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Late Holocene Relative Sea-Level Change and Climate in Southern Britain

Volume One :

Main Text, Tables, and References

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Robin James Edwards

Thesis submitted for the degree of Doctor of Philosophy.

Department of Geography,
University of Durham

May 1998

In memory of Great Grandma Payne,

For tigers and coral....

"Science, my boy, is made up of mistakes, but they are mistakes which it is useful to make, because they lead little by little to the truth."

Professor Lidenbrock

From *Journey to the Centre of the Earth* by Jules Verne (1864)

Declaration

This thesis is the result of my own work. Data from other authors contained herein are acknowledged at the appropriate point in the text.

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Robin Edwards,

Durham, May 1998.

Abstract

Late Holocene Sea-Level Change and Climate in Southern Britain

Robin J. Edwards, University of Durham, May 1998.

The late Holocene period has witnessed a number of widely recorded changes in climate of comparable magnitude to those predicted to occur in the next century as a consequence of human-induced global warming. Whilst late Holocene sea-level data may provide information on the future response of sea level, UK records of relative sea-level change from the last 2000 calendar years are conspicuous by their absence. This thesis redresses this imbalance by generating high resolution records of relative sea-level change from four study sites in southern Britain.

Vertically zoned modern saltmarsh foraminiferal distributions are employed *via* a transfer function to reconstruct past water level changes from fossil sequences. These intra-core lateral translations in depositional environment are placed in an altitudinal and temporal context through the development of AMS radiocarbon-dated sea-level index points.

Whilst the paucity of late Holocene organic sequences in the UK frustrates development of detailed radiocarbon-based chronologies, data collected in this study appear to indicate five phases of relative sea-level change in southern Britain during this period. Whilst age errors associated with both climate and sea-level data preclude the unequivocal identification of a climate signal in the sea-level record from southern Britain, the observed changes are consistent in timing, sequence, and magnitude with those expected to arise on the basis of known climate change during this period. Furthermore, ocean records imply that changes in Gulf Stream strength may play a critical role in translating atmospheric climate variability to changes in sea level. This suggests that the volumetrically derived, spatially homogeneous predictions of future relative sea rise employed by the IPCC may be inappropriate at the local to regional scale.

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Introduction

1.1 BACKGROUND

It is now widely accepted by the scientific community that we, the most recent and destructive version of the *Homo* genus, are altering the climate of the planet we inhabit. The 1995 report of the Intergovernmental Panel on Climate Change (IPCC) embodies the current opinion of an international scientific group and states that:

"*The background of evidence suggests that there is a discernible human influence on global climate.*" (Houghton *et al.*, 1996).

Instrumental records indicate that in the Northern Hemisphere, temperatures have generally risen since the 18th Century (Jones & Bradley, 1992a) (Figure 1.1). Furthermore, evidence from a range of proxy climate data suggests that the 20th Century is warmer than any period in the last 1000 years (Bradley & Jones, 1993; Briffa *et al.*, 1995). At the 1997 'Earth Summit' in Kyoto, Japan, representatives from countries around the globe met to discuss implementing measures to combat the spectre of human-induced climate change and its associated dangers. One of these threats is the possibility of an acceleration in the rate of sea-level rise, and the associated flooding and land loss that will result from this (Figure 1.2).

Climate and sea level were changing in sympathy long before human activities began to affect the natural order. The glacial and interglacial cycles of the **Quaternary Period** were characterised by temperature changes of *c.* 10 °C and sea-level fluctuations of *c.* 120 m (Warrick, 1993) (Figure 1.3). During the early **Holocene**, after the termination of the last glacial epoch, sea levels rose rapidly as terrestrial ice masses wasted releasing huge volumes of water into the oceans. It is generally believed that most of this ice melt had ceased by around 5000 to 6000 Cal. BP and that the smaller sea level variations after this time primarily reflect redistribution of water masses (e.g. Mörner, 1995). These more subtle, natural variations in climate and sea level that typify the **late Holocene** period are thought by some to provide good analogues for possible future change (Scott *et al.*, 1995a, 1995b; Scott & Collins, 1996).

