraminiferal analysis
Core 3	40	Archiving
Core 4	92	AMS dating
Core 5	94	Archiving

Transects Galpins Salt Marsh, Kariega

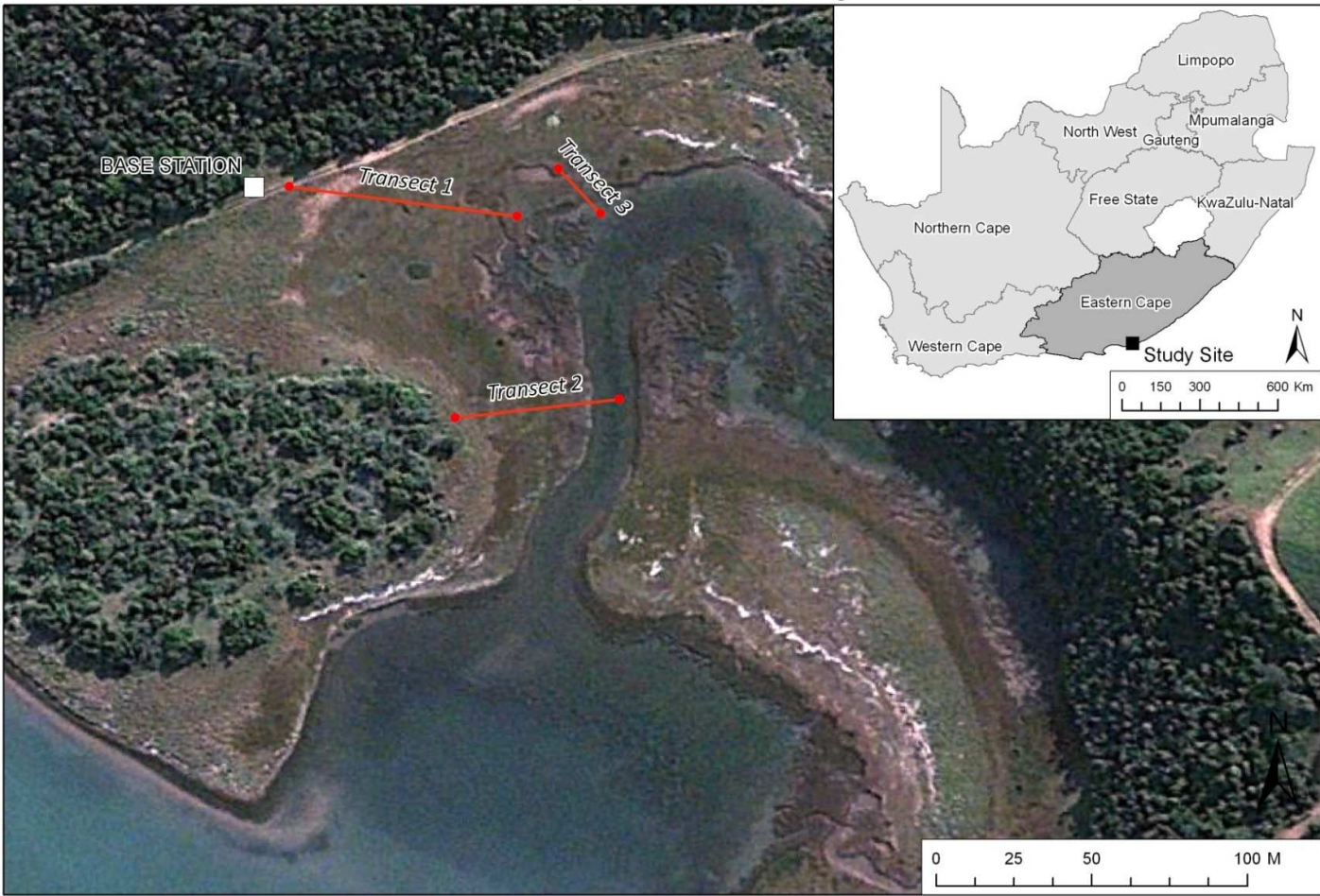


Figure 5.1: Position of transects on Galpin's salt marsh, Kariega

A total of 188 sub-samples were extracted at a 0.5 cm resolution from the core and stored. Four cm³ of each sub-sample are processed and stored in buffered ethanol in the refrigerator. The remainder of each sub-sample was placed into a polypropylene sealable bag, labeled and refrigerated. Once the sedimentation rate was determined, it was decided that 48 samples at a 2 cm resolution were used for microscopic foraminiferal analysis.

Both surface and core samples were individually washed through 500 and 63 µm mesh sieves. The sample collected in the 63 µm sieve was washed into a beaker. Once the sediment had settled the suspended organic material was removed. The sample was then suspended in 1.5 liters of water and sub-divided into eight aliquots using a wet splitter (Gehrels, 2002).

5.2.2 Microscopy

The samples were kept wet throughout the counting process thus preventing aggregation of organic particles and minimizes degradation. A Leica M205C stereomicroscope with an attached camera was used for counting and identifying foraminiferal assemblages under 60X, 80X and 100X magnification. For each sample, one aliquot was placed into a spiraled grooved foraminiferal counting tray and counted, if the subsample did not provide sufficient specimens another aliquot was counted; 250-300 foraminifera per sample were counted, identified and photographed (Appendix E). From a statistical perspective a count of 250 individuals is necessary especially if significant indicator species are present but in low numbers (Gehrels, 2002). Those that were not identified immediately under the microscope were photographed and identified at a later stage. Samples extracted from the upper reaches of the salt marsh and samples at certain depths in the core had low concentrations of foraminifera and the desired number of 250 was not obtained. In the surface samples, both stained and unstained foraminifera were counted and identified (Appendix F-I). Both living and dead assemblages were counted, as that total assemblages provide a better indication of the environmental conditions as both season and temporal fluctuations are integrated (Scott *et al.*, 2001; Horton *et al.*, 2005b).

Specimens that were brightly or partially stained were considered alive and those that were not stained were considered non-living at the time of collection (Berkeley *et al.*, 2008). Several references were used to aid in taxonomic identifications of foraminifera found along each transect and down the core (Murray, 1979; Horton and Edwards, 2006). A taxonomic list of each species of foraminifera encountered and its description was compiled (Appendix D). Tests that were either fragmented, including linings, too small or could not be identified were designated unknown.

5.3 Data Analysis

Modern assemblages were used to construct models of environmental information and much of the data were subject to statistical examination throughout the data analysis stage. Software packages used in this analysis include the Classical Age Modeling source code (Blaauw, 2010), Psimpoll (Bennett, 2005), Constrained Incremental Sum of Squares (Grimm, 1987), CANOCO computer program (Ter Braak, 1995) and the software package C₂ (Juggins, 2003).

5.3.1 Radiocarbon dating and chronology

Five subsamples were extracted at selected intervals of 18 cm, 66 cm, 53 cm, 89 cm and 92 cm for AMS radiocarbon analysis, based on stratigraphic boundaries. All large rootlets and pieces of debris were removed. Samples were then placed in an 80°C oven overnight. Once each sample had cooled, each was weighed and placed into a polypropylene bag. All samples were sent to BETA Analytic inc. in Miami for radiocarbon analysis. The organic fraction was dated for four bulk sediment samples and carbonate fraction for the 92 cm sample as it consisted of shell fragments.

An age-depth model was established for the sediment core to place the sea-level record within a chronological framework. To create a suitable age-depth model using the dates obtained from the AMS radiocarbon dating, the dates needed to be calibrated. Calibration and the development of the model were achieved using CLAM (Classical Age Modeling) source code (Blaauw, 2010), in conjunction with R open-source statistical environment (Team, 2010). Standard procedures were followed as described by Blaauw (2010). All the radiocarbon dates were calibrated using ShCal 04 calibration (McCormac *et al.*, 2004) and linear interpolation model.

5.3.2 Data zonation and diagrams

The results from the foraminifera counts along the transect and down the core were plotted using Psimpoll Version 4.263 (Bennett, 2005). Foraminifera counts from each transect were plotted against the length of the relevant transect along with the stratigraphic information and calibrated dates as established using an age-depth model. To assist in the interpretation of the fossil foraminiferal assemblage data down the core, it was divided into five zones using Constrained Incremental Sum of Squares (CONISS) (Grimm, 1987). CONISS is a stratigraphically constrained ordination technique devised for analysis of sediment core data. Foraminiferal zonation using CONISS was run on this dataset to ensure that zonations would not be influenced by personal observations.

5.3.3 Transfer function

Multivariate regression analyses were used to quantify the relationship between surface assemblage data and mean sea level in the contemporary environment. This produced a transfer function that could be applied to down-core assemblage data allowing historical sea levels to be inferred. For the purpose of this research both living and dead (total) surface assemblages were used to construct a training set. This does not take into consideration how live assemblages are affected following death (Murray and Alve, 2000; Horton *et al.*, 2005b). Some researches argue that dead assemblages are suitable for a training set, especially in temperate salt marshes as they are less vulnerable to seasonal variations as even though live assemblages are variable they do not move from place to place in subsurface environments (Horton, 1999; Scott *et al.*, 2001; Horton *et al.*, 2005b)

From the data set of living and dead surface foraminifera assemblages, a transfer function was constructed using software C_2 (Juggins, 2003), thus quantifying the relationship between elevation and foraminiferal assemblages. Developing a transfer function to reconstruct past sea levels from fossil foraminifera records is an interactive process. The data set was a combination of all data from all three transects of total assemblages. The relationship between total foraminiferal assemblages and elevation above sea level was determined using a multivariate statistical analysis. This was done by performing a Detrended Canonical Correspondence Analysis (DCCA) using CANOCO (Ter Braak, 1995). Running a DCCA provides one with an estimate of the gradient length of standard deviation (SD) units, which helps to determine the most suitable model for the total assemblage training set (Birks, 1995). The gradient lengths for this research's data set were greater than 2 SD units (Appendix E) and therefore the most appropriate regression and calibration technique is a unimodal response model. C_2 (Juggins, 2003) was used to calculate regression statistics. As recommended for species data with gradients $>2SD$ units, unimodal regression model was used, and in this case weighted averaging (WA).

CHAPTER SIX: RESULTS

This chapter describes the results of the vegetation and foraminiferal analysis, stratigraphic descriptions and radiocarbon dating. Radiocarbon dating is evaluated and calibrated providing a chronology and used in conjunction with the transfer function data to construct a past sea level curve. Vegetation zonation is compared with the compositional changes in the surface foraminiferal assemblages and elevation determining the accuracy of the foraminiferal assemblages as indicators.

6.1 Vegetation zonation

Galpin's salt marsh exhibits upper and low marsh zonations based on the vascular flora present along each of the elevational transects across the salt marsh (Figure 6.1). Approximately 70% of the marsh area can be described as high marsh, with this zone being denser in vegetation compared to both the upper and lower sections of the marsh. The marsh is dominated by *Sarcocornia perennis* (Plate 6.1), which is present in all three zones. The higher marsh zone is dominated by *Chenolea diffusa* (Plate 6.1) and *Limonium linifolium* (Plate 6.1) with some occurrence of *Triglochin straita* (Figure 6.2). The lower marsh or tidal flats are covered by *Spartina maritima* and *Zostera capensis*, a pattern which is clearly evident in Transect One (Figure 6.2). The dominant plants in the upper part of the intertidal zone closer to the terrestrial edge are *Carpobrotus edulis* (Plate 6.1) and *Sporobolus virginicus* (Plate 6.1).

6.2 Training set analysis

A total of 13 foraminifera taxa were identified along the transects forming the training set for Galpin's salt marsh. Eight of the taxa occurred in both transects, with the dominant species being agglutinated. These eight taxa in decreasing order of abundance (living and dead) are; *Jadammina macrescens*, *Miliammina fusca*, *Trochammina inflata*, *Quinqueloculina spp*, *Scherochorella moniliforme*, *Fissurina spp*, *Trochammina ochracea* and *Spirillina vivipara*. However, there were five taxa that were exclusive to Transect Four, viz., *Ammonia batavus*, *Ammonia aberdoveyensis*, *Haynesina germanica*, *Textularia earlandi* and *Textularia gramen*. The majority of tests were well preserved especially those that were close to the intertidal zone. Preservation declined ever so slightly towards the terrestrial edge, whereby small and more fragile foraminiferal tests showed signs of dissolution.

The relative abundance of foraminiferal species increases towards the intertidal channel as elevation decreases compared to the abundance of species near the terrestrial edge where elevation is greater (Figure 6.3). A higher percentage of *Jadammina macrescens* are present at higher elevations, while it rapidly

declines as elevation decreases. In the high marsh zone closer to the terrestrial edge there is a small number of *Trochammina inflata* present. *Trochammina inflata* and *Miliammina fusca* predominantly occur in conjunction with one another. However, closer to the intertidal channel *Trochammina inflata* decreases in abundance and *Miliammina fusca* increases. At the lower elevations where the tidal mudflats would be

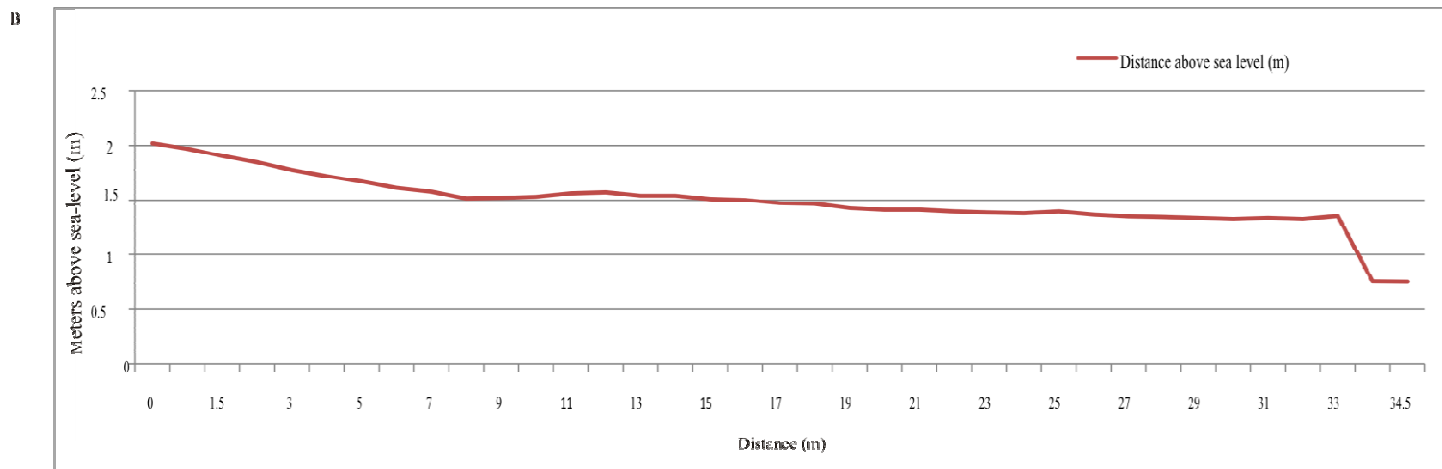
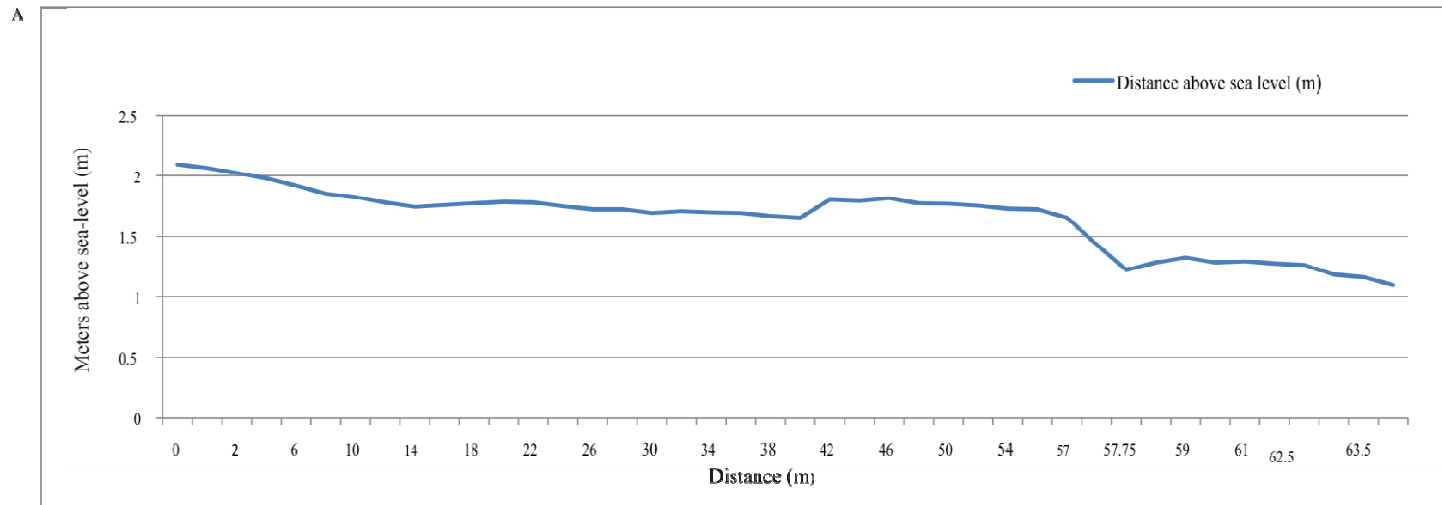


Figure 6.1: Elevation profiles of (A) Transect Four and (B) Transect Two, Galpin's salt marsh