During the last 3000 to 4000 Cal. years, there have been a number of widely documented climate variations. A general deterioration in climate around 2500 Cal. BP was inferred from plant macrofossils contained within Scandinavian peat bogs (Blytt, 1876; Sernander, 1908). This transition from the 'Sub-Boreal' to 'Sub-Atlantic' Blytt-Sernander climate periods was widely held as a sub-division of the Holocene Period until its oversimplification was recognised (Smith & Pilcher, 1973; Smith, 1981). More recent periods of climate change include the Medieval Warm Period that extended from *c.* 1400 Cal. BP to *c.* 700 Cal. BP, and the Little Ice Age, that encompassed the intervals from *c.* 400 Cal. BP to *c.* 100 Cal. BP (Lamb, 1979, 1984). As more detailed climate records for this period are developed, it is becoming increasingly apparent that these warm and cold epochs may not be globally or even regionally recorded (Jones & Bradley, 1992b) and, as with the Blytt-Sernander scheme before them, their timing varies considerably (Briffa & Schweingruber, 1992). Nevertheless, there is evidence that these changes were sufficiently widespread in the northern Hemisphere to leave their mark in ice cores (Meese *et al.*, 1994), ocean cores (Hass, 1996), and a variety of terrestrial proxies (e.g. Grove & Switsur, 1994).

The link between these small magnitude, short period fluctuations in climate and sea level has only recently begun to be explored in North America *via* high resolution multi-proxy investigations of saltmarsh environments (Varekamp *et al.*, 1992; Fletcher *et al.*, 1991, 1993; Nydick *et al.*, 1995). These intimate that local **relative sea-level records** may exhibit evidence of climate forcing (van de Plassche, 1991) or the influence of variations in oceanic circulation (Fletcher *et al.*, 1993). Such studies offer the means to examine the relationship between sea level and climate at centennial timescales which will be significant in facilitating accurate prediction of change in the 21st Century.

In contrast to North America, there is a paucity of sea-level index points in the UK from the late Holocene period in general, and the last 2000 Cal. years in particular (Figure 1.4). This means that the relationship between past climate change and sea level during this period is poorly understood. In this thesis, I present the first research explicitly seeking to document and explain late Holocene sea-level change in southern Britain and its relationship to climate. I use a combination of research methodologies previously employed in the UK in conjunction with techniques recently developed in North America, to produce a high resolution record of relative sea-level change that is compared with existing climate data. A foraminiferal-based transfer function is used to produce quantitative reconstructions of palaeo-mean tide level variations from fossil foraminiferal assemblages. The results of these analyses permit investigation of trans-Atlantic teleconnections of the relative sea-level changes recorded in Atlantic North America, and their links to climate.

I conclude that the records from southern Britain exhibit fluctuations in the rate of late Holocene relative sea-level rise consistent in timing and magnitude with those expected to arise from documented climate change during this period. Caution should be exercised in ascribing a climatic origin to these variations however, since the chronology of change lacks sufficient resolution and precision to unequivocally date individual climate or sea level events.

1.2 RESEARCH RATIONALE

1.2.1 Research Aims

This thesis has the following research aims:

1. To generate a high resolution record of late Holocene relative sea-level change for southern Britain;

2. To investigate the hypothesis that late Holocene relative sea-level change in southern Britain reflects documented variations in climate during this period;

3. To evaluate the applicability and efficacy of existing research techniques to the study of late Holocene sea-level change.

1.2.2 Research Approach

The research aims are addressed in the following manner:

1. Development of a methodology for studying centennial, low magnitude relative sea-level change (typically < 1 m) including the development and application of a foraminiferal-based transfer function;

2. Application of this methodology to determine late Holocene relative sea-level change from a number of sites in southern Britain;

3. Combination of site-specific records to provide reliable local sea-level histories;

4. Comparison of local scale sea-level records to identify any regional scale sea-level variation;

5. Comparison of sea-level records at all scales with documented changes in climate and reported relative sea-level variations in North America.

1.3 THESIS STRUCTURE

This thesis is presented in two parts:

PART I - Text and Tables

Chapter Two considers the mechanisms by which climate and sea level are related, and evaluates the existing evidence for late Holocene climate and sea-level change in the North Atlantic system. On this basis, a qualitative prediction of late Holocene sea-level change in southern Britain is developed which will be tested in *Chapter Seven*;

Chapter Three describes site selection criteria and the research methodology employed in this thesis;

Chapter Four introduces the study sites and presents a summary of results arising from lithostratigraphic, biostratigraphic, and chronostratigraphic investigation of their fossil deposits;

Chapter Five describes the contemporary foraminiferal distributions of the study marshes, and uses these data to develop a transfer function capable of making quantitative mean tide level reconstructions from fossil foraminiferal assemblages;

Chapter Six applies the foraminiferal-based transfer function developed in *Chapter Five* to the fossil assemblages presented in *Chapter Four* in order to reconstruct late Holocene relative sea-level change in southern Britain;

Chapter Seven compares the sea-level record produced in *Chapter Six* with the climate-based predictions made in *Chapter Two* to investigate the relationship between late Holocene sea-level change and climate in southern Britain. It also examines the evidence for trans-Atlantic teleconnections with the North American studies;

Chapter Eight presents the conclusions of this thesis, evaluates the success of the research techniques employed, makes recommendations for future work, and briefly considers the implications for future sea level rise predictions.

PART II - Figures and Appendices

Figures in this thesis are grouped according to chapter;

Appendix One presents the full lithostratigraphy for each site with accompanying figures;

Appendix Two describes the protocol used in the collection and processing of foraminifera, and presents the results for each site in tabular and diagrammatic form;

Appendix Three presents the results of the radiocarbon age estimations in association with the dating strategy. It then describes the generation of new sea-level index points and an evaluation of their reliability. In addition, existing radiocarbon dates used in this thesis are presented in tabular form, including the source publications;

Appendix Four presents the full output of the foraminiferal-based transfer function;

In addition, a glossary is included where all terms highlighted in **bold** throughout the thesis are defined.

Sea-Level Change and Climate

Jones & Hulme (1997) describe temperature as the fundamental measure of climate, and its variations are of particular significance to the study of climate-related sea level movements. As noted in Chapter One, the Medieval Warm Period and the Little Ice Age are examples of times when climate was generally warmer or colder than today. An important question however, is whether these changes were of great enough magnitude, duration, and spatial extent to have influenced past sea level. Equally important is the extent to which existing techniques used in sea-level research can detect such fluctuations.

In this chapter I consider the following issues:

- The mechanisms by which climate and sea level are related, and the spatial and temporal scales at which these links operate;
- The role played by North Atlantic oceanic and atmospheric circulation in influencing climate;
- The evidence for late Holocene climate change and its potential influence on sea level;
- The techniques used to investigate changes in sea level and their limitations;
- The existing evidence for late Holocene sea-level change in the North Atlantic region.

Climate and sea-level data are not presented in a consistent manner in the literature and chronologies are frequently defined using the radiocarbon timescale (^{14}C yr. BP), calibrated years BP where present is taken as AD 1950 (Cal. BP), calendar years

(AD\BC), and archaeological periods (e.g. Bronze Age). To facilitate comparison of these various data, all chronologies constructed in this thesis are presented in Cal. BP. Details concerning the calibration of radiocarbon age estimates are provided in Appendix Three.

2.1 Scales and Mechanisms of Change

The macroscale relationship between climate and sea level is well established. Figure 1.3 shows globally averaged changes in temperature over the last 150 000 years, and the associated variations in global sea level. As temperature falls, water is removed from the oceans and stored on land in the form of ice sheets and glaciers, resulting in a **glacio-eustatic** lowering of sea level. This is concomitant with a reduction in temperature and volume of the ocean waters, and this **steric** effect serves to further lower sea level. Upon deglaciation, these processes are reversed and sea level rises, as was the case at the end of the **Devensian** glaciation (e.g. Fairbanks, 1989).

The relationship between climate and sea level during the late Holocene is less easy to quantify. Atmospheric and oceanic circulation are important agents of sea-level change at the multi-decadal to centennial timescale, and this introduces a significant element of temporal and spatial variability. As a consequence of this, it is no longer appropriate to consider global changes in single variables such as temperature or sea level; instead regional to local scale fluctuations must be evaluated. The importance of regional to local scale sea-level records is particularly note-worthy given that the earth's crust is still experiencing the effects of the Devensian glaciation and its spatially heterogeneous distribution of terrestrial ice. This legacy of differing **glacio-isostatic** movements is further complicated by **hydro-isostatic** and **sedimento-isostatic** processes. These factors are influential in producing a wide variety of **relative sea-level** histories from around the UK (Figure 2.1).

A final complication associated with, and perhaps instrumental in, changing oceanic circulation is the **equipotential surface** of the oceans, termed the **geoid** (Mörner, 1976). The geoid imparts vertical relief to the ocean surface even before the influence of waves and currents is considered, and means that the altitude of sea level determined by geodetic methods varies around the globe (Figure 2.2). This led Mörner (1976) to speculate that any variation in the earth's equipotential surface would instil a regional change in sea level, a process he termed **geoidal-eustatic** change.