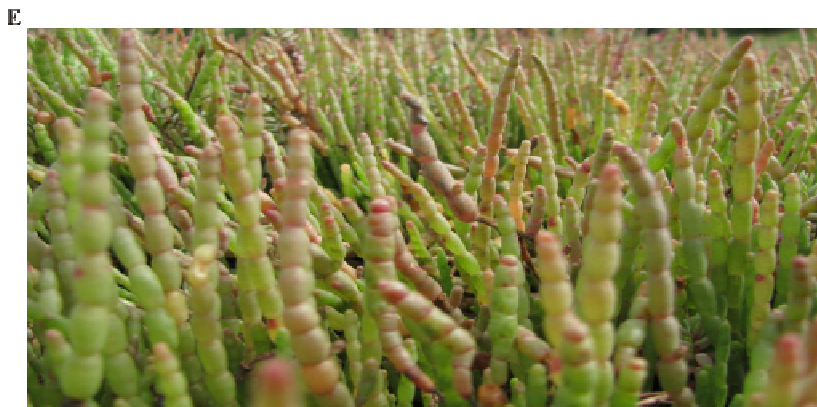
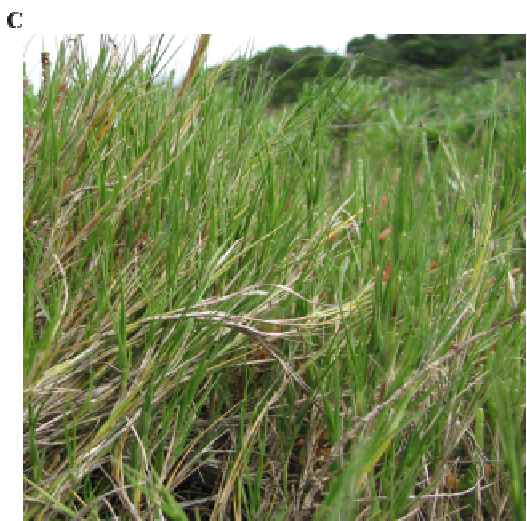
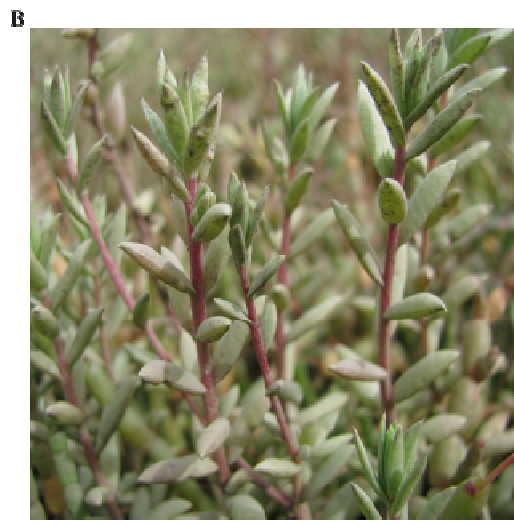
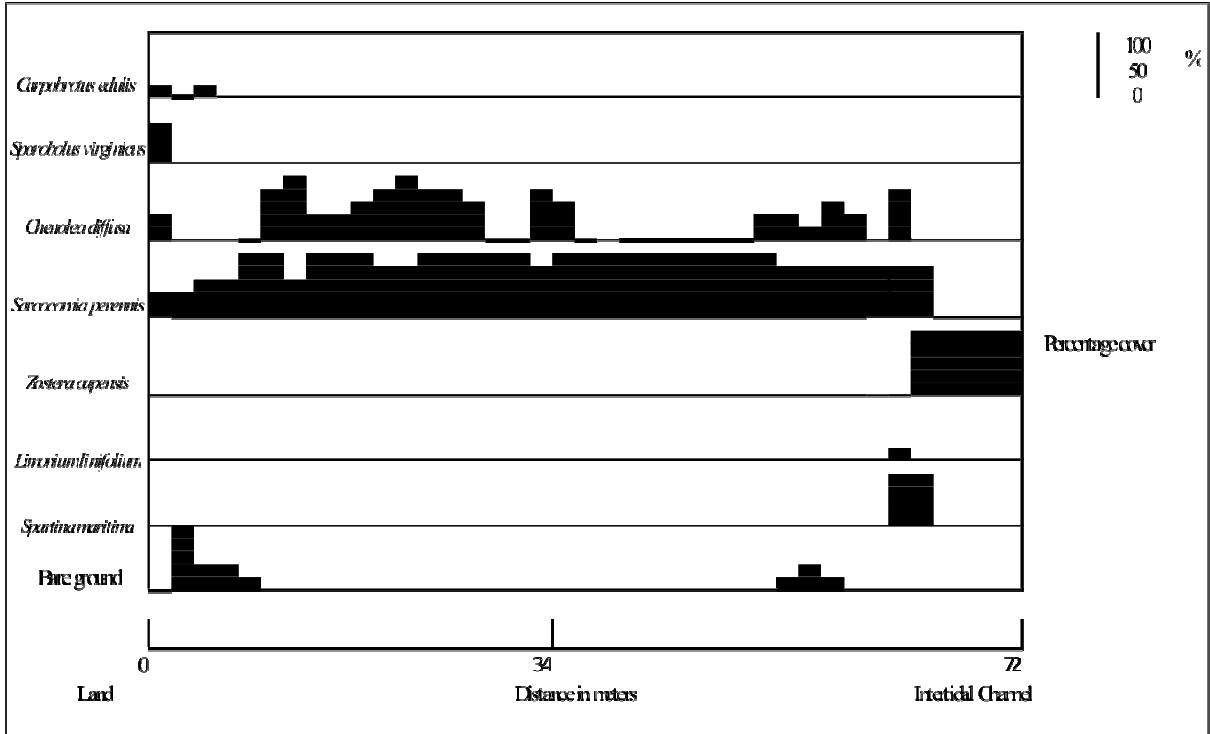


Plate 6.1: (A) *Carpobrotus edulis* (B) *Chenolea diffusa* (C) *Sporobolus virginicus* (D) *Limonium linifolium* (E) *Sarcocornia perennis*

A



B

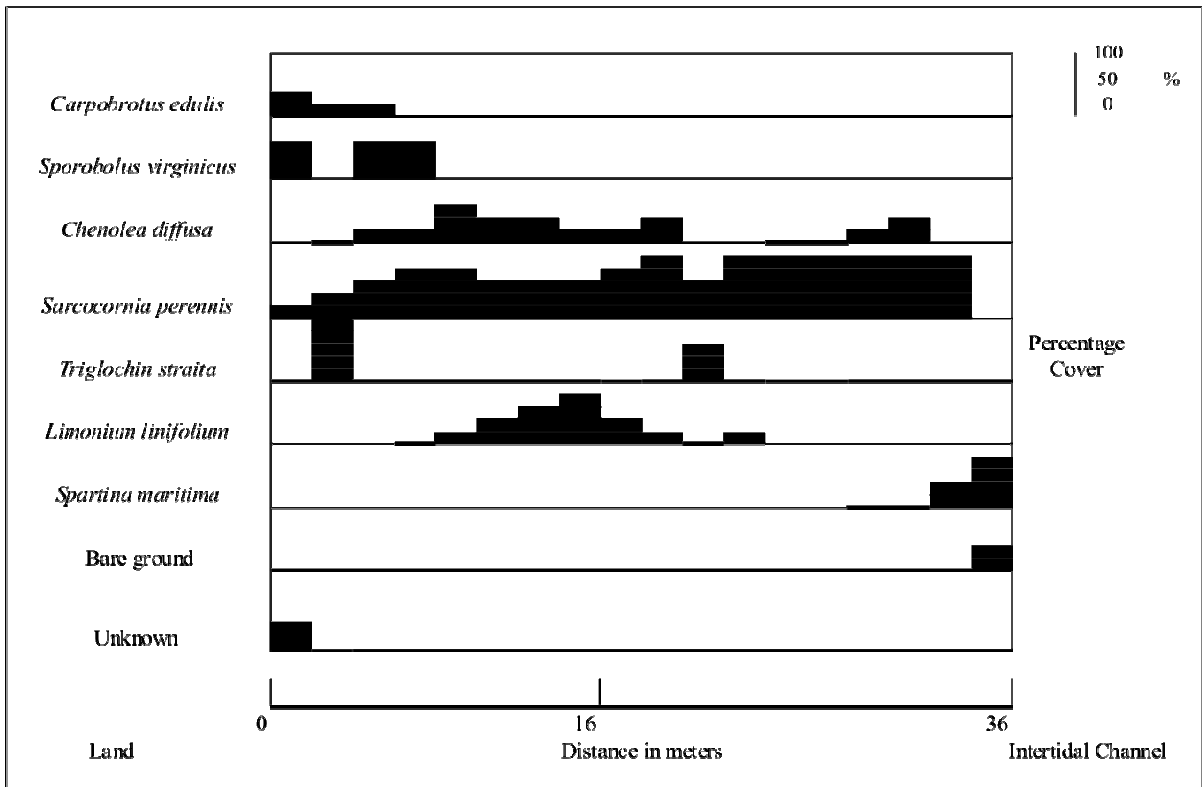


Figure 6.2: Zonations of the dominant plants along (A) Transect Four and (B) Transect Two

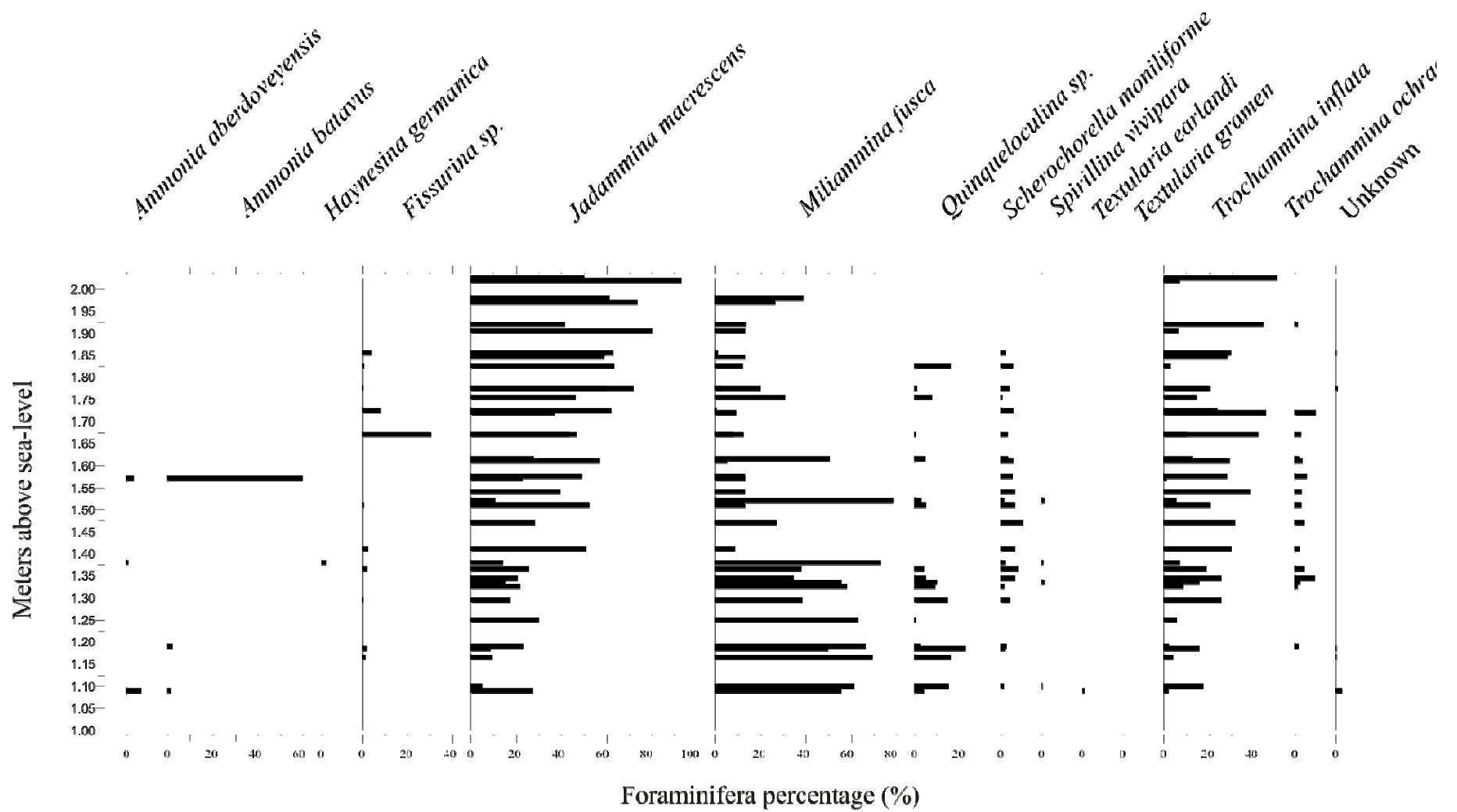


Figure 6.3: Surface foraminiferal assemblages (40% live and 60% dead) (X-axis scales are independent)

present, the proportion of agglutinated species decreases, and are replaced by an increase in calcareous species. Here there is a rapid decrease in *Trochammina inflata* and *Miliammina fusca* and an increase in *Ammonia batavus*, *Ammonia aberdoveyensis* and *Haynesina germanica*. The greatest abundance of foraminiferal species is present at lower elevations in the lower to tidal mudflat zones (Figure 6.3).

6.3 Foraminiferal assemblages and dominant plant ranges

The numbers of live species in the samples were relatively higher towards the lower marsh, with dead species abundance being greater towards the terrestrial edge (Figure 6.4 and 6.5). There is some correlation in the distribution of the zonations of the plants and foraminiferal assemblages. However, the correlation in distribution could be a manifestation of salinity or other physical parameters as opposed to a direct causation. In the tidal mudflat zone closest to the intertidal channel there is a noticeable increase in *Zostera capensis* and in calcareous species such as *Ammonia batavus*, *Ammonia aberdoveyensis*, *Spirillina vivipara* and *Haynesina germanica*. Where *Spartina maritima* is present along Transect Two (Figure 6.5) the percentage abundance of *Miliammina fusca* increases. As one moves towards the terrestrial edge the abundance of *Trochammina inflata* increases. Where the abundance of *Trochammina inflata* is relatively constant, there is an even distribution of *Chenolea diffusa* and *Sarcocornia perennis*. Where there is a clear increase in the number of *Jadammina macrescens* species *Sporobolus virginicus* is present.

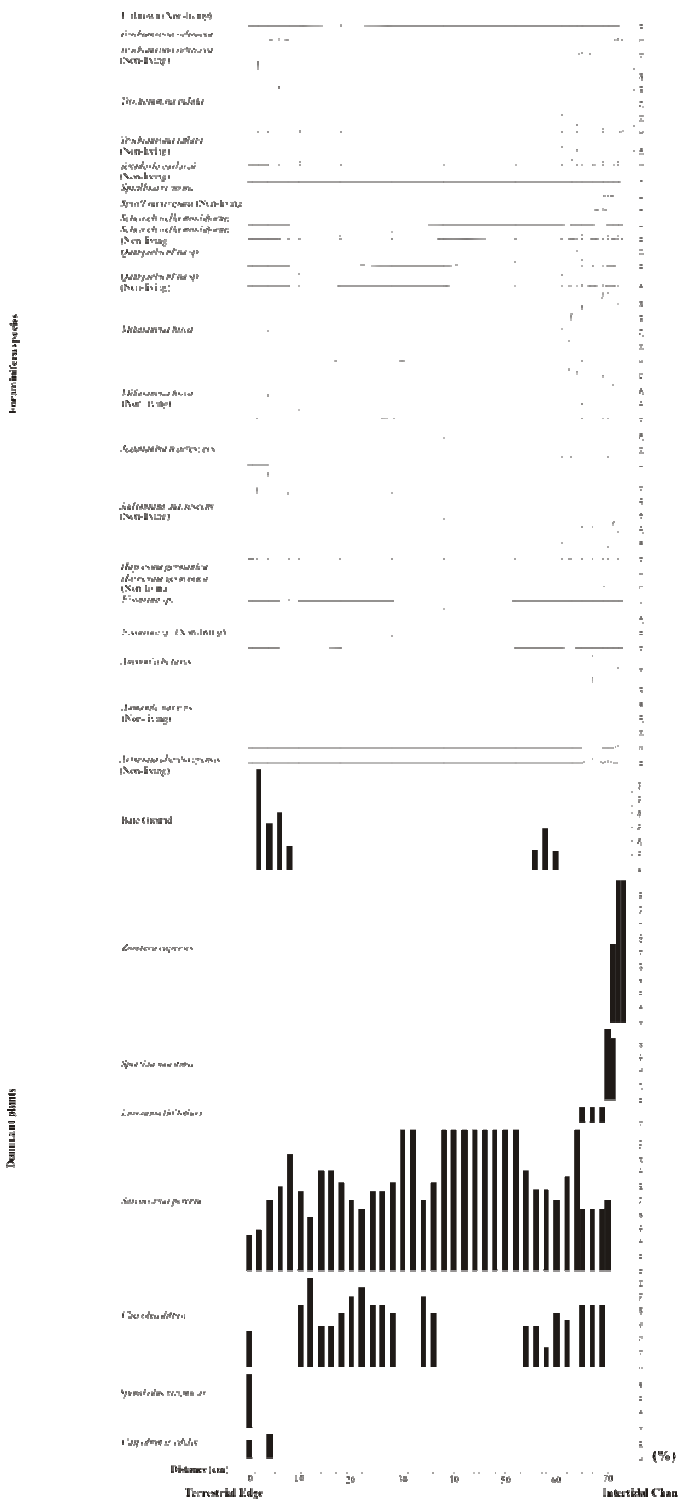


Figure 6.4: Vertical zonation along Transect Four of foraminifera in Galpin's salt marsh, with ranges of dominant plants

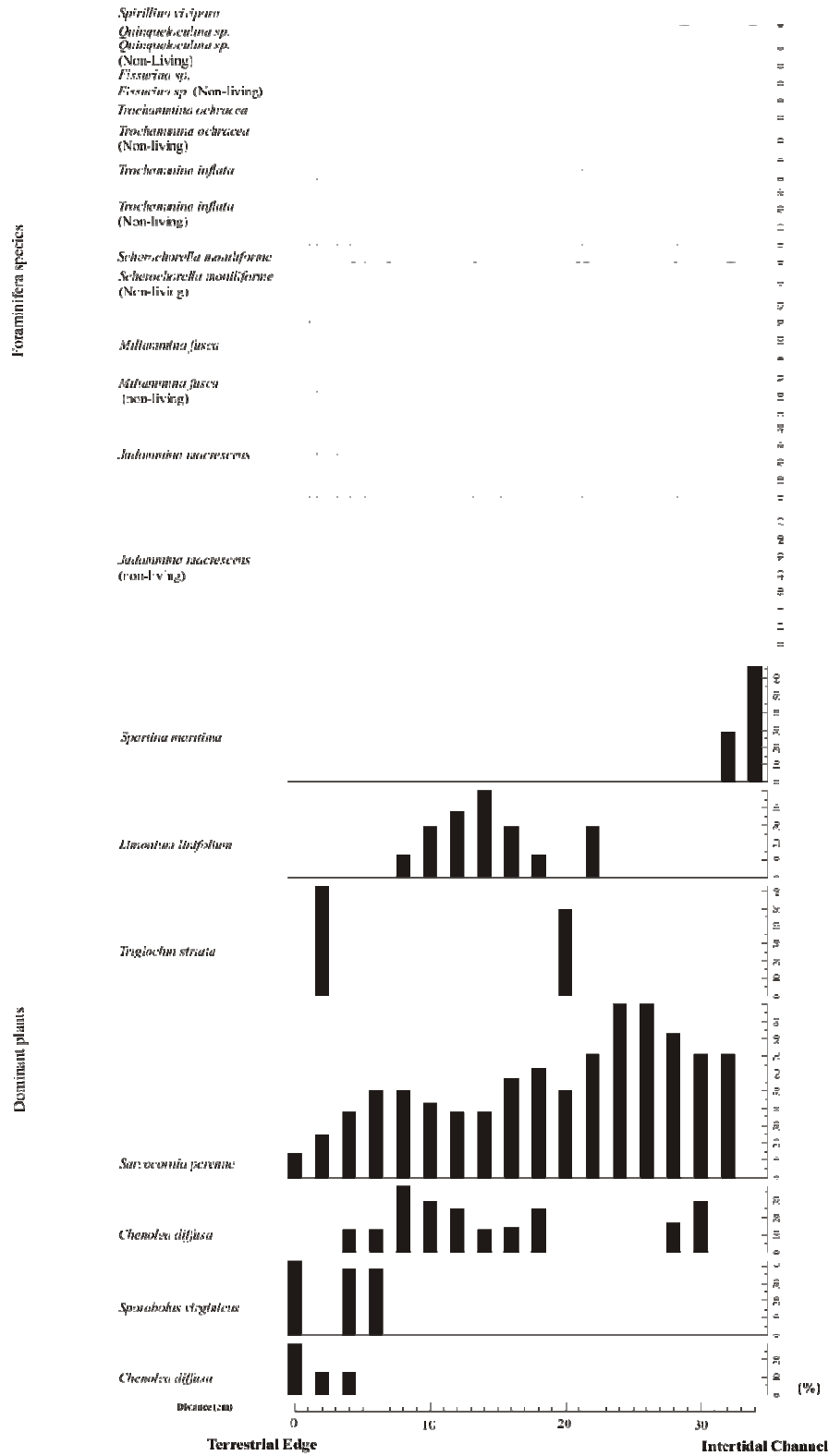


Figure 6.5: Vertical zonation along Transect Two of foraminifera in Galpin's salt marsh, with ranges of dominant plants

6.3.1 Optima and tolerance of foraminiferal species

The weighted-averaging analysis of the height-normalized training set yields tolerance ranges for each species present. It identifies the optimum elevation for each species relative to mean sea level (Figure 6:6). It appears that the optimum for most species occurs in the middle to slightly lower marsh. However, those taxa whose number of occurrence is low in the training set seem to appear closer to lower marsh habitats.