Whilst it is evident that relative sea-level change must be approached *via* local records, the scale at which climate must be considered is less obvious. Modern climate is spatially variable and, as the quality and distribution of palaeo-climate records increases, there is growing evidence that past climates were equally inconstant (Grove, 1988). Whilst relative sea-level records are constructed at the local scale, their climatic component (**eustatic** changes in ocean level) is regional to global in extent. It is therefore important to compare sea-level records with a variety of climate data collected from a wide geographic area.

2.2 ATMOSPHERIC AND OCEANIC CIRCULATION

In this section, the modern relationship between oceanic and atmospheric circulation and its influence on the present climate of northwest Europe is described. This review concentrates on identification of the principal mechanisms most likely to cause climate change.

2.2.1 The Gulf Stream

The climate of the UK is mild in comparison with areas at similar latitudes on the western margin of the North Atlantic. This situation arises as a result of the northward flowing Gulf Stream and its extension, the North Atlantic Drift (Figure 2.3), which are responsible for the transport of warm equatorial water to the west coast of the UK (Klein Tank & Können, 1997). The effect of this northerly heat transport is apparent when comparing sea surface temperatures (SST) on either side of the Atlantic. In the west, the difference in SST between Florida and Labrador is around 25 °C, whilst between the same latitudes in the east (between North Africa and Scotland), the difference is only 10 °C (Pickard & Emery, 1990). This northward heat transport means that the northern North Atlantic is around 4 °C warmer than comparable regions in the Pacific (Levitus, 1994). Intuitively therefore, the climate of the UK and northwest Europe will be particularly sensitive to variations in the strength of the Gulf Stream and North Atlantic Drift.

Variability of the Gulf Stream has added significance since it forms an integral part of the 'Great Ocean Conveyor', a density driven **thermohaline circulation** implicated in major climate change (Broecker *et al.*, 1985; Broecker & Denton, 1989, 1990; Broecker, 1991). Convective overturning of water in the Labrador, Greenland, and Norwegian Seas results in a subsurface outflow of cold, dense water termed North Atlantic Deep Water (NADW). This deep, southward outflow requires a surface, northward inflow, and is thought to be one of the reasons why the Gulf Stream extends much further north, and transports so much more water in comparison to the western boundary current in the Pacific (Rossby, 1996). If this thermohaline circulation were to weaken or completely shut down, the Gulf Stream and North Atlantic Drift would no longer bathe the British Isles in warm water, and the accompanying reduction in polar heat transport would result in dramatic high latitude cooling. In addition, the act of deepwater formation releases heat equivalent to around 30% of the yearly direct input of solar energy to the surface of the North Atlantic (Broecker & Denton, 1990), and is therefore a powerful climate forcing in its own right.

Toward the end of the last glacial period, during a time of ameliorating climate, there was an abrupt return to cold glacial conditions termed the Younger Dryas event. Evidence for this is found in ocean sediments (e.g. Ruddiman & McIntyre, 1981), ice core data (e.g. Dansgaard *et al.*, 1989), and terrestrial records (e.g. Watts, 1980). Boyle & Keigwin (1987) suggest that ocean foraminifera indicate a dramatic reduction in NADW formation at this time, which was most likely caused by a high latitude freshening of surface waters instigated by huge meltwater inputs from ablating glacial ice (Broecker *et al.*, 1985).

In Chapter One it is noted that during the late Holocene period, ice melt was a less significant forcing of sea-level change than it was at the beginning of the Holocene. Nevertheless, it is conceivable that subtle variations in the salinity of the Greenland, Norwegian, and Labrador seas could produce significant changes in ocean circulation and, in turn, abrupt climate change *via* fluctuations in the thermohaline circulation (Rahmstorf, 1997). Many computer models have sought to investigate the effects of high latitude freshening on oceanic circulation (e.g. Rahmstorf, 1994, 1995; Manabe & Stouffer, 1995). The necessary simplifications associated with modelling such a complex system means that these simulations are not well suited to making quantitative predictions (Rahmstorf, 1997); rather their strength lies in identifying patterns and trends that give clues to the qualitative response of the system. These modelling studies are in general agreement that high latitude freshening does reduce convective overturning, resulting in a weakened thermohaline circulation and a reduction of SST across the North Atlantic (Rahmstorf, 1995, 1997; Manabe & Stouffer, 1995). The simulations also seem to suggest that the ocean conveyor can recover relatively rapidly, and that prolonged periods of change require large inputs of freshwater. Alternatively, smaller volumes of freshwater could be applied over prolonged periods, such as would result from increased precipitation (Weaver & Hughes, 1994).

Lehman & Keigwin (1992) suggested that variations in the strength of the thermohaline circulation could be triggered by relative changes in discharge from the Norwegian and Labrador Seas. Norwegian Sea outflow is colder and deeper than that of the Labrador Sea and consequently a reduction in Norwegian NADW output would lead to cooling in the North Atlantic. Furthermore, the results of ocean-atmosphere models suggest that decreases in outflow from the Norwegian Sea are concomitant with increases in discharge from the Labrador Sea (Rahmstorf, 1994). Recent studies of convective overturning in the Greenland and Labrador Seas reveal changes of a compensatory nature have taken place during the last few decades (Sy *et al.*, 1997; Dickson, 1997). Furthermore, these data indicate that the Sargasso Sea which is another important source of **mode water**, has experience variations in convective overturning that are in-phase with those noted in the Greenland Sea (Dickson, 1997). From this it follows that localised atmospheric fluctuations in critical areas such as the Greenland, Norwegian or Labrador Seas have the potential to cause widespread adjustments of the ocean system capable of influencing climate.

2.2.2 Atmospheric Pressure in the North Atlantic Region

The oceans are not the only factor directing the climate across northwest Europe: atmospheric circulation is also significant, and its influence is best summarised in terms of pressure systems since they control the strength and direction of winds that, in turn, determine temperature and precipitation.

In the upper **troposphere** of the mid-latitudes, atmospheric circulation is dominated by the **Ferrel westerlies** which flow around the polar high pressure in a 'circumpolar vortex' (Lamb, 1972). These upper winds arise in response to the temperature differences experienced between polar and equatorial regions and are at their strongest during the winter months. Surface features such as mountain ranges deflect the circumpolar flow and induce long 'Rossby' waves in the westerlies (Rossby 1939, 1941). Wave troughs formed by geographical features such as the Rocky Mountains, vary little in position throughout the year, but the secondary troughs such as that located over eastern Europe, migrate with changes in flow strength (Lamb, 1972). When velocities are high, Rossby Waves are few, flow is more zonal and located toward higher latitudes. In lower velocity situations, the number of Rossby waves increases and the flow experiences much larger meridional swings that can produce cellular pressure distributions (Namais, 1972).

The importance of upper atmospheric flow is that it exerts a strong influence on the pressure patterns experienced at the earth's surface, and therefore on climate. The three major features of surface pressure distribution that influence the climate of Europe are the Icelandic Low, the Azores High, and the Siberian Winter High (Barry & Chorley, 1995). The Icelandic Low is formed by the westerly passage of depressions that originated in the zone of cyclogenesis associated with the Rossby wave trough in the lee of the Rocky Mountains (Davies *et al.*, 1997). The Azores High is an extension of the of the permanent tropical high pressure over the Atlantic, whilst the Siberian Winter High owes its existence in part to flow in the westerlies, but is intensified by extensive winter snow cover (Barry & Chorley, 1995).

Walker (1924) noted that the Icelandic Low and Azores High strengthened and weakened simultaneously, and termed this phenomenon the North Atlantic Oscillation (NAO). The NAO is most pronounced during the Northern Hemisphere winter (Barnston & Livezey, 1987) and accounts for more than one-third of the total variance in the sea-level pressure field over the North Atlantic (Deser & Blackmon, 1993). During periods when the **North Atlantic Oscillation Index** is high, a large pressure difference exists between the Icelandic Low and the Azores High, and both pressure systems migrate northwards (Angell & Koshover, 1974). This situation is associated with strong westerlies, higher SSTs in the North Atlantic Drift, and relatively mild winters in northwest Europe (van Loon & Rogers, 1978; Rogers & van Loon, 1979; Kelly *et al.*, 1997). At the same time, Greenland experiences colder conditions and cyclones occur at higher latitude with greater intensities (Carleton 1988; Serreze *et al.*, 1997). Hurrell (1995) demonstrated that during phases of high NAO index, northern Europe and Scandinavia experiences more precipitation whilst there is a reduction in moisture reaching southern Europe and the Mediterranean.