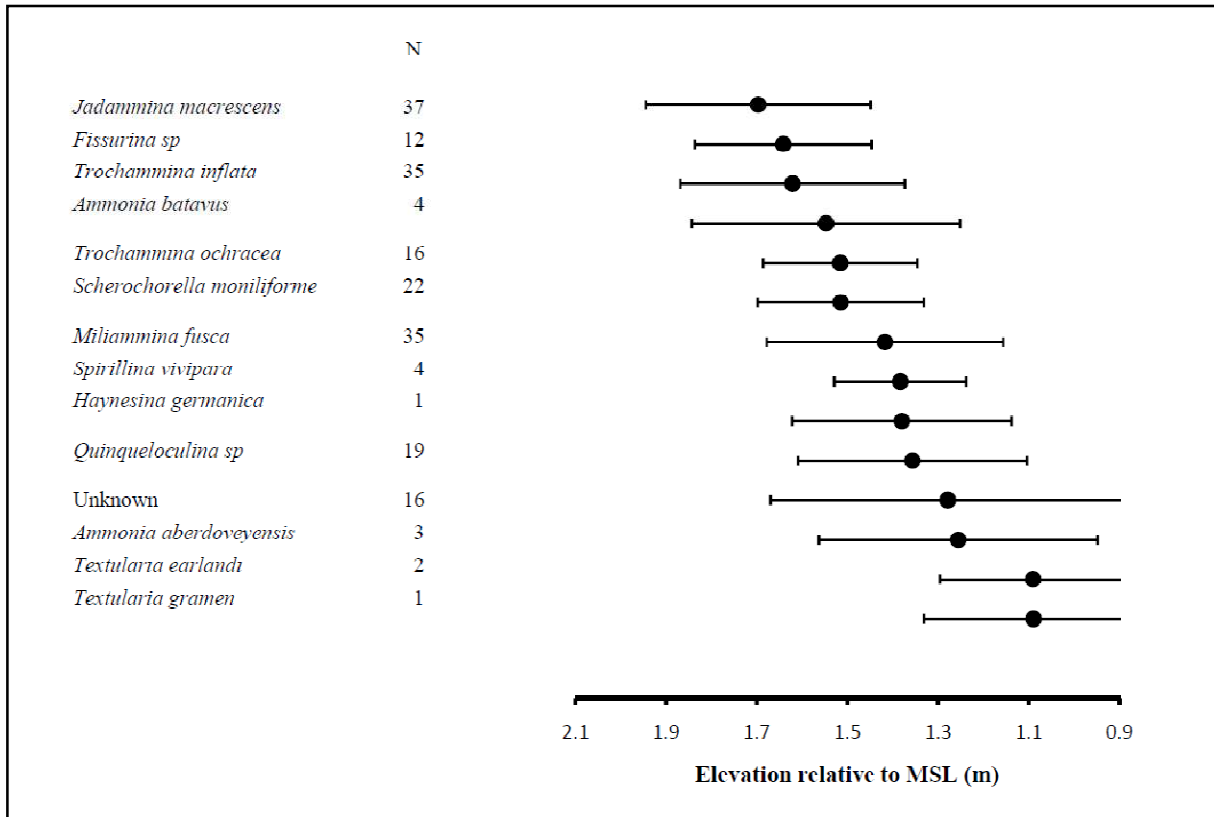


Figure 6:6: Optima and tolerance of total (live and dead) foraminiferal species. N is the number of occurrences in the training set of 37 samples.

6.4 Stratigraphic transect

Seven cores were recovered along Transect Two, starting 6 m from the channel edge moving towards the terrestrial edge (Figure 6.7). The majority of the cores were extracted from the middle marsh zone, an area with abundant *Chenolea diffusa* and *Limonium linifolium*. The upper 10 cm of each core had a high clay and humus content with the presence of silt and rootlets. The rich salt-marsh peat layer becomes deeper as one moves towards the terrestrial edge. This layer is evident in the three cores extracted at 18 m, 21 m and 26 m away from the channel edge, with the salt-marsh peat layer lying between 10 cm and 30 cm down the cores. This layer could be identified by a grey-brown colour with the presence of iron oxide stain, high concentration of clay and the high organic content. As one moves down each of the cores the organic-rich salt-marsh sediment is underlain either by a large layer of silt-clay with the presence of sand. A sand, shell basal layer was found in the core removed 21 cm from the channel edge. The sand content was otherwise restricted to the upper reaches of each core approximately 10 – 30 cm down, core dependent.

6.5 Core Two stratigraphy

The sediment core (Figure 6.8) used for foraminiferal analysis was chosen based on the stratigraphic information and the organic material present. This core was subdivided into nine stratigraphic units; these zones are based on colour and composition as described by Munsell Colour Code and Troels-Smith sediment description. The boundaries of the nine zones down the core appear at the following depths: 2, 7, 26, 30, 38, 42, 53, 56 and 97 cm.

Based on Troels-Smith sediment descriptions the bottom (31 -97 cm) of the core is predominately comprised of *Argilla steatodes*, *Argilla granosa* and *Grana minora*. The zone above, lying 26 - 30 cm has a high presence of *Limus ferrugineus*, with the preceding zone from 7 – 26 cm comprising solely of *Argilla steatode*. Above this zone there was a portion (4 -7cm) with a high *Argilla steatodes* and *Substantia humosa* content along with loose plant and root material. The upper two centimeters based on Troels-Smith sediment descriptions is dark brown in colour and comprises of *Detritus herbosus* and *Argilla steatodes*. The colour of the sediment core varies from very dark gray to dark grayish brown (plate 6.2).

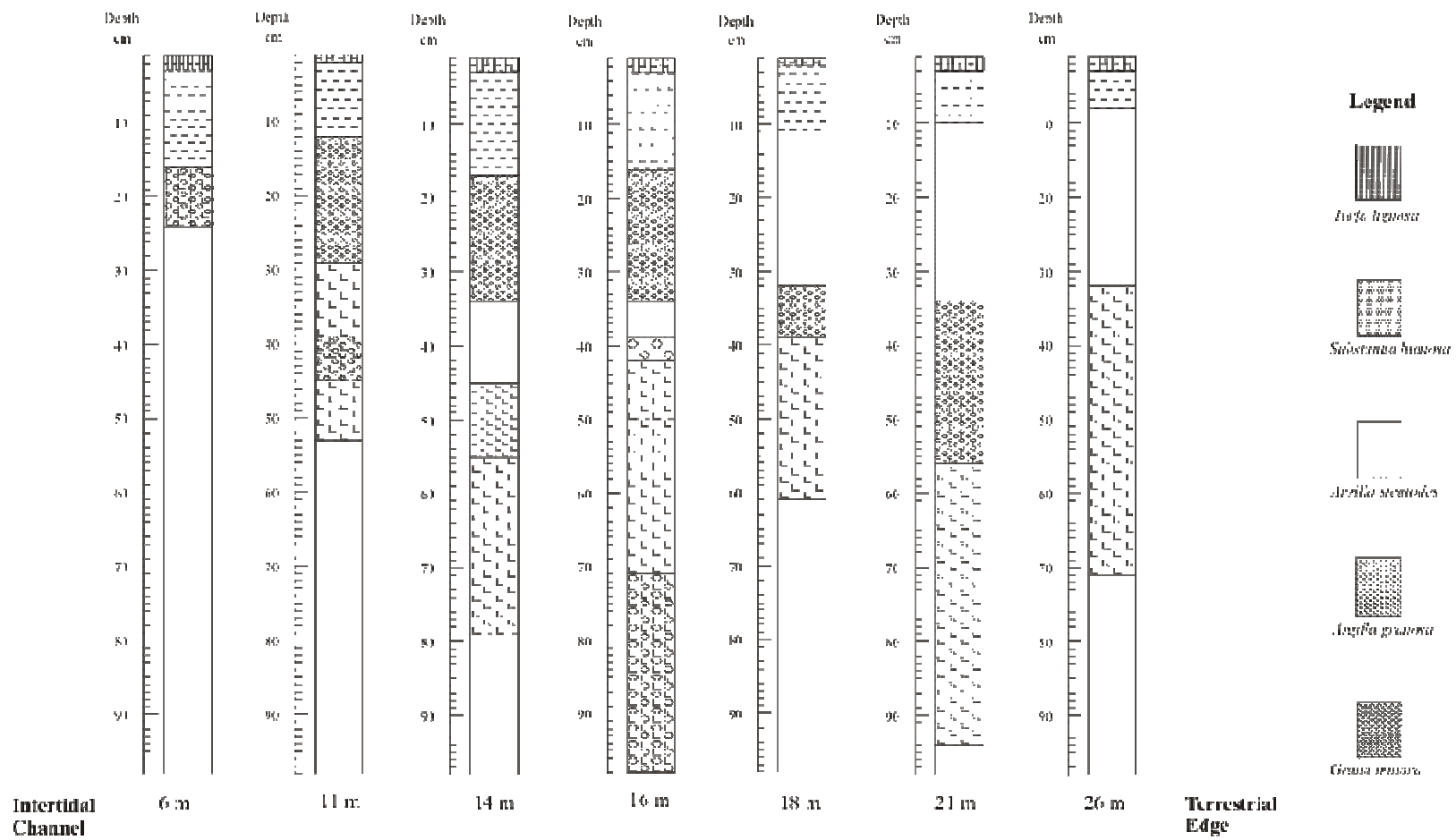


Figure 6.7: Troels-Smith stratigraphy of Transect Two Galpin's salt marsh, Kariega Estuary

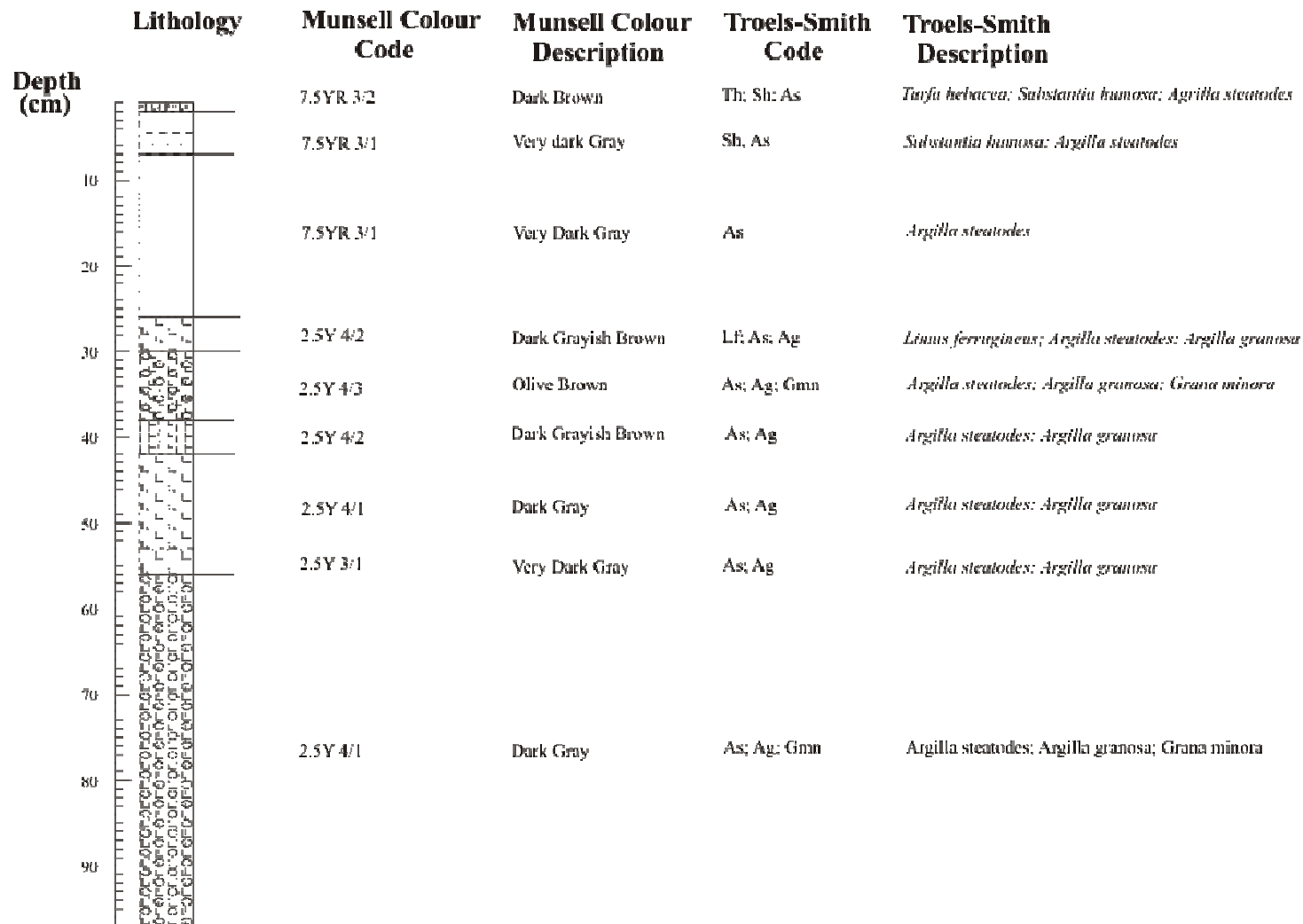


Figure 6.8: Stratigraphic description of Core Two (Appendix B)

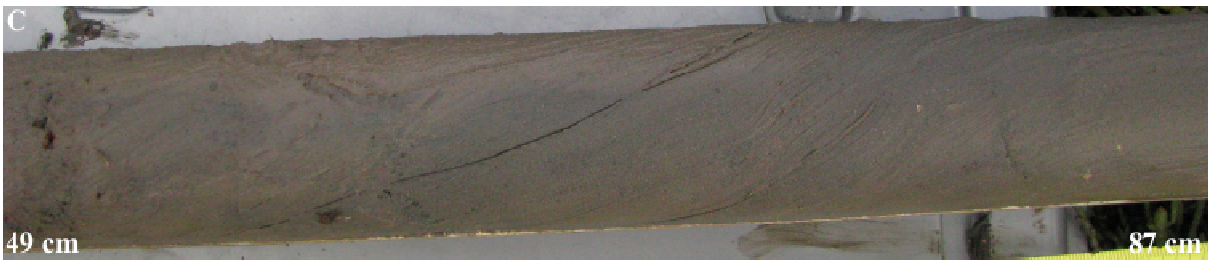
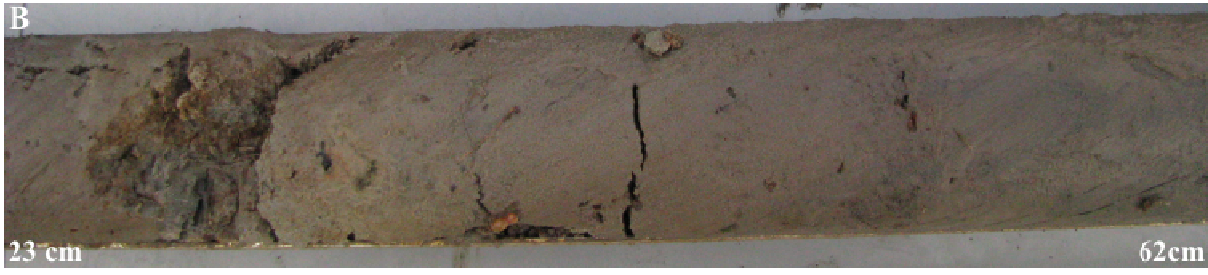


Plate 6.2: Core Two sediments

6.6 Age model and sedimentary history

The chronology for the core was determined through AMS dating on four bulk organic sediment and one shell sample (Table 6.1). The basic age-model (Figure 6.9), constructed by linear interpolation, for Core Two indicates two age reversals. The basal shell date is a clear outlier to all the bulk sediment dates. The chronology is non-linear and places the record within the late Holocene period with a basal age of 1331-1424 cal years BP.