Conversely, during periods of low NAO index the pressure systems become less intense and migrate southwards, westerlies weaken, SSTs fall, and the stronger meridional winds bring warmth to Greenland whilst northern Europe experience cold conditions. Periods of weakened westerlies are more common in early spring and summer, and the cellular pressure patterns can give rise to 'blocking anticyclones'. These high pressure cells can persist for several weeks to months and bring prolonged periods of stable weather. Whilst the precise position and timing of anticyclone development determines the type of weather experienced at the surface, blocking anticyclones over northern Europe have been invoked as causes of climate change during the late Holocene (Lamb, 1977). A recent example is the UK drought of 1975 to 1976, when a blocking anticyclone caused one of the driest 18 month periods ever recorded (Davies *et al.*, 1997).

2.2.3 Atmosphere-Ocean Interaction

The ocean and atmosphere interact though exchanges of heat, mass and momentum, but the nature of this linkage is extremely complex and imperfectly understood (Rahmstorf, 1995; McCartney, 1997). Winds and atmospheric pressure influence the strength and direction of oceanic currents and can also affect SST *via* **Ekman pumping**. Lamb & Johnson (1959) suggested that cyclonic wind stress across the northwest Atlantic would enhance the poleward flow of warm water in the North Atlantic Drift, elevating SSTs in the northeast Atlantic. Lamb (1972) speculated that such a change in SST would alter the thermal gradient across the North Atlantic, steering cyclones further north into the Greenland and Norwegian Seas whilst permitting the northerly expansion of the Azores High. The resultant situation would be similar to that observed during phases of high NAO index, bringing warm weather to northwestern Europe.

Hansen & Bezdek (1996) presented evidence that SST anomalies are advected along the path of the North Atlantic Drift and, since their thermal signature extends below the seasonal thermocline, they have the ability to persist for a number of years. Sutton & Allen (1997) identified coherent SST fluctuations propagating across the Atlantic from ship-board SST observations compiled between 1945 and 1989. They noted that variations in SST in the 'storm forming region' between Florida and Cape Hatteras, are closely correlated to changes in the North Atlantic sea-level pressure. In this way, the propagation of SST anomalies has been linked with the NAO, opening the possibility of predicting general weather patterns many years in advance (McCartney, 1997). Irrespective of whether climate is driven by oceanic or atmospheric forcing, it is clear that the oceans are potential mechanisms by which short-period atmospheric variations can be attenuated to slower, longer period climate change (Dickson, 1997).

2.3 THE MODERN RELATIONSHIP BETWEEN CLIMATE AND SEA LEVEL

Instrumental records of sea level and climate extend back over last 200 years and, whilst strongly biased to the Northern Hemisphere, provide information on their relationship at the decadal timescale. An important question is whether these short period records accurately reflect the interaction between sea level and climate at centennial to millennial timescales.

2.3.1 The Instrumental Record of Temperature Change

The 'Central England Temperature' record developed by Manley (1974), is the longest **homogeneous temperature series** of any region in the world, extending back to 1659 AD (Jones & Hulme, 1997). The record indicates that temperatures have risen during this period, with the 1980s and 1990s being around 0.8 °C warmer than the 1660s (Jones & Hulme, 1997). Jones & Bradley (1992a) combine data from 12 stations in the Northern Hemisphere to investigate temperature changes from 1700 AD (Figure 1.1). All records show a warming between 1851 and 1980, and suggest that from the mid 19th Century, there has been a rise in temperature of around 0.5 °C throughout the Northern (and indeed Southern) Hemisphere (Jones & Bradley, 1992a). Whilst the greatest rise in Northern Hemisphere temperatures occurred between 1920 and 1940, the Central England Temperature record shows that three of the six warmest years have occurred since 1989 (Jones & Hulme, 1997).

2.3.2 The Instrumental Record of Sea-Level Change

Tide gauges measure relative sea-level change, and so a method of filtering out land level movements is required if eustatic (climate driven) sea level changes are to be discriminated. The exact length of record required to achieve this is a moot point. Warrick & Oerlemans (1990) suggest a minimum record length of 15-20 years, whilst Pirazzoli (1986) and Douglas (1991) prefer the longer periods of 50 and 60 years respectively. Ultimately, a compromise has to be reached between selecting longer records and maintaining sufficient numbers to ensure representative spatial coverage.

There are also differences in the methods used to distil the eustatic signal from tide gauge records. Many early studies limited data screening to the removal of stations located in areas of known uplift, subsidence, or tectonic activity, and assumed that by using a sufficiently large number of tide gauge stations, crustal movements would be compensated for over the large spatial scales involved (e.g. Fairbridge, 1961; Lisitzin, 1974). Other studies attempted to filter out the isostatic component through the use of geophysical models, or by establishing the long term rate of change from geologically based sea-level reconstructions (e.g. Gornitz *et al.*, 1982; Gornitz & Lebedeff, 1987; Douglas, 1991).

Despite the different approaches employed, there is general agreement that ocean levels are rising at a rate of 0.5 to 3.0 mm a^{-1}, with most values in the range 1.0 to 2.0 mm a^{-1} (Douglas, 1991; Gornitz, 1993; Pirazzoli, 1996). It is tempting to view this similarity as giving added credence to the results of such studies, but Pirazzoli (1986) warns that an inherent bias exists within the tide gauge dataset. The majority of tide gauge stations are situated on continental shelves and margins that are liable to subsidence as a result of human activities, hydro- and sedimento-isostasy. The result is an artificially elevated calculated rate of eustatic sea level rise. Fairbridge (1987) echoes this concern, describing the database as "hopelessly flawed". Therefore, whilst it is clear that the rate of temperature rise has increased during the last 100 to 150 years, there is currently no conclusive evidence of an associated acceleration in the rate of relative sea-level rise (Woodworth, 1990; Douglas, 1992; Gornitz, 1995). This may be due to difficulties in separating the eustatic and isostatic components of the tide gauge records, or may reflect the slower operation of the ocean system and associated lag effects.

2.4 THE EVIDENCE FOR LATE HOLOCENE CLIMATE CHANGE

Computer model simulations of surface air temperature anomalies indicate that high frequency changes tend to have relatively small spatial scales and contribute little to the variability of global mean surface air temperature (Manabe & Stouffer, 1997). Conversely, at multi-decadal to centennial scales, surface air temperature anomalies observed across the continents are more representative of larger scale conditions. Therefore, whilst many climate records are characterised by high frequency interannual to decadal variations in temperature (e.g. Pfister, 1992), it is their multi-decadal to centennial signals that are most likely to exert an influence on sea level. In this section late Holocene changes in temperature and precipitation derived from a number of different sources are examined to identify general patterns of variability.

2.4.1 Temperature

Late Holocene temperature variations can be reconstructed from a range of sources including ice core data, tree ring records, periods of glacial advance or retreat, documentary and instrumental records. For the purposes of this study, records are selected that possess long time series and which exhibit some evidence of multi-decadal to centennial variability.

2.4.1.1 The Ice Core Record

Ice core records are commonly analysed for variations in oxygen isotopes ($\delta^{18}O$) since these are related to temperature (Dansgaard & Tauber, 1969). The relationship is frequently expressed in the form of a linear function with constants being derived from the modern correlations between temperature and isotopic composition at a variety of locations (Robin, 1983). Oxygen isotope values are influenced by factors other than temperature, including moisture source, transport pathways, altitude, latitude, seasonal effects and ice movement (Dansgaard *et al.*, 1969, 1973; Koerner & Russell, 1979). As a result of these limitations, some workers suggest that ice core records are best interpreted as showing relative changes in temperature (Meese *et al.*, 1994). Alternative approaches include attempting to quantify the influence of other factors *via* general circulation models (Charles *et al.*, 1994), or by using direct measurements of borehole temperature (Cuffey *et al.*, 1992). Bradley (1985) suggests that since many of the complicating variables can be considered constant in short period ice core records, direct temperature estimations during the last *c.* 1000 years are reasonable.

Analysis of the Camp Century ice core (north-west Greenland) indicates that a marked cooling in climate occurred between *c.* 3500 Cal. BP and 3000 Cal. BP, and that relatively cold conditions persisted until *c.* 2000 Cal. BP (Dansgaard *et al.*, 1984). The interval from *c.* 2000 Cal. BP to *c.* 800 Cal. BP appears to have been slightly milder than the preceding cold period, but remained cool nevertheless. A second, abrupt shift to very cold conditions is intimated by the low $\delta^{18}O$ values recorded between *c.* 800 Cal. BP to *c.* 100 Cal. BP (Dansgaard *et al.*, 1984).

Dansgaard *et al.* (1975) examined variations in $\delta^{18}O$ extending back to before 1350 Cal. BP from the Crête ice core from central Greenland (Figure 2.4). The record indicates that between *c.* 1350 Cal. BP and *c.* 750 Cal. BP conditions were relatively mild, but that after this point there was a transition to a phase of colder climate. Conditions appear to have been particularly severe from *c.* 600 Cal. BP to 450 Cal. BP, and from *c.* 200 Cal. BP to 50 Cal. BP. There is some evidence that these two cold phases were separated by a period of relative warmth although conditions still appear to have been cooler than the warmth that characterised the early portion of the record.