Table 6.1: Radiocarbon dating results from Galpin's salt marsh, Kariega Estuary

Sample No	Lab Code	Depth (cm)	Sample material	¹⁴ C yr BP	cal year BP Range	¹³ C/ ¹² C (o/oo)
1	Beta-334778	18-19	Organic sediment	120±30	970 -1142 (92.1%)	-20.5
2	Beta-301135	53-54	Organic sediment	620±30	591-638 (49.7%)	-20.8
3	Beta-334779	66-67	Organic sediment	1460±30	1281-1363 (95%)	-19.8
4	Beta-334780	89-90	Organic sediment	1570±30	1331-1424 (69.6%)	-19
5	Beta-3301136	92-93	Shell	673±30	557-654 (95%)	1.5

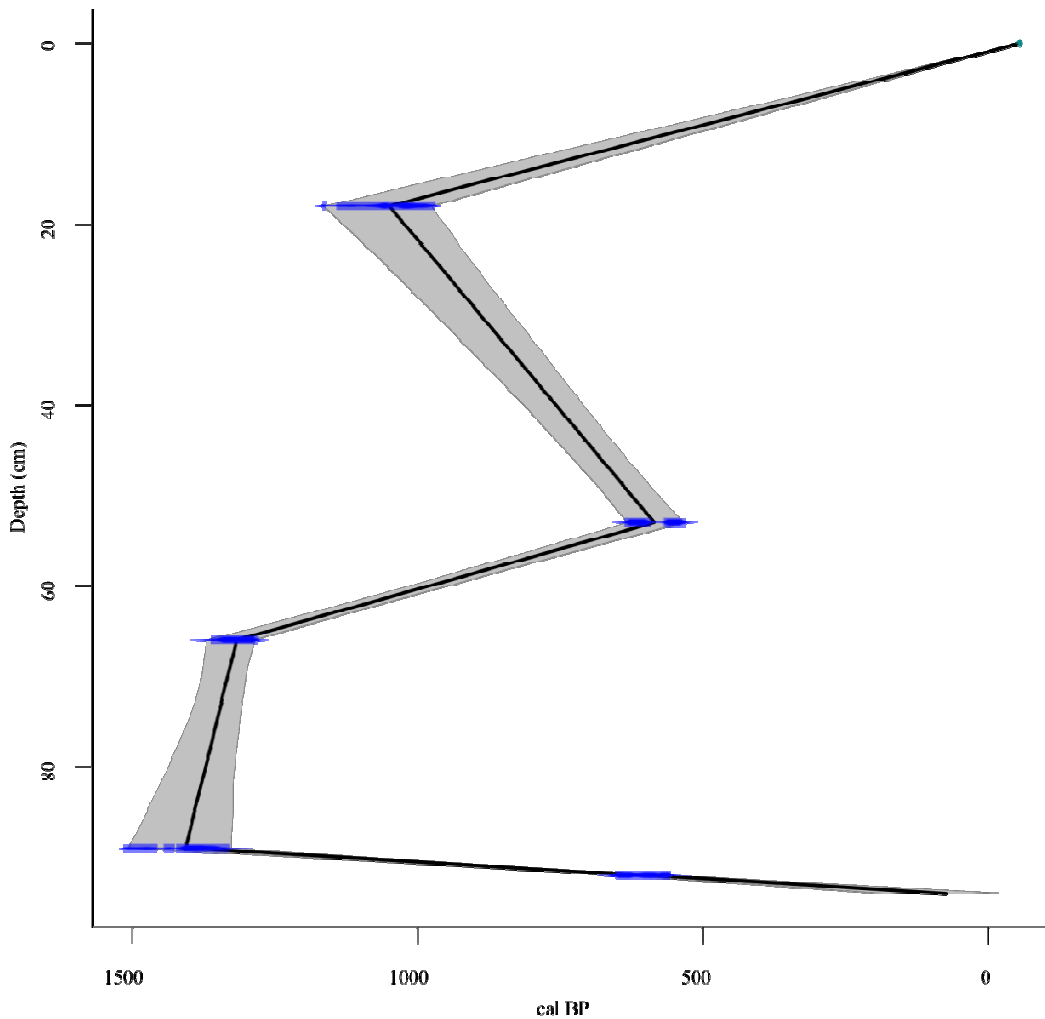


Figure 6.9: Basic age-depth model for Core Two, Galpin's salt marsh, Kariega Estuary

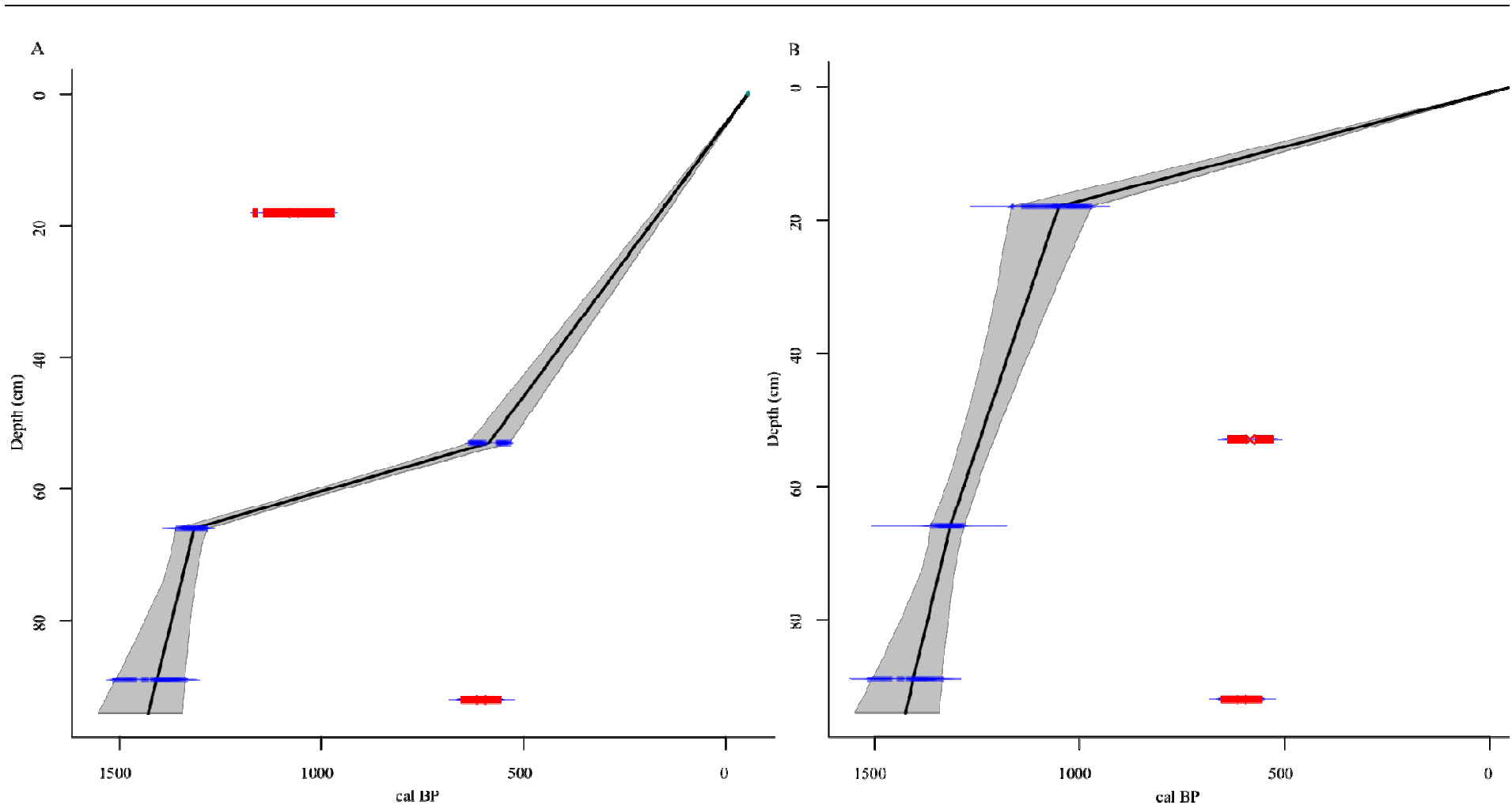


Figure 6.10: Age-depth models for (A) Scenario One (B) Scenario Two for Galpin's salt marsh, Kariega (outliers indicated with an x)

The fossil marine shell material has a date of 557-654 cal years BP which conflicts with the age model. Carbonate shells are notoriously more difficult to date because the carbon in their structure undergoes an exchange with carbon dioxide resulting in a younger age. The fossil shell date is 772 years younger than the 1331-1424 cal years BP date above it in the sediment core, possibly due to recrystallisation (Compton, 2011). The shell has a 1.5 ‰ for $\delta^{13}\text{C}$ but usually marine shells analysed show a range of $\delta^{13}\text{C}$ from +4.2 to -1.7‰, whereas mollusk shells derived from fresh water have a range between -0.6 to -1.5‰ with similar differences in $\delta^{18}\text{O}$ content. Without the oxygen content there is no guarantee that recrystallisation has taken place (Compton, 2011).

To calibrate radiocarbon ages based on samples with a marine carbon element it is important to know the marine carbon reservoir correction value (Dewar *et al.*, 2012). Finding the appropriate correction factor for marine radiocarbon dates is difficult and the problem has yet to be fully resolved (Walker, 2005). In this particular research we use a west coast weighted mean ΔR of 146 ± 85 ^{14}C years to correctly calibrate marine shell (Dewar *et al.*, 2012). The radiocarbon date of the shell material was not included in the final age-depth models as possible recrystallisation has taken place and conflicts with the rest of the dates obtained.

Although the radiocarbon dates are fairly consistent with one another as they increase in age down the sediment core, there are apparent conflicts. Results indicate two possible age reversals in the upper reaches of the core, which are best viewed as separate scenarios. In Scenario One the 970-1142 cal years BP date at 18 cm and in Scenario Two the 591-638 cal years BP date at a depth of 53 cm have been identified as the possible outliers (Figure 6.10). Age reversals evident in the upper sections of the core suggest that there was either reworked organic matter from bioturbation, contamination by aquatic plant rootlets, erosion or deposition occurring as a result of wave action or river deposition. Scenario One shows a varying curve while Scenario Two shows a relatively linear curve.

In Scenario One the radiocarbon date 970-1142 cal years BP at a depth of 18 cm has been excluded. This age reversals suggest that there was either reworked organic matter from bioturbation, contamination, erosional or deposition hiatus occurring as a result of wave action or the probability of river deposition. The 970-1142 cal years BP date would suggest that the sample was either contaminated by old carbon as the sample extracted below is younger. The contamination could be as a result of older sediments being deposited (Retelle *et al.*, 1989) by wave action or bioturbation. The material in a salt marsh can be reworked by spring tides and some of it distributed elsewhere in the marsh (Stoddart *et al.*, 1989). The eroding marsh edge is a major source of sediment, with sediment being transported into the marsh during tidal fluctuations over the marsh surface (Stoddart *et al.*, 1989).

Scenario One's sedimentation rates (Table 6.2) show a considerable increase in recent times. The increase in sedimentation rate can be attributed to the increase in tidal inundation which allows greater opportunity for sediment to reach the marsh surface (Stoddart *et al.*, 1989). A study conducted on Long Island marsh in Connecticut suggests that over the last 100 years a high rate of deposition has occurred. This study concluded that with a slow rise in sea level, the marsh accumulates at a rate determined by sea level variation, with the level of the marsh surface relative to mean sea level determined by sediment supply and frequency of tidal flooding (Stumpf, 1982). An increase in sedimentation is in conjunction with studies that propose that sedimentation continues throughout the Holocene and modern salt marshes generally have higher rates of sedimentation than previously (Stumpf, 1982; Retelle *et al.*, 1989; Stoddart *et al.*, 1989). An erosional hiatus would explain the sedimentation rate of 0.02 cm/year between 1281-1363 cal years BP and 591-638 cal years BP. According to the sediment description of Core Two, the stratigraphy indicates that the bottom (31 – 97 cm) is predominately comprised of *Argilla steatodes*, *Argilla granosa* and *Grana minora*. Near the top of the core change in sedimentation is more evident and is almost solely comprised of *Argilla steatodes*, with small fractions of *Substantia humosa*.

In Scenario Two the radiocarbon date of 591-638 cal years BP extracted at a depth of 53 cm has been excluded. This age reversal suggests that there is either an erosional or depositional hiatus at this depth; it was contaminated by younger carbon as a result of bioturbation. As discussed in Scenario One, spring tides could rework the sediment and mixing can occur (Stoddart *et al.*, 1989).

Scenario Two is close to linear, which implies a steady sedimentation rate (Table 6.3) except between 0 cm and 18 cm where there is a significant decrease in sedimentation rates in recent years. The layer in the sediment core suggests steady sedimentation as it is fairly homogeneous between 56 cm and 94 cm with a sedimentation rate of 0.22 cm/year. A slight decrease in sedimentation rates is possible if the elevation of the marsh surface increases over time (Stoddart *et al.*, 1989) or the tidal range has decreased as it was found that areas with higher tidal ranges had higher sedimentation rates (Stumpf, 1982).

This suggests that estuarine sedimentation transport process are complex, with varying influences on the sediment delivery to the marsh surface and that sediment availability for deposition on the marsh surface is not only controlled by flooding. In terms of parsimony, Scenario Two is the most likely as it is linear and has a steady sedimentation rate which correlates well with findings that sedimentation was continuous throughout the Holocene, however if this scenario is seen to be correct then the period between 970-1142 cal years BP and the present would be considered as unreliable as sedimentation rates decrease and a possible erosional hiatus present. On the other hand if Scenario One was correct, as they are more comparable to findings that sedimentation rates have continued to increase as sea-level increases, then the period between 1331-1424 cal years BP and 1281-1363 cal years BP would be unreliable as during this period of time sedimentation rates decreased, again due to a possible erosional hiatus.

Table 6.2: Sedimentation rates for Scenario One

Sample No.	Depth range (cm)	Depth thickness (cm)	Sedimentation rate (cm/yr)
2	0 - 53	53	0.09
3	53 - 66	13	0.02
4	66 - 89	23	0.41

Table 6.3: Sedimentation rates for Scenario Two

Sample No.	Depth range (cm)	Depth thickness (cm)	Sedimentation rate (cm/yr)
1	0 - 18	18	0.02
3	18 - 66	71	0.22
4	66 - 89		

6.7 Down core species description

Foraminifera occurred throughout the length of the core and were well preserved, however counts were low at certain depths particularly between 30 and 65 cm. To assist in the interpretation of the foraminiferal assemblages down Core Two, the foraminiferal sequence was divided into five zones, viz., K1 (65-94 cm), K2 (46-65 cm), K3 (39-46 cm), K4 (31-39 cm) and K5 (0-31 cm), using the Constrained Incremental Sum of Squares (CONISS) technique (Figure 6.11). Agglutinated foraminifera were most abundant in the upper reaches of the core and gradually decreased down the core. The base of the core was dominated by calcareous foraminiferal species (Figure 6.12).

In zone K1 *Ammonia aberdoveyensis* and *Haynesina germanica* appear to be dominant down to the base of the core, with agglutinated species especially *Textularia gramen* appearing sporadically. Zone K2 is the transition zone in the core, where calcareous species become present and the proportion of agglutinated species decreases. There is an apparent peak in the abundance of *Ammonia batavus* in K2; especially from 54 to 62 cm. *Miliammina fusca* exhibits an increase at a depth of 50 to 52 cm. *Jadammina macrescens* is the most abundant species in zone K3, with a small presence of *Miliammina fusca* and *Trochammina inflata*. A very low number of foraminifera were recovered from between the depths of 30 cm and 46 cm. Zones K4 and K3 are similar in sediment composition and were made up of approximately equal parts clay and silt. In zone K4 *Jadammina macrescens* is the dominant species, with three species appearing in low counts between 31 and 36 cm, viz., *Miliammina fusca*, *Trochammina inflata* and *Trochammina ochracea*. Zone K5 had five dominant species, viz., *Jadammina macrescens*, *Miliammina fusca*, *Scherochorella moniliforme*, *Trochammina inflata* and *Trochammina ochracea*. Species such as *Jadammina macrescens*, *Miliammina fusca*, *Trochammina inflata* and *Trochammina ochracea* were generally restricted to the upper 46 cm,

whereas *Ammonia aberdoveyensis*, *Ammonia batavus* and *Haynesina germanica* were restricted to the lower half of the core.

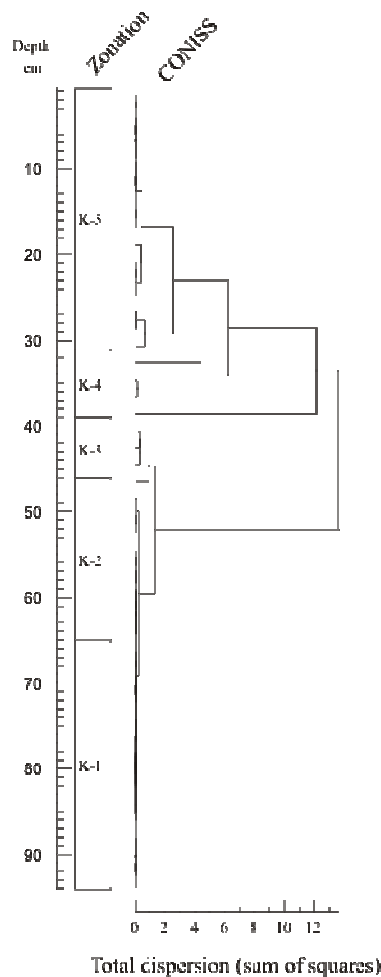


Figure 6.11: Zonation and CONISS of foraminiferal assemblages down Core Two

6.8 Sea-level reconstruction

The transfer function was applied to calibrate the fossil assemblages from Core Four and the results reconstructions compared. Summary of the performance of the transfer function can be seen in Appendix K. Relative sea-level change is reconstructed for Scenario One and Scenario Two from the data obtained from the transfer function and the age-depth model, relative sea-level change was also constructed and plotted against depth (Figure 6.13). For both Scenario One (Figure 6.14) and Two (Figure 6.15) the sea level was within one meter of the present day level, going back to 1500 cal years BP. The reconstruction for both scenarios indicate that sea level was 0.6 m higher than the present between 1300 cal years BP and 1500 cal years BP.

In Scenario One (Figure 6.14) after 1300 cal years BP sea-level fell relatively rapidly to between 0.2 m and 0.4 m below present until 600 cal years BP, at which point it rose again to above the present level. Once again between 600 cal years BP and 500 cal years BP sea-level dropped and reached the lowest in the record during this 1500 cal years BP period. From 500 cal years BP until present, sea-level fluctuated between 0 and 0.6 m below present.

In Scenario Two (Figure 6.15) between 1300 cal years BP and 1400 cal years BP sea-level dropped rapidly. A sea-level highstand is then evident between 1300 cal years BP and 1200 cal years BP. After 1250 cal years BP sea-level fell to a minimum around 1200 cal years BP. Between 1100 cal years BP and 1200 cal years BP sea-level fluctuated between -0.3 m and -0.6 m. There was the relative stability in the sea level between 100 cal years BP and the present.

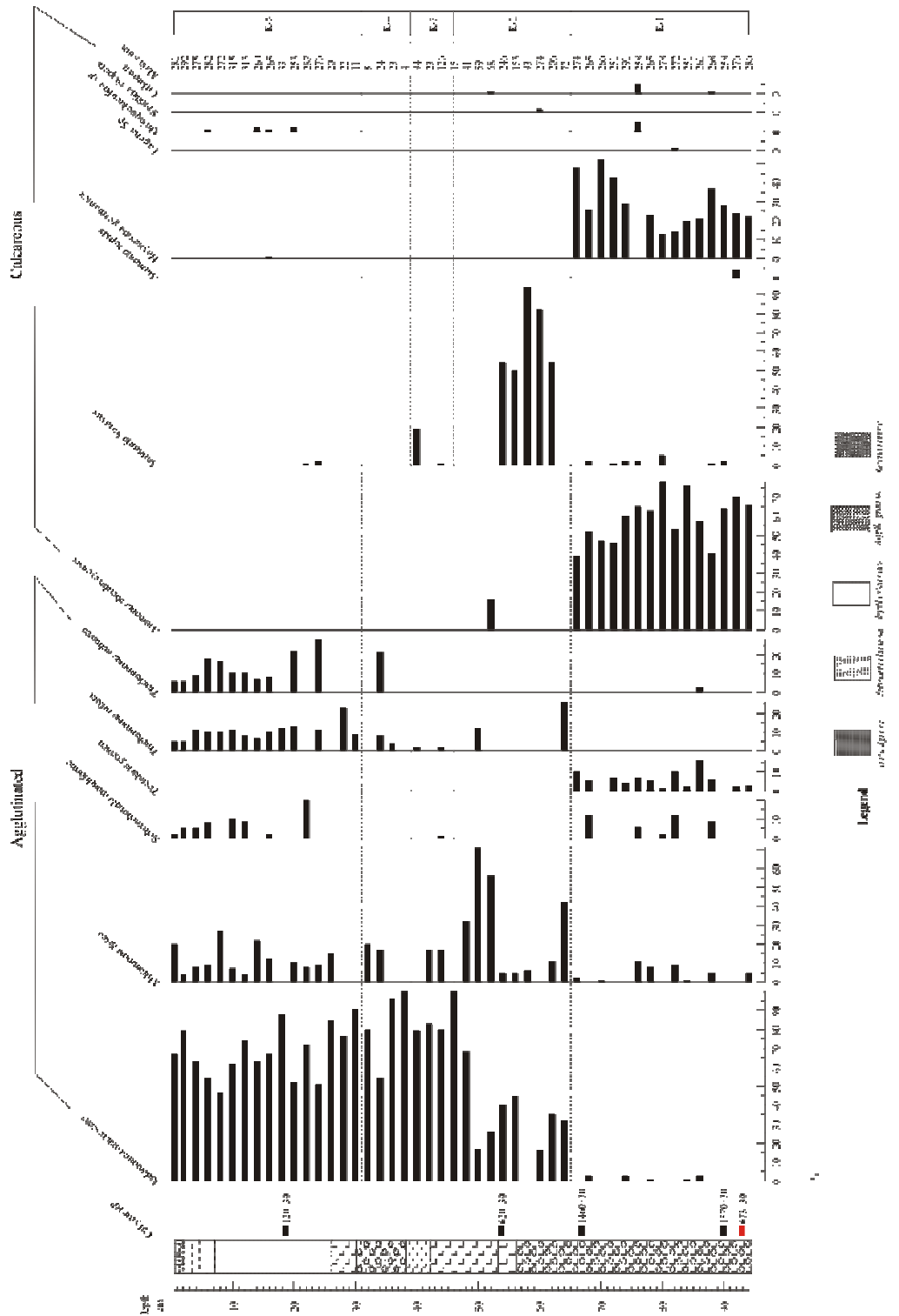


Figure 6.12: Foraminiferal concentrations down Core Two, Galpin's salt marsh. Indicated in the figure is Troels-Smith stratigraphy and CONISS derived zonation

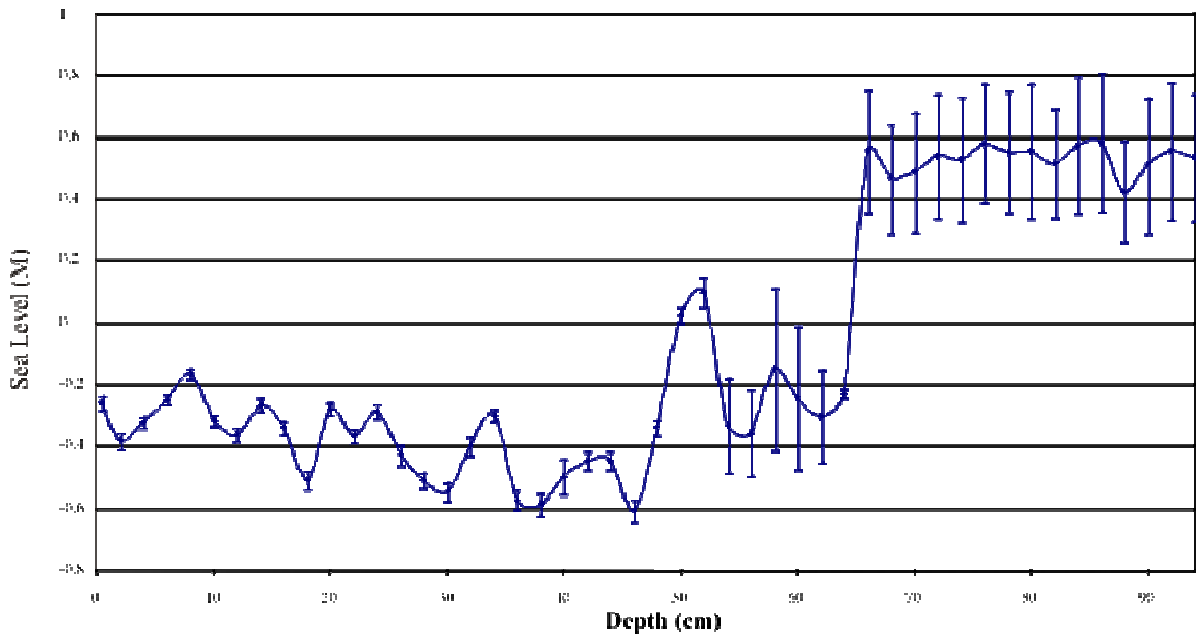


Figure 6.13: Reconstruction of relative sea level plotted against depth

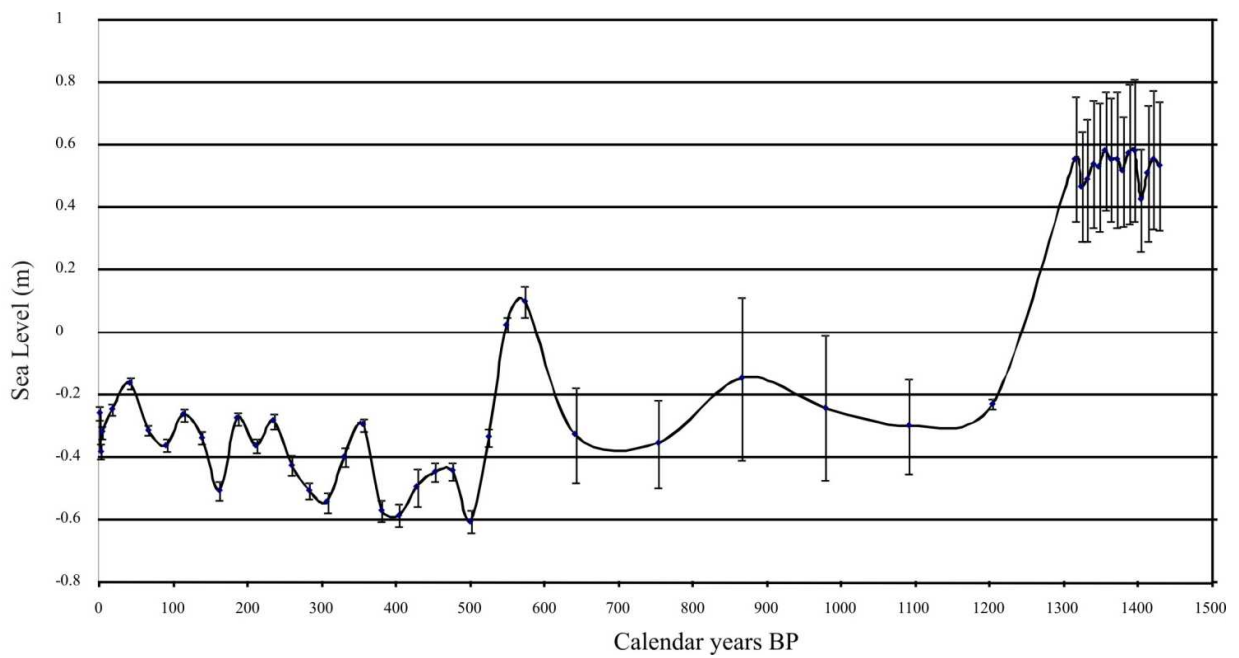


Figure 6.14: Reconstruction of relative sea level for Scenario One

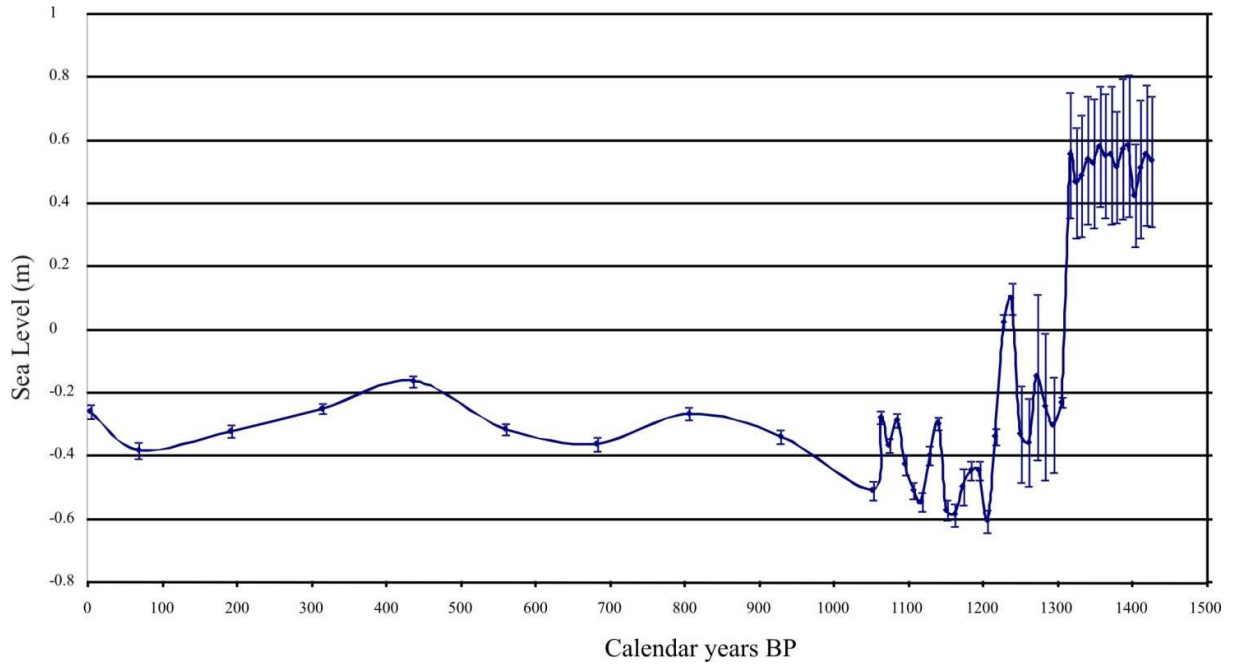


Figure 6.15: Reconstruction of relative sea level for Scenario Two

CHAPTER SEVEN: DISCUSSION

For sea level to be reconstructed with the aid of salt-marsh foraminifera, one needs to have a comprehensive understanding of their surface distributions, in conjunction with understanding their relationship with environmental variables that influence these distribution patterns (Scott and Medioli, 1978; Patterson, 1990; Edwards and Horton, 2000; Horton and Edwards, 2006). In the following chapter the distribution of modern foraminiferal assemblages within the intertidal zone are described and compared to the dominant salt-marsh plants zonation and tidal fluctuation. This established relationship between foraminiferal distributions and tidal levels, is used to develop a foraminifera transfer function which is able to produce a relative reconstruction of past sea levels. Finally, the sea-level reconstruction is discussed and compared with previous results from local and international literature.

7.1 Transitional zones determined by foraminifera assemblage changes

The results (Figure 6.12) suggest that foraminiferal assemblages can determine vertical zonations within Kariega Estuary. The distribution of foraminiferal assemblages across the surface of the intertidal zone is a direct result of altitude, though the duration and frequency of intertidal exposure must be considered as these factors may affect distribution (Scott and Medioli, 1978, Jennings *et al.*, 1995; Franceschini *et al.*, 2005).

Kariega Estuary has a relatively diverse abundance of foraminiferal assemblages (Table 7.1) compared with studies conducted along the west coast of South Africa (Compton, 2001; Franceschini *et al.*, 2005), which tend to be dominated by agglutinated species. The terrestrial edge of the salt marsh has a high percentage of *Jadammina macrescens* with a determined optimum of 1.69 m above sea level (Figure 6.12), this increases higher up the marsh. This concurs with previous studies which suggest that *Jadammina macrescens* is typically found in the highest parts of the intertidal zone and its relative abundance rapidly declines towards the intertidal channel (Gehrels and van de Plassche, 1999; Sen Gupta, 2003; Horton and Edwards, 2006). However, in this research the high marsh zone contains a small number of *Trochammina inflata*, which has been identified as a high and middle marsh species in samples taken from the Fraser River delta in British Columbia (Canada) where it was found to co-dominate with *Jadammina macrescens* (Williams, 1989; Patterson, 1990). The co-inhabitation of these two species in Kariega Estuary was observed and described for Langebaan (Franceschini *et al.*, 2005).

Miliammina fusca and *Trochammina inflata* predominantly occur in conjunction with each other (Figure 6.9) and are good representatives of vegetated marshes worldwide (Scott and Medioli, 1978; Berkeley *et al.*, 2007). According to Horton and Edwards (2003), investigating the seasonal distribution of foraminiferal

species at Cowpens marsh in England, the transition zone between high and low marsh is characterized by a decrease in *Trochammina inflata* and an increase in *Miliammina fusca*. This trend is evident in Kariega Estuary surface samples, with *Miliammina fusca* abundance increasing and *Trochammina inflata* numbers decreasing as altitude decreases towards the intertidal channel.

Table 7.1: Dominant foraminifera identified in this research at Kariega Estuary and their preferred niche

Foraminifera	Niche
<i>Ammonia aberdoveyensis</i>	Low marsh and tidal flats
<i>Ammonia batavus</i>	Low marsh and tidal flats
<i>Haynesina germanica</i>	Middle/low marsh and tidal flats
<i>Fissurina spp</i>	Marine shelves and transported into estuarine systems
<i>Jadammina macrescens</i>	Brackish marshes
<i>Miliammina fusca</i>	High marsh and upper estuary
<i>Quinqueloculina spp</i>	Estuary mouths, inner shelf and middle/low marsh
<i>Scherochorella moniliforme</i>	Inner shelf and estuary mouths
<i>Spirillina vivipara</i>	Inner shelf
<i>Textularia earlandi</i>	Coastal environments
<i>Textularia gramen</i>	Coastal environments
<i>Trochammina inflata</i>	High marsh and upper estuary
<i>Trochammina ochracea</i>	Inner shelf

(Adapted from: Murray, 1979; Horton and Edwards, 2006)

The transition between low marsh and tidal mudflats corresponds with the percentage of agglutinated species decreasing and being replaced by an increased number of calcareous species in Kariega Estuary's surface foraminiferal assemblages (Horton and Edwards, 2006). Similar studies from Europe, North America and Langebaan Lagoon salt marsh in South Africa identified the mudflats as being dominated by calcareous foraminiferal species (Patterson, 1990; Horton, 1999; Franceschini *et al.*, 2005). This transition zone at Kariega Estuary is characterized by a rapid decrease in *Trochammina inflata* and *Miliammina fusca* and being replaced by *Haynesina germanica* and *Ammonia* species, as identified at Langebaan Lagoon salt marsh (Compton, 2001) and at Cowpen marsh (Horton and Edwards, 2003). Species of the genus *Haynesina* and *Ammonia* are often found to populate the lower tidal to intertidal zones (Berkeley *et al.*, 2007). Tidal mudflats are usually characterized by the highest foraminiferal diversity (Franceschini *et al.*, 2005), as is the case at Kariega Estuary where the highest abundance of ten foraminiferal species were present in the tidal mudflats. *Miliammina*, *Trochammina* and *Ammonia* assemblages usually characterize modern paralic and deltaic environments (Jones, 2006).

Species of the genus *Ammonia* are said to dominate lower tidal to intertidal zones (Berkeley *et al.*, 2007), however in a salt marsh in north Queensland, Australia *Ammonia aberdoveyensis* were found isolated at 1.54

m above sea level, thus not corresponding with Compton (2001), Horton and Edwards (2003), and Berkeley *et al.* (2007). This may be an indication that the distribution of *Ammonia aberdoveyensis* is not directly proportional to altitude in this particular situation as hypothesized by Scott and Medioli (1978), but rather linked with the salinity of the salt marsh, as well as natural variables that would influence salinity such as precipitation and tidal fluctuation. Unfortunately, this cannot be verified for Kariega Estuary as salinity was not determined in this research.

7.2 Vegetation - foraminiferal zonations

Many salt marshes exhibit distinct plant zonations (Table 7.2) which are controlled by their tolerance to tidal inundation, competition (Gehrels, 1994) and biotic environmental factors such as altitude and salinity (Horton, 1999), such a strong correlation with elevation makes them extremely useful to sea-level research (Jennings *et al.*, 1995; Horton, 1999).

Three vegetation zones can be identified; mudflats, low marsh and high marsh, using the relative abundance of the dominant plants (Jennings *et al.*, 1995). Galpin's mud flat zone corresponds with Lubke and de Moor (1998) who describe successional changes in salt marshes, whereby the mud flats are colonized by *Zostera capensis*, followed by a low marsh zone stabilized by *Spartina maritima* and the water is shallower than that of the mud flats (Lubke and de Moor, 1998). These three distinct zones are apparent at Valdivia tidal marshes in Chile (Jennings *et al.*, 1995) and Langebaan, South Africa (Franceschini *et al.*, 2005). The low marsh zone occupies the zone between the low water mark and neap tide or high water mark (Gehrels, 1994), whilst the high marsh zone can be subdivided into the upper *Sporobolus virginicus* zone and the lower *Chenolea diffusa*, *Sarcocornia perennis* and *Limonium linifolium* zone.

According to the described plant zonations present at Kariega Estuary, both the tidal mudflats and the lower marsh zones are relatively narrow, resulting in the majority of the marsh being classified as a high marsh zone (Lubke and de Moor, 1998), similar to that of the Kowie and Swartkops Rivers in the Eastern Cape, South Africa (Lubke and de Moor, 1998). *Triglochin linifolium* indicates a middle marsh zone, however even though it is present it is rather dispersed throughout the marsh, and therefore there is very little evidence of a middle marsh zone (Lubke and de Moor, 1998). The vegetation zonation is fairly distinct as it progresses from the intertidal channel towards the terrestrial edge.

The findings for the salt marsh plant zonations in Galpin's salt marsh (Table 7.3) are in part similar to earlier observations of the transitional zones indicated by surface foraminiferal assemblages, but there are some important differences. Foraminiferal high marsh zone (Horton and Edwards, 2006) coincides with the high marsh vegetation zone. Similar observations have been made elsewhere in the northern hemisphere (Scott

and Medioli, 1978; Gehrels, 1994; Jennings *et al.*, 1995 and Horton, 1999), with a similar trend at Langebaan, South Africa (Franceschini *et al.*, 2005).

Table 7.2: General zonation of salt marsh plants for South Africa

Salt marsh plants	Salt marsh zone	Tide mark
Upper halophytic community	high marsh	
<i>Sporobolus virginicus</i>		
<i>Chenolea diffusa</i>		Spring tide mark
<i>Sarcocornia perenne</i>		
<i>Limonium linifolium</i>		
<i>Triglochin bulbosum</i>	middle marsh	
<i>Spartina maritima</i>	low marsh	Neap tide/high water mark
<i>Zostera capensis</i>	tidal mudflats	Low water mark

(Adapted from: Lubke and de Moor, 1998)

The low marsh zone, as identified by foraminiferal assemblages, weakly coincides with that of the low marsh vegetation zone. The transition zone between high and low marsh is characterized by a decrease in *Trochammina inflata* and an increase in *Miliammina fusca* (Horton and Edwards, 2003). However, this transition between the two zones is further away from the intertidal channel compared to the *Spartina maritima*, which is a low marsh species (Gehrels, 1994; Jennings *et al.*, 1995). Scott and Medioli (1978), Patterson (1990), Gehrels (1994) and Horton (1999) found low marsh floral zones to be dominated by *Miliammina fusca* with low numbers of *Haynesina germanica*, concordant with the findings of this research.

The foraminiferal and marsh vegetation tidal mudflat zones overlap, though the plant mudflat zone is more limited compared to that of the mudflat zone identified by the foraminiferal assemblages. This zone is characterized by a high diversity of foraminiferal species of which a significant number are calcareous (Horton, 1999; Franceschini *et al.*, 2005). Similar zones have been identified in mid-latitudes and cool temperate intertidal environments (Patterson, 1990; Horton, 1999) including Langebaan (Franceschini *et al.*, 2005).

Table 7.3: General zonation of plant and foraminifera species for Galpin's salt marsh

Dominant salt marsh plants	Salt marsh zone	Dominate foraminiferal assemblages
<i>Sporobolus virginicus</i>	high Marsh	<i>Jadammina macrescens</i>
<i>Chenolea diffusa</i>		
<i>Sarcocornia perenne</i>		<i>Trochammina inflata</i>
<i>Limonium linifolium</i>		
<i>Spartina maritima</i>	low Marsh	<i>Miliammina fusca</i>
<i>Zostera capensis</i>	tidal mudflats	<i>Haynesina germanica</i>
		<i>Ammonia batavus</i>
		<i>Quinqueloculina sp</i>

(Adapted from: Murray, 1979; Lubke and de Moor, 1998; Horton and Edwards, 2006)

7.3 Sea-level reconstruction

The proposed late Holocene sea levels from Kariega Estuary generally agree with previously proposed sea levels from the west, south and east coasts of South Africa. Evidence from Kariega Estuary supporting the highstand is noticeable in both Scenario One (Figure 6.14) and Scenario Two (Figure 6.15) taking place between 1300 cal years BP and 1500 cal years BP. Evidence at a low-resolution from previously proposed sea-level curves suggests that there was a late Holocene sea-level highstand around 1500 cal years BP (Miller *et al.*, 1995; Compton, 2001). The sea-level curve from the south-western coast (Compton, 2006) and to some extent the bedrock from the east coast along KwaZulu-Natal (Ramsay and Cooper, 2002), indicate that highstands have taken place between 3000 cal years BP and 1200 cal years BP. The *Zostera* facies recovered from Langebaan (Compton, 2001) and the pollen record at Verlorenvlei (Baxter, 1997) present evidence that the sea-level rose +0.5 m by 1300 cal years BP, along with oyster shells recovered from Langebaan (Flemming, 1997). This early period is known as the Little Climatic Optimum as it was a period of temperate warming which coincides with a period of sea-level rise in the Pacific Ocean (Nunn, 1998).

The two proposed sea-level reconstructions for Kariega show evidence of a lowstand following the highstand. According to Compton (2001), this lowstand of -0.5 m took place between 700 cal years BP and 400 cal years BP. This lowstand is evident on the south coast around 700 cal years BP as indicated from *in-situ* tree stumps exposed at Knysna Estuary during the low tide (Marker, 1997), although there is not a complete overlap of this lowstand at Kariega Estuary with previous sea-level curves for South Africa. However, the resolution of this particular proposed sea level is much higher compared to previous reconstructions. In Scenario One there is a lowstand of approximately -0.5 m between 500 cal years BP and 300 cal years BP. Between 700 cal years BP and 500 cal years BP there is a relatively small highstand taking

place around 0.1 m above the present sea level. Prior to this a minor highstand between 700 cal years BP and 1200 cal years BP there is a lowstand which is relatively stable around -0.3 m. According to available sea-level data for the Pacific Ocean (Nunn, 1998) the sea-level fell abruptly around 700 – 650 cal years BP.

The proposed sea-level curve for Scenario Two shows that around 300 cal years BP and 700 cal years BP the sea level was between -0.2m and -0.3m with a lowstand of -0.5 m taking place between 100 cal years BP. Scenario Two is not in accordance with the lowstand record from both Knysna and Langebaan of -0.5 m between 300 cal years BP to 700 cal years BP (Marker, 1997; Compton, 2001).

In Scenario Two there is no apparent increase but a gradual increase is evident since 1000 cal years BP. According to findings at Langebaan Lagoon (Compton, 2001) sea level has risen to its present levels since 400 cal years BP. In Scenario One there is evidence of steep fluctuations of the sea level and a noticeable feature in this construction is that the sea level is at its lowest during this period. Scenario One coincides with findings by Nunn (1998) in the Pacific, whereby the early part of the Little Ice Age approximately 550 cal years BP is marked by a rise in sea level a little above the present level approximately 430 cal years BP. This is evident in Scenario One whereby the sea level was a little above the present level between 600 cal years BP and 500 cal years BP and then drops to its lowest between 1500 cal years BP and the present. The Little Ice Age appears to be marked in the Pacific Ocean by slowly falling sea levels reaching a low of -0.9 m around 200 BP (Nunn, 1998).

Based on the above discussion and the literature review there is no rationale to suggest that one of the scenarios is a better interpretation of past sea-level changes that have occurred on the east coast of South Africa than the other.

The highstand evident at Galpin's salt marsh is supported by previous proposed sea levels and provides not only a better understanding of the east coast of South Africa but provides a high-resolution record of sea-level change on a regional scale. The lowstand which overlaps with the Little Ice Age suggests that climate change may have influenced the late Holocene sea-level fluctuations (Compton, 2001). Observed fluctuations in sea level during the late Holocene have been small compared to those during the early Holocene, in the order of $\pm 1-2$ m. Even though these fluctuations are relatively small they could potentially have an impact on the surrounding coastal environments.

Salt-marsh foraminifera show a strong correlation with elevation and salt-marsh vegetation. As they demonstrate a strong correlation with elevation they are considered to be accurate indicators of sea-level change. Foraminiferal assemblages found in Kariega Estuary have similar elevations to those found in the northern hemisphere and the west coast of South Africa.

A chronological framework was constructed for the core based on five AMS radiocarbon age determinations of both bulk sediment and shell fragment samples placing the record within the last 1500 years. For each foraminiferal sample, a transfer function calculated the former elevation at which the sample once existed thus enabling the use of these samples in conjunction with the age model to produce a relative sea-level reconstruction. The reconstructed curve indicates a 0.5 m sea-level highstand at 1500 cal years BP followed by a lowstand and several low amplitude fluctuations during a relatively stable period. Based on a problematic age model and the possibility of an erosional hiatus, two possible scenarios are presented. This interpretation of the reconstruction for east coast of South Africa is broadly in agreement with existing literature from South Africa's coastlines.

CHAPTER EIGHT: CONCLUSION

8.1 Introduction

This research investigated the applicability of modern and fossil foraminifera as a proxy for determining sea-level change. This was achieved by calculating former elevations at which the sample once existed with the use of a transfer function thus enabling a past sea-level curve to be reconstructed for the last 1500 years. To conclude this research, the initial aim and objectives of this thesis will be reviewed and future research directions will be explored.

i. To map estuarine vegetation and compare this with compositional changes in the surface foraminiferal assemblages

The foraminiferal and marsh vegetation zones were in part similar and overlapped to a certain extent, identifying three zones; high, middle and tidal mudflats. The transitional zones identified by the foraminiferal assemblages and marsh vegetation had similar trends to those seen at Langebaan, South Africa and salt marshes studied in the northern hemisphere.

ii. To determine the degree to which foraminiferal assemblages in the Kariega Estuary show evidence of vertical zonations

The foraminiferal assemblages from Kariega Estuary exhibit vertical zones that are similar to those found in salt marshes both in the northern hemisphere and along the west coast of South Africa. However, there are differences that relate specifically to elevation and vertical ranges of species. According to the foraminiferal assemblages the marsh is divided into three zones, whereby the agglutinated species are restricted to the upper reaches of the marsh and the calcareous species are more dominating towards the intertidal channel. The high marsh area is dominated by *Jadammina macrescens* with a small existence of *Trochammina inflata*. The middle marsh has a presence of both *Trochammina inflata* and higher number of *Miliammina fusca*. Calcareous species found in the mudflats consist of the species *Haynesina germanica*, *Ammonia batavus* and *Quinqueloculina spp.*

iii. To extract a series of sediment cores from the salt marsh and establish a stratigraphy

The stratigraphy of the salt marsh was determined through an extensive program of coring along Transect Two. Each core was described using Troels-Smith sediment classification system (Troels-Smith, 1955), based on these findings a suitable coring site was determined and larger sediment cores were extracted for foraminiferal analysis.

iv. To determine foraminiferal species composition and abundance at increments along a selected sediment core

Foraminifera occurred throughout the core but counts were low at certain depths particularly between 30 cm and 65 cm. Agglutinated foraminifera were most abundant in the upper reaches of the core and gradually decreased down the core. The base of the core was dominated by calcareous foraminiferal species. The species *Jadammina macrescens*, *Miliammina fusca*, *Trochammina inflata* and *Trochammina ochracea* were generally restricted to the upper 46 cm, whereas *Ammonia aberdoveyensis*, *Ammonia batavus* and *Haynesina germanica* were restricted to the lower half of the core.

v. To develop an age-depth model for this sediment core

The chronology for the core was determined through AMS dating on four bulk organic sediment and one shell sample. The basic age model for core two indicated two age reversals. The basal shell date is a clear outlier to all the bulk sediment dates. Results indicated two possible age reversals in the upper reaches of the core, which were best viewed as separate scenarios. Age reversals evident in the upper sections of the core suggest that there was either reworked organic matter from bioturbation, contamination by aquatic plant rootlets, erosion or deposition occurring as a result of wave action or river deposition. Both scenarios indicate a possible erosional hiatus at different depths within each scenario chronologies where sedimentation rates decreased and were viewed as unreliable. The chronology placed the record within the late Holocene period with a basal age of 1424-1331cal years BP.

vi. To develop transfer functions using ecological information contained within modern distributions of salt-marsh foraminifera

Multivariate regression analyses were used to quantify the relationship between surface assemblage data and mean sea-level in the contemporary environment. This produced a transfer function that could be applied to down-core assemblage data allowing historical sea levels to be inferred. For the purpose of this research both living and dead (total) surface assemblages were used to construct a training set. For each foraminiferal sample a transfer function calculated the former elevation at which the sample was formed at Kariega Estuary.

vii. To apply transfer function to fossil assemblage data to reconstruct a sea-level curve

A transfer function calculated the former elevation at which the fossil foraminifera samples once existed in conjunction with the age model producing a relative sea-level reconstruction. This has in return increased our understanding of ± 1 m sea-level changes during the late Holocene along the east coast of South Africa. Sea level was at least 0.5 m above the present day level between 1500 cal years BP and 1300 cal years BP, followed by a lowstand. After the highstand, sea levels remained below the present day level until 400 cal years BP at which it started to rise to present levels. Late Holocene sea levels from Kariega are similar to

those proposed previously from the west and south coast of South Africa and studies conducted elsewhere in the Pacific Ocean.

Conducting sea-level research through the use of intertidal foraminifera along the east coast of South Africa has inherent limitations, mainly the lack of suitable, relatively undisturbed and accessible salt marshes and the significant lack of information into foraminiferal assemblages found along the South African coastline. Many of South Africa's estuarine systems are not permanently open, therefore salt marshes are not receiving tidal fluctuations daily or many of the salt marshes are inhabited by mangroves. Kariega Estuary is one of the few permanently open systems with salt marshes that are relatively undisturbed. One of the major limitations was that the modern day training set was limited to Kariega, due to the lack of knowledge of modern foraminiferal assemblages and their relationship to past environments in South Africa.

Seasonal variations of living foraminiferal assemblages need to be taken into consideration when examining a dataset (Gehrels, 1994). For the purpose of this research live foraminiferal species were not used independently, they formed part of the total assemblages that the statistical analysis was based on. To investigate sedimentation history, particles size analysis downcore sediment should be considered.

Further research sites, if possible should be investigated along the east coast, thus establishing a greater modern day analogue for South Africa that could facilitate and strengthen further studies of sub-surface and core fossil assemblages to reconstruct sea-level changes. This will enable a greater understanding of how sea levels have changed at a regional scale. There is much scope for further sea-level reconstructions in South Africa. The potential for other salt marshes to provide insight into past sea levels through the use of foraminifera as a proxy is high and should be kept in mind if more suitable salt marshes or intertidal zones are identified along the east coast of South Africa.

8.2 Conclusion

Sea-level change along the east coast of South Africa and the southern hemisphere are under studied, there is a need for further research. The protection of coastal ecosystems is essential to sustaining human livelihoods and past sea-level records serve as a guide to what may take place when it comes to environment climate changes in the future. An understanding of the effects of climate changes and how the coastline will be impacted upon as a result of climate change is crucial. To gain this understanding and knowledge, further palaeoenvironmental research is needed to understand how sea levels have risen and impacted coastal systems and how they have adapted to such changes. Further reconstructions along the coastline could aid in identifying patterns and refining predictions of future sea-level fluctuations. Palaeoceanographic information at this resolution is necessary for understanding past sea levels and climatic changes. Such sea-level history is a reminder and warning that one cannot expect sea levels to be constant, as they have not been stable over

the last 1500 cal years BP and will continue to fluctuate into the future. Therefore this research is testimonial to the fact that salt-marsh foraminifera are precise indicators of relative sea-level change having reconstructed relative sea-level change for an estuarine system along the east coast of South Africa.

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Appendices

Appendix A: Troels-Smith stratigraphic description sheets for Transect Two

site: Campins Transect number: T2S1 Core name: 6m Dist. from site edge: 6m Date: 25-04-2011
 (water edge)

Depth	Physical Features				Upper Boundary	Mosses	Woody plants	Herbs	Woody detritus	Herb detritus	Fine detritus	Charcoal	Organic lake mud	Humus	Organosilicates	Carbonates	Iron oxides	Clay	Silt	Sand	Gravel	Comments	
	0-4	Dryness	Elasticity	Stratification																			Darkness
Lower	3	2	0	0	0									2				+					mid brown humus + clay
Upper	3	2	0	0	0	7.5YR3/1								1				3					humus, clay mid brown
	16	2	0	0	0	7.5YR3/2												1	1	2			lighter brown at beginning
	24	2	0	0	0	2.5Y5/3																	clay, silty sand 2 iron oxide stain.

Site: Scalpins Transect number: T2 S1 Date: 25-04-2011

Core name: 11m Dist. from site edge: 11m

Core name: 11m

Depth	Physical Features				Components (Total = 4)													Comments						
	Lower	Upper	0-4	Colour (Munsell)	Structure	Upper Boundary	Mosses	Woody plants	Herbs	Woody detritus	Herb detritus	Fine detritus	Charcoal	Organic lake mud	Humus	Organosilicates	Carbonates		Iron oxides	Clay	Silt	Sand	Gravel	
0	2	2	2	7.5YR 2.5/1		0	1	1							1				2					+ salt marsh flat mid brown clayed humus
2	12	2	0	7.5YR 3/1		1								1					3					mid brown stiff, humus clay & iron oxide stain
12	29	1	0	2.5Y 5/4		0													1	++	3			light brown to beige sand & clay + silty.
29	39	2	0	2.5Y 5/1		0													3	1				Brown grey silty clay + iron oxide stain
39	45	2	0	2.5Y 4/2		0													1	1	2			brown grey silty, clay & iron oxide stain
45	53	2	0	2.5Y 4/1		0													++	3	1			

Site: Carpins Transect number: Tz S1 Core name: 14m Dist. from site edge: 14m Date: 25-04-2011

Depth	Physical Features					Components (Total = 4)											Comments																			
	Upper	Lower	0-4				Upper Boundary	Structure	Colour (Munsell)	Darkness	Stratification	Elasticity	Dryness	Mosses	Woody plants	Herbs		Woody detritus	Herb detritus	Fine detritus	Charcoal	Organic lake mud	Humus	Organosilicates	Carbonates	Iron oxides	Clay	Silt	Sand	Gravel						
0	3	2	2	0	0	2	7.5YR 2.5/1				0				1											2								mid brown z fine roots, clay humus		
3	17	2	2	0	0	2	7.5YR 3/1				0				1											3	+							mid brown humus clay z fine silt layer		
17	34	1	0	0	2	2	2.5Y 5/3				0															1	+	3						clay sand z some silt, light brown		
34	45	2	0	0	2	2	2.5Y 4/1				0															4	+							light grey brown clay		
45	55	2	0	0	2	2	6YR 4/2				0															2	2							mid grey		
55	79	2	0	0	2	2	6.5Y 1.5/N				0															3	1							grey		

Site: Scarpins Transect number: T2 S1

Core name: 16m

Dist. from site edge: 16m

Date: 25-04-2011

Depth	Physical Features					Upper Boundary	Components (Total = 4)													Comments					
	Lower	Upper	0-4				Structure	Colour (Munsell)	Herbs	Woody plants	Woody detritus	Herb detritus	Fine detritus	Charcoal	Organic lake mud	Humus	Organosilicates	Carbonates	Iron oxides		Clay	Silt	Sand	Gravel	
			2	0	0	2	7.5YR 2.5/1		0	0	0	0	0	0	1					2					mid brown & fine roots clayey humus
	3	16	2	0	0	2	7.5YR 3/1		0	0	0	0	0	0	1				3	+					mid brown, humus clay, grey silt clay
	16	34	1	0	0	2	2.5Y 5/3		0	0	0	0	0	0					1	+	3				clayey sand with silty, light brown
	34	39	2	0	0	2	2.5Y 5/1		0	0	0	0	0	0	+				+						mid brown clay c iron oxide stain + silt
	39	42	2	0	0	2	2.5Y 5/1		0	0	0	0	0	0	3				3		1				Sand lens
	42	50	2	0	0	2	2.5Y 4/1		0	0	0	0	0	0	3				3	1					Silty clay grey brown
	50	71	2	0	0	2	grey 1.3/N		0	0	0	0	0	0	+				+	1					Silty clay, mid grey, brown oxide
	71	98	2	0	0	2	grey 1.3/N		0	0	0	0	0	0	2				2	1	1				mid grey, sandy silty, clay

Site: Crapins Transect number: T2 S1 Core name: 18m Dist. from site edge: 18m Date: 25-04-2011
 (cores 1-5)

Depth	Physical Features					Upper Boundary	Mosses	Woody plants	Herbs	Woody detritus	Herb detritus	Fine detritus	Charcoal	Organic lake mud	Humus	Organosilicates	Carbonates	Iron oxides	Clay	Silt	Sand	Gravel	Comments
	0-4	Darkness	Stratification	Elasticity	Dryness																		
0 2	2	0	0	2	75YR 2.5/1	0			1					1					2				mid brown, clayey humus
2 12	2	0	0	2	75YR 3/1	0								1					3	+			mid brown, humus clay & fine silt layer
12 32	2	0	0	2	75YR 3/1	0												+	4				grey, brown clay
32 39	2	0	0	2	10YR 4/2	1												+	1		3		mid beige brown
39 61	2	0	0	2	6.5Y 1 3/N	0												++	3	1			grey mid

Site: Celpins Transect number: T2 S1 Core name: 21m Dist. from site edge: 21m Date: 25-04-2011

Depth	Physical Features				Components (Total = 4)												Comments							
	Lower	Upper	0-4	Colour (Munsell)	Structure	Upper Boundary	Mosses	Woody plants	Herbs	Woody detritus	Herb detritus	Fine detritus	Charcoal	Organic lake mud	Humus	Organosilicates		Carbonates	Iron oxides	Clay	Silt	Sand	Gravel	
	0	3	2	7.5YR 2.5/1		0			1						1				2					mid brown ± fine roots, clay humus
	3	10	2	7.5YR 3/1		0									1				3	+				mid brown, humus clay ± fine silt layer
	10	34	2	10YR 4/1		0									1				4	+				mid, but grey clay silt + some organic detritus, iron oxide stain.
	34	56	2	grey 5/104		1													1	+	3			light grey clay sand ± some silt
	56	94	2	grey 1/3N		0													3	1				big broken shell material (gastropods) very rare mid to about grey silty clay.

Site: Grappins Transect number: T2 S1 Core name: 26m Dist. from site edge: 26m Date: 25-04-2011
 (Edge terrestrial)

Depth	Physical Features				Components (Total = 4)													Comments							
	0-4		Colour (Munsell)	Structure	Upper Boundary	Mosses	Woody plants	Herbs	Woody detritus	Herb detritus	Fine detritus	Charcoal	Organic lake mud	Humus	Organosilicates	Carbonates	Iron oxides		Clay	Silt	Sand	Gravel			
Upper																									
Lower	3	2	7.5 YR 3/1		0									1				3	+						
38	2	0	10 YR 4/1		0													4							
832	2	0	6.5 Y 4/N		0													3	1						
3271	2	0			0																				

mid brown, humus
 clay & silt layer
 grey brown clay
 & iron oxide stain
 From 50cm Gra
 (r) mid to
 dark grey silty
 clay. Ig (r) roue
 but present.

Appendix B: Troels-Smith stratigraphic description sheets for Core Two

Site: Galpins Transect number: T2 Core name: C2 Dist. from site edge: 8m Date: 15-06-2011

Depth	Physical Features				Components (Total = 4)													Comments						
	Lower	Upper	0-4	Colour (Munsell)	Structure	Upper Boundary	Mosses	Woody plants	Herbs	Woody detritus	Herb detritus	Fine detritus	Charcoal	Organic lake mud	Humus	Organosilicates	Carbonates		Iron oxides	Clay	Silt	Sand	Gravel	
0	2	2	2	7.5YR 3/2		0		1							2				1					brown to dark brown
2	4	2	0	7.5YR 3/1		0								1					3	+				light brown
4	7	2	0	7.5YR 3/1		0									1				3					Brown
7	12	2	0	7.5YR 3/1		0									+				4					molting begins
12	26	2	0	7.5YR 3/1		0									+				4	+				Silt becoming evident
26	30	2	0	2.5Y 4/2		0													1	3	1			Silt presence
30	38	2	0	2.5Y 4/3		0													+	1	2	1		clay less, sand layer
38	42	2	0	2.5Y 4/2		0													+	2	2			No sand
42	53	2	0	2.5Y 4/1		0													+	3	1			clay begins again
53	56	2	0	2.5Y 3/1		0													+	3	1	+		lots of molting
56	97	2	0	2.5Y 4/1		0													2	1	1	1		lots of shell fragments

Appendix C: Troels-Smith stratigraphic description sheets for Core Four

Site: Grampins Transect number: T2 Core name: C4 Date: 14-06-2011
 Dist. from site edge: 18m

Depth	Physical Features				Components (Total = 4)										Comments														
	0-4		Colour (Munsell)		Upper Boundary	Structure	Moisture	Darkness	Stratification	Elasticity	Dryness	Mosses	Woody plants	Herbs		Woody detritus	Herb detritus	Fine detritus	Charcoal	Organic lake mud	Humus	Organosilicates	Carbonates	Iron oxides	Clay	Silt	Sand	Gravel	
Lower	0	2	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	2	+		+						Int brown to dark Some clay
Upper	2	4	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	3		3	+				light brown layer	
	4	7	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	+	3		3	1				Dark brown	
	7	12	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	+	4		4					Beige brown	
	12	21	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	+	4		4					Dark & molting origins	
	21	25	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	+	4		4					Very Dark	
	25	32	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	+	2		2	2				Sand starts to become evident	
	32	41	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	+	1		1	2				is broken in places	
	41	47	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	+	2		2	2	1				Darker grey 1
	47	53	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	+	3		3	1				clay becoming more	
	53	58	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	+	2		2	1	1				Large shell fragments near bottom shell becomes more dense
	58	94	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0										

Appendix D: Taxonomic list of species

In this appendix we describe and illustrate the dominant agglutinated and calcareous foraminifera found in both surface and sub-surface sediments from Galpin's salt marsh, Kariega, South Africa. The agglutinated foraminifera characterize the vegetated marsh surface, while the calcareous species dominate the intertidal mudflats. Besides describing the features, a summary of their distribution is provided.

Suborder TEXTULARINA

Scherochorella moniliforme (Siddall, 1886):

Description: the test walls are agglutinated and pale brown in colour. They are made up of tiny detrital grains held together in an organic cement. *Scherochorella moniliforme* is cylindrical, consisting of a single column of chambers, with a terminal aperture (Murray, 1979; Horton and Edwards, 2006). Their average length is 0.8 mm (Murray, 1979).

Distribution: This is a sediment dwelling species found usually around the inner shelf. It is essentially a marine species however, is occasionally found in estuary mouths (Murray, 1979).

Cribrostomoides jeffreysii (Williamson, 1858):

Description: *Cribrostomoides jeffreysii* is an agglutinated species and therefore are composed of very small silt grains held together by organic cement. They appear smooth and pale brown in colour. The test is made up of six to seven chambers and is planispirally coiled. The average length it reaches is 0.5 mm (Murray, 1979).

Distribution: They are a marine species and are found along the inner shelf (Murray, 1979).

Miliammina fusca (Brady, 1879):

Description: Test walls are agglutinated composed of sand-sized detrital grains, held together by a pale brown in colour organic cement. However, morphology will vary slightly with grain size. The test is elongated and rounded in section. This is made up of many chambers coiled on a milioline plan. Aperture terminal with a tooth (Murray, 1979; Horton and Edwards, 2006). With an average length of 0.4 mm (Murray, 1979) (Appendix D, Plate 1, Taxa 3).

Distribution: *Miliammina fusca* is a sediment dwelling euryhaline species and is restricted to a narrow elevation zone within brackish lagoons, estuaries and tidal marshes (Murray, 1979; Horton and Edwards,

2006). This species was found in abundance in the intertidal channels at Kariega Estuary (personal observation).

Trochammina ochracea (Williamson, 1858):

Description: *Trochammina ochracea* have agglutinated walls which are made up of very fine detritus in organic cement. They appear reddish-brown in colour. The test is trochospirally coiled and very compressed with a small periphery. The spiraled side structure is slightly depressed and curved backwards. They usually have a maximum diameter of 0.2mm (Murray, 1979).

Distribution: *Trochammina ochracea* are an inner shelf species and spend their lives clinging onto pebbles and seaweed. (Murray, 1979).

Trochammina inflata (Montagu, 1808):

Description: Test is agglutinated and composed of very fine detrital grains with the outer layer being organic cement, brown in colour. The test is trochospirally coiled with globular chambers. The sutures are gently depressed on the spiral side, with the sutures deeply depressed between the inflated chambers on the umbilical side. The aperture is a narrow slit with bordering lip. The average size is 0.4 mm in diameter (Murray, 1979; Horton and Edwards, 2006) (Appendix D, Plate 1, Taxa 2).

Distribution: *Trochammina inflata* is a sediment dwelling species, commonly found in higher elevation zones within salt marshes. They are also typically able to survive in environments that vary including periods of little available oxygen (Murray, 1979).

Jadammina macrescens (Brady, 1870) and *Balticammina pseudomacrescens*:

Description: Test agglutinated and made up of incredibly fine detrital grains held together by brown in colour organic cement. The walls are particularly thin and when wet are flexible, however when they are dry the chambers tend to collapse. The test is trochospirally coiled and compressed with a shallow umbilicus (Murray, 1979; Horton and Edwards, 2006). The sutures are slightly raised when the chambers collapse (Murray, 1979). The aperture is a slit at the bottom of the apertural face with areal pores spread across the face. The average diameter of *Jadammina macrescens* is 0.3 mm (Murray, 1979; Horton and Edwards, 2006) (Appendix D, Plate 1, Taxa 12).

Distribution: *Jadammina macrescens* is one of the most abundant species to be found. They are commonly found around the world in brackish marshes and are able to survive extreme environmental changes (Murray, 1979). At Kariega this species was recorded in abundance in the high marsh (person observation).

Textularia earlandi (Parker, 1952):

Description: The test is agglutinated and the chambers are biserially arranged and laterally compressed. The aperture found at the base of the last chamber and is a simple arch (Scott *et al.*, 2001) (Appendix D, Plate 1, Taxa 8).

Distribution: This species is found in most coastal environments; however they are usually not the dominant taxon.

Textularia gramen (d'Orbigny, 1846): – (Variant) (Appendix D, Plate 1, Taxa 9)

Suborder MILIOLINA

***Quinqueloculina* species** (d'Orbigny, 1826):

Description: The walls of the test are porcellaneous with a translucent to opaque appearance. The chambers are coiled on a quinqueloculine plan, with the aperture usually having a tooth (Murray, 1979; Horton and Edwards, 2006). *Quinqueloculina* usually has an average length of 0.3mm (Horton and Edwards, 2006) (Appendix D, Plate 1, Taxa 4-5).

Distribution: *Quinqueloculina* are sediment dwelling and colonize mouths of estuaries if the conditions are favorable. They are also commonly found in the inner shelf as well as middle/low marshes and tidal flats (Horton and Edwards, 2006).

Suborder ROTALINA

***Lagena* species** (d'Orbigny, 1839):

Description: *Lagena* species are calcitic, which are radially arranged crystallites and have very fine pores and appear transparent. The test is a single chamber, but is always circular in the cross section with an elongated neck. The aperture is found at the end of the neck. The test walls may appear smooth or embellished with ridges and grooves. They can grow up to an average of 0.3 – 0.6 mm (Murray, 1979).

Distribution: They are a marine species and reside in the muddy substrates, however postmortem are able to be transported during storms and deposited in estuarine muds (Murray, 1979).

Ammonia batavus (Hofker, 1951):

Description: *Ammonia batavus* have calcite tests with radially arranged crystallites and pores. They appear glassy and translucent and can appear brownish in colour. They generally have a flattened dorsal side with the periphery sub-angular to sub-rounded. It is common to see eight to nine chambers on the ventral side (Horton and Edwards, 2006) (Appendix D, Plate 1, Taxa 7).

Distribution: *Ammonia batavus* are commonly found in low marsh and tidal flat environments (Horton and Edwards, 2006; personal observation). They occasionally appear in high abundance (Horton and Edwards, 2006).

Ammonia aberdoveyensis (Haynes, 1973): This species differs to that of *batavus* in that it is lacking an umbilical boss. It also has eight to nine rather than six to eight chambers per spiral (Murray, 1979) (Appendix D, Plate 1, Taxa 11). In Kariega this species was found in low marsh and tidal flat environments (personal observation).

Ammonia tepida (Cushman, 1926): This species is similar to *A. aberdoveyensis*, however it has slightly fewer chamber per whorl (Murray, 1979).

Spirillina vivipara (Ehrenberg, 1843):

Description: *Spirillina vivipara* have calcite walls which behave optically as a single crystal. They appear glassy with pores. The test is planispirally coiled and compressed. The test is a long tubular chamber which is not divided up, and consists of proloculus. The terminal aperture is simple. The greatest diameter it can be is 0.1mm (Murray, 1979) (Appendix D, Plate 1, Taxa 1).

Distribution: *Spirillina vivipara* is a marine species found along the inner shelf. It will be found clinging to solid structures, such as pebbles and sea weed. They occasionally are found in estuary mouths (Murray, 1979).

Fissurina species (Reuss, 1850):

Description: generally the walls of *Fissurina* genus are calcitic of radially arranged crystallites. The walls also have very fine pores that are translucent. The single-chambered test is oval in shape. The aperture is a

lenticular slit in the periphery, with an entoselenian tube that extends into the chamber of the test. The ornament consists of peripheral thickening or translucent to opaque bands (Murray, 1979).

Distribution: The genus is commonly found along marine shelves on muddy substrates. However, it can be transported into estuarine systems and deposited in the mud as empty tests (Murray, 1979).

Haynesina germanica (Ehrenberg, 1840):

Description: *Haynesina germanica* has a calcitic test with radially arranged crystallites which is finely perforated. They appear glassy and are planispirally coiled with five to eleven chambers. The aperture consists of a row of pores at the base of the last chamber (Horton and Edwards, 2006) (Appendix D, Plate 1, Taxa 10).

Distribution: *Haynesina germanica* is a common marsh species. It is the most abundant calcareous species present. Even though they are found in middle and low marshes, the highest abundance of *Haynesina germanica* are found in tidal flats (Horton and Edwards, 2006; personal observation).

Appendix E: Dominant foraminiferal taxa identified for Kariega Estuary

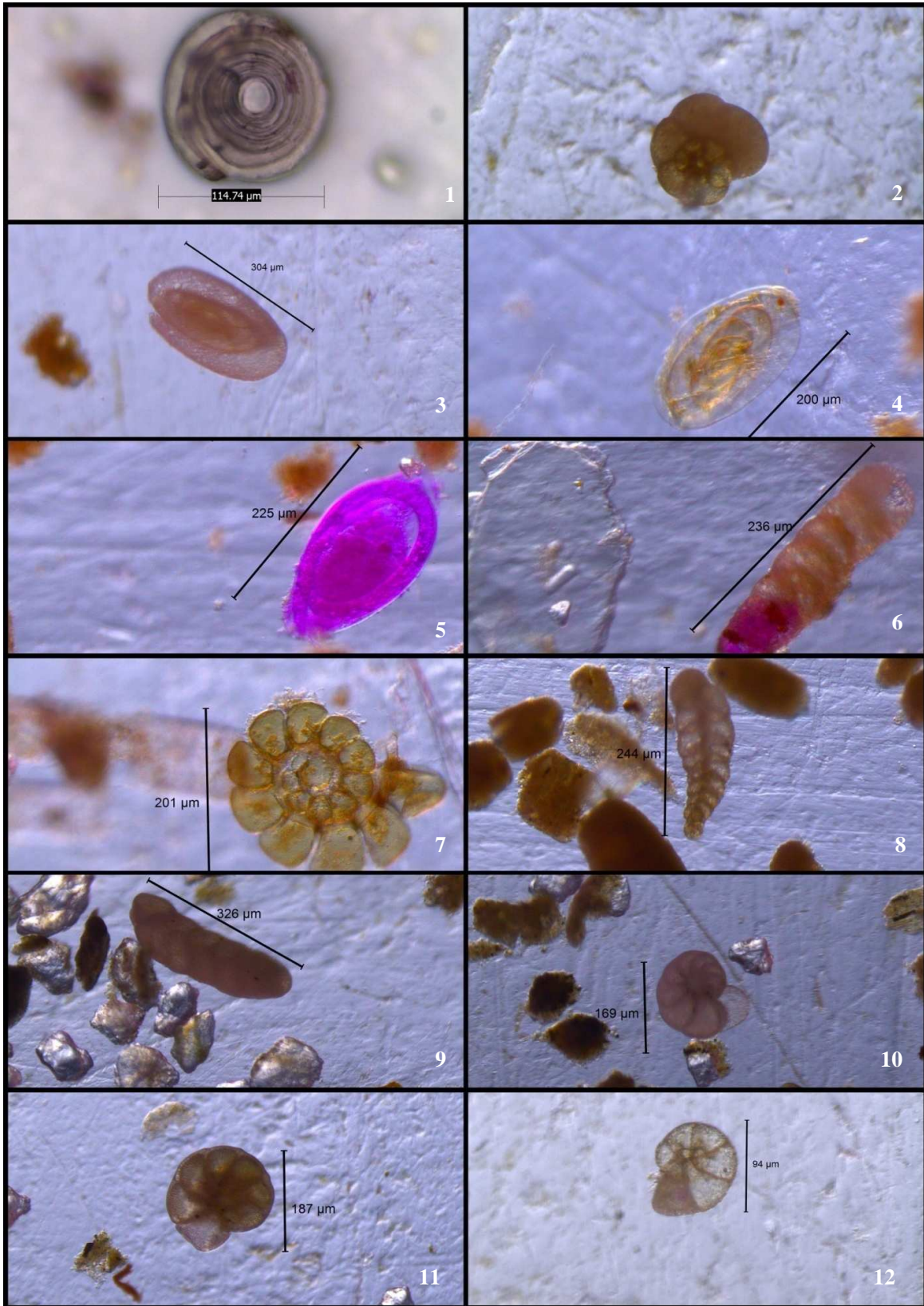


Plate 1. Taxa 1. *Spirillina vivipara* (Ehrenberg, 1843), living 2. *Trochammina inflata* (Montagu, 1808), living 3. *Miliammina fusca* (Brady, 1870), living 4. *Quinqueloculina spp*, non-living 5. *Quinqueloculina spp*, living, 6. *Scherochorella moniliforme* (Siddall, 1886), living 7. *Ammonia batavus* (Hofker, 1951), non-living 8. *Textularia earlandi* (Parker, 1952), non-living 9. *Textularia gramen* (d'Orbigny, 1846), living 10. *Haynesina germanica* (Ehrenberg, 1840), living 11. *Ammonia aberdoveyensis* (Haynes, 1973), living 12. *Jadammina macrescens* (Brady, 1870), living.

Appendix F: Transect One count data

Species Name	Status	0	1	2	4	6	8	10	18	28	38	52	61	62.5	63	64	Total
<i>Fissurina</i> spp.	Non-Living	0	0	0	0	0	2	0	0	1	5	0	0	1	0	0	9
	Living	0	0	0	0	0	10	3	1	30	72	0	1	4	4	0	125
<i>Jadammina macrescens</i>	Non-Living	0	0	1	11	17	139	165	120	173	70	73	33	18	11	10	841
	Living	0	0	0	0	17	46	81	73	60	49	55	19	8	17	6	431
<i>Miliammina fusca</i>	Non-Living	0	0	0	3	1	2	29	7	1	26	72	46	107	106	108	508
	Living	0	0	0	4	10	1	17	0	1	5	14	71	42	101	85	351
<i>Quinqueloculina</i> spp.	Non-Living	0	0	0	0	0	0	34	1	0	0	14	22	25	17	13	126
	Living	0	0	0	0	0	0	28	2	0	0	7	22	42	30	34	165
<i>Scherochorella moniliforme</i>	Non-Living	0	0	0	0	0	6	15	6	19	0	2	13	2	0	4	67
	Living	0	0	0	0	0	0	6	5	3	0	0	0	3	0	0	17
<i>Spirillina vivipara</i>	Non-Living	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	1
	Living	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Trochammina inflata</i>	Non-Living	0	0	0	0	10	54	6	19	58	11	21	39	35	8	35	296
	living	0	0	1	0	26	33	4	32	31	15	20	38	12	4	20	236
<i>Trochammina ochracea</i>	Non-Living	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
	Living	0	0	0	0	1	0	0	0	0	0	0	0	0	0	0	1
Unknown	Non-Living	0	0	0	0	0	1	0	2	0	0	0	0	1	1	0	5
	Living	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Total number of non-Living		0	0	1	14	28	204	249	155	252	112	182	153	189	143	171	1853
Total number of living		0	0	1	4	54	90	139	113	125	141	96	151	111	156	145	1326
Total number of sample species		0	0	2	18	82	294	388	268	377	253	278	304	300	299	316	3179

In Transect One it was found that 42 % of species were living while 58 % were non-living.

Appendix G: Transect Two count data

Species Name	Status	0	1	1.5	2	3	4	5	6	7	13	15	17	21	26	28	31.5	32	Total
<i>Fissurina</i> spp.	Non-Living	0	0	0	0	0	0	0	0	0	0	0	0	2	4	0	0	0	6
	Living	0	0	0	0	0	0	2	0	0	0	2	0	6	2	0	0	0	12
<i>Jadammina macrescens</i>	Non-Living	11	13	114	28	111	81	90	137	117	104	116	55	159	45	40	20	21	1262
	Living	2	1	56	75	82	17	30	26	21	23	41	34	19	37	19	22	41	546
<i>Miliammina fusca</i>	Non-Living	0	1	27	22	24	20	13	12	24	22	17	43	20	54	52	74	56	481
	Living	0	4	1	1	39	4	8	3	12	20	22	42	10	68	47	78	109	468
Quinqueloculina spp.	Non-Living	0	0	0	0	0	0	0	0	0	0	7	0	0	9	4	6	6	32
	Living	0	0	0	0	0	0	2	0	0	0	8	0	1	4	10	21	20	66
Scherchorella moniliforme	Non-Living	0	0	0	0	0	0	7	12	14	20	11	23	21	16	18	0	0	142
	Living	0	0	0	0	0	0	2	4	1	0	8	8	1	9	0	0	4	37
Spirillina vivipara	Non-Living	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	0	1
	Living	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	2	0	2
<i>Trochammina inflata</i>	Non-Living	1	0	13	20	32	89	90	66	60	73	39	70	86	30	48	24	8	749
	living	0	0	1	29	33	31	25	17	18	50	22	30	18	30	25	19	16	364
<i>Trochammina ochracea</i>	Non-Living	0	0	0	0	1	20	3	7	14	5	7	6	3	7	16	3	1	93
	Living	0	0	0	0	0	4	4	2	0	4	1	6	4	6	9	3	2	45
Unknown	Non-Living	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
	Living	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Total number of non-Living		12	14	154	70	168	210	203	234	229	224	197	197	291	165	178	128	92	2766
Total number of living		2	5	58	105	154	56	73	52	52	97	104	120	59	156	110	145	192	1540
Total number pf sample species		14	19	212	175	322	266	276	286	281	321	301	317	350	321	288	273	284	4306

In Transect Two it was found that 36 % of species were living while 64 % were non-living.

Appendix H: Transect Three count data

Species Name	Status	0	2	4	4.5	5	5.5	7	Total
<i>Ammonia aberdoveyensis</i>	Non-Living	0	5	0	2	0	0	19	26
	Living	0	0	0	0	0	0	0	0
<i>Ammonia batavus</i>	Non-Living	0	83	0	1	0	4	4	92
	Living	0	16	0	0	0	2	0	18
<i>Fissurina spp.</i>	Non-Living	0	0	0	0	0	0	0	0
	Living	0	0	0	0	0	0	0	0
<i>Haynesina germanica</i>	Non-Living	0	0	0	5	0	0	0	5
	Living	0	0	0	0	0	0	0	0
<i>Jadammina macrescens</i>	Non-Living	69	37	17	22	70	54	61	330
	Living	15	1	13	18	11	11	21	90
<i>Miliammina fusca</i>	Non-Living	33	8	86	71	64	70	76	408
	Living	120	14	134	135	106	116	93	718
<i>Quinqueloculina spp.</i>	Non-Living	2	0	2	0	1	4	2	11
	Living	12	0	6	0	1	3	10	32
<i>Scherochorella moniliforme</i>	Non-Living	10	0	2	5	0	6	0	23
	Living	0	0	2	0	0	1	0	3
<i>Spirillina vivipara</i>	Non-Living	0	0	3	0	0	0	0	3
	Living	0	0	0	2	0	0	0	2
<i>Textularia earlandi</i>	Non-Living	0	0	0	1	0	0	3	4
	Living	0	0	0	0	0	0	0	0
<i>Textularia gramen</i>	Non-Living	0	0	0	0	0	0	1	1
	Living	0	0	0	0	0	0	0	0
<i>Trochammina inflata</i>	Non-Living	25	0	7	11	16	6	4	69
	living	13	2	9	9	0	0	2	35
<i>Trochammina ochracea</i>	Non-Living	4	0	0	0	0	0	0	4
	Living	1	0	0	0	0	4	0	5
Unknown	Non-Living	0	0	0	0	0	0	8	8
	Living	0	0	0	0	0	0	0	0
Total number of non-Living		143	133	117	118	151	144	178	984
Total number of living		161	33	164	164	118	137	126	903
Total number pf sample species		304	166	281	282	269	281	304	1887

In Transect Three it was found that 48 % of species were living while 52 % were non-living.

Appendix I: Core Two count data

Count table of individual foraminiferal species found down the core, 0.5 cm - 48 cm

Species Name	0.5	2	4	6	8	10	12	14	16	18	20	22	24	26	28	30	32	34	36	38	40	42	44	46	48
<i>Ammonia aberdoveyensis</i>	0	0	0	0	0	0	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Ammonia batavus</i>	0	0	0	0	0	0	0	0	0	0	0	2	6	0	0	0	0	0	0	0	18	0	1	0	0
<i>Ammonia tepida</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Cribromoides jeffreysii</i>	0	0	14	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Fissurina spp.</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Haynesina germanica</i>	0	0	0	0	0	0	0	0	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Jadammina macrescens</i>	187	231	172	153	128	194	231	163	178	29	132	206	140	17	17	10	4	13	22	4	74	19	101	15	28
<i>Lagena Sp.</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Miliammina fusca</i>	56	13	22	25	74	23	14	56	31	0	26	22	24	3	0	0	1	4	0	0	0	4	21	0	13
<i>Quinqueloculina spp.</i>	0	0	0	4	0	0	0	5	2	0	5	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Scherochorella moniliforme</i>	7	16	13	22	0	33	12	0	4	0	0	57	0	0	0	0	0	0	0	0	0	0	1	0	0
<i>Scherochorella moniliforme</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Spirillina vivipara</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Textularia gramen</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Trochammina inflata</i>	14	15	29	28	26	34	25	18	27	4	34	0	29	0	5	1	0	2	1	0	2	0	2	0	0
<i>Trochammina ochracea</i>	17	17	25	50	44	31	30	18	21	0	56	0	77	0	0	0	0	5	0	0	0	0	0	0	0
Unknown	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Total number of sample species	281	292	275	282	272	315	313	260	265	33	253	287	276	20	22	11	5	24	23	4	94	23	126	15	41

Count table of individual foraminiferal species found down the core, 50 cm - 94 cm

Species Name	50	52	54	56	58	60	62	64	66	68	70	72	74	76	78	80	82	84	86	88	90	92	94	Total	
<i>Ammonia aberdoveyensis</i>	0	11	0	0	0	0	0	0	107	139	125	116	175	167	168	218	144	191	148	106	165	193	189	2363	
<i>Ammonia batavus</i>	0	0	134	76	87	226	111	0	0	5	0	2	6	4	0	15	0	0	0	4	5	0	1	703	
<i>Ammonia tepida</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	12	0	12	
<i>Cribromoides jeffreysii</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	14
<i>Fissurina spp.</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
<i>Haynesina germanica</i>	0	0	0	0	0	0	0	0	132	70	139	109	83	0	62	37	38	50	54	100	71	65	62	1074	
<i>Jadammina macrescens</i>	10	18	99	69	0	43	72	23	1	8	0	1	10	0	3	0	0	2	9	0	0	0	0	2836	
<i>Lagena Sp.</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	4	0	0	0	0	0	0	4	
<i>Miliammina fusca</i>	42	38	13	8	6	0	23	30	6	0	2	0	0	29	20	0	25	3	0	13	0	0	14	704	
<i>Quinqueloculina spp.</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	12	0	0	0	0	0	0	0	0	0	0	28
<i>Scherochorella moniliforme</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	165
<i>Scherochorella moniliforme</i>	0	0	0	0	0	0	0	0	0	31	0	6	0	15	0	6	33	0	0	23	0	0	0	114	
<i>Spirillina vivipara</i>	0	0	0	0	0	5	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	5
<i>Textularia gramen</i>	0	0	0	0	0	0	0	0	28	12	0	17	13	19	12	2	28	4	42	16	0	6	9	208	
<i>Trochammina inflata</i>	7	0	0	0	0	0	0	19	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	322
<i>Trochammina ochracea</i>	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	8	0	0	0	0	399	
Unknown	0	1	0	0	0	0	0	0	0	0	0	0	0	12	0	0	0	1	0	2	0	0	0	16	
Total number of sample species	59	68	246	153	93	274	206	72	274	265	266	251	291	258	265	278	272	251	261	268	258	276	286	9003	

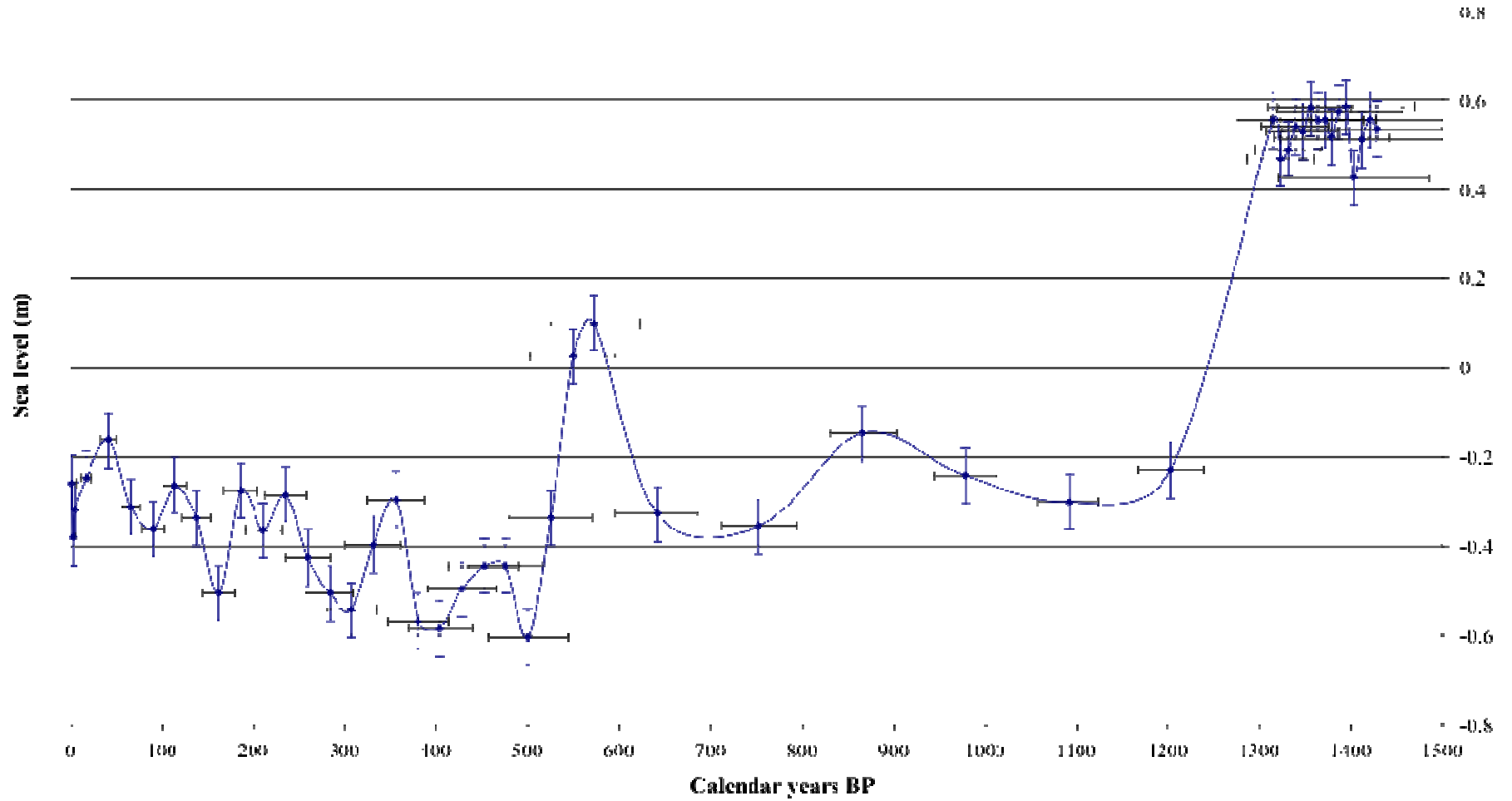
Appendix J: Results of the Detrended Canonical Correspondence Analysis

Axes	1	2	3	4	Total inertia
Eigenvalues	0.592	0.195	0.013	0.002	1.505
Lengths of gradient	2.816	2.736	2.494	2.493	
Cumulative percentage variance of species data	39.4	52.3	53.1	53.3	
Sum of all eigenvalues					1.505

Appendix K: Statistics summary of the performance of Weighted Averaging (WA) for the foraminiferal assemblages from the Galpin's Salt Marsh.

Code	Model performance			Bootstrapping		
	RMSE	r^2	Maximum bias	r^2_{boot}	Maximum bias _{boot}	RMSEP
WA_Inv (full data set)	0.156156	0.672425	0.222709	0.453209	0.240299	0.222693
WA_Cla (full data set)	0.19043	0.672425	0.16612	0.462612	0.253159	0.288695
WATOL_Inv (full data set)	0.150445	0.695945	0.205123	0.366669	0.252474	0.256246
WATOL_Cla (full data set)	0.18034	0.695945	0.147286	0.379331	0.30727	0.318598

Appendix L: Reconstruction of relative sea-level change for Scenario One with age error bars



Appendix M: Reconstruction of relative sea-level change for Scenario Two with age error bars