Meese *et al.* (1994) analysed the GISP-2 core (central Greenland) and used accumulation rate as an indicator of temperature change (Figure 2.5). These results closely mirror the changes identified from the $\delta^{18}O$ records of Camp Century and Crête. The period between 1325 and 800 Cal. BP appears to have been typified by mild conditions with a higher incidence of melt layers. Meese *et al.* (1994) proposed that this could be an early indication of the warm phase identified in Iceland around 1150 Cal. BP by Lamb (1977), which he attributed to the beginning of the Medieval Warm Period. After 800 Cal. BP, there was a dramatic reduction in accumulation, indicating a fall in temperature until 700 Cal. BP, after which temperatures oscillated significantly until *c.* 200 Cal. BP (Figure 2.5). Once again there appears to have been a phase of relative warmth between *c.* 400 Cal. BP and 250 Cal. BP when accumulation rates were only slightly below those in the early part of the record.

A record of summer melt from the Lomonosov Ice Cap, Spitzbergen, extends back to 550 Cal. BP, and reveals that most of the last 600 years have been typified by cool conditions relative to the last century (Tarussov, 1992). The climate appears to have been particularly cold between 400 Cal. BP and 200 Cal. BP, after which there is a definite warming to the present day. The record from Spitzbergen is important since it is located near the deepwater forming areas of the Greenland and Norwegian Seas.

2.4.1.2 Glacial Evidence

Glacier advance or retreat is driven by changes in mass balance that reflect variations in temperature and precipitation, modified by the physical properties of the glacier itself and the surrounding terrain (Grove, 1988). The timing of changes in glacier extent can be determined *via* lichenometric dating of moraines (e.g. Beschel, 1961) or by dendrochronological analysis of trees growing on them or damaged by their advance (e.g. Carrara & McGimsey, 1981). Alternatively, organic material underlain or overlain by glacial deposits can be radiocarbon dated (Grove, 1988). A full discussion of these various dating methods and their associated errors is given in Porter (1981) and Bradley (1985). The biggest problem with glacial evidence when examining changes at the multi-decadal to centennial scale is that advances and retreats reflect variations in both temperature, precipitation and local glaciological conditions. However, the fact that variations in glacial extent is one of the proposed mechanisms by which climate and sea-level may be linked means that this record is still important (Houghton *et al.*, 1996).

The Swiss Alps possess the most detailed history of glacial changes during the Holocene (Grove & Switsur, 1994). Patzelt (1974), identified three periods of widespread glacier advance during the late Holocene (Figure 2.6). The Löbben advance, which occurred between 3800 Cal. BP and 3300 Cal. BP in the eastern Alps, was bigger than any advance to have occurred in recent centuries. Glaciers that expanded during the succeeding Göschener I advance (*c.* 3000 Cal. BP to 2300 Cal. BP) probably remained in their advanced positions until 1900 Cal. BP. A third advance (Göschener II) is also suggested to have occurred around 1600 Cal. BP to 1200 Cal. BP, a proposition supported by evidence from Bezinge & Vivian (1976).

Data from Scandinavia are less abundant and more widely scattered than the Alpine record (Grove & Switsur, 1994) with much of the work being conducted by Karlén and co-workers (1973, 1976, 1979, 1995). Karlén (1973) noted that Holocene moraines from 53 systems in northern Sweden could be divided into four groups on the basis of age. In the late Holocene period, there is evidence for glacial advance around 2800 Cal. BP to 2200 Cal. BP, and a later set of moraines which Karlén (1973) attributed to the Little Ice Age. Karlén (1976) mentions intermediate advances *c.* 3300 Cal. BP, between 2000 Cal. BP and 1800 Cal. BP, *c.* 1600 Cal. BP, from 1300 Cal. BP to 1000 Cal. BP, and between 700 Cal. BP and 400 Cal. BP (Figure 2.6), although questions about the reliability of this lichenometric chronology have been raised (Innes, 1984). Later work in Sweden by Karlén *et al.* (1995), identified periods of glacier advance from postglacial lake sediments (Figure 2.6). General periods of advance occurred around 3300 Cal. BP to 3000 Cal. BP, 2600 Cal. BP, 2000 Cal. BP, 1700 Cal. BP, 1200 Cal. BP, and from 800 Cal. BP to the present, broadly supporting the earlier findings.

Studies of the Okstindan glacier on the Swedish Border, indicate it advanced between 3200 Cal. BP and 2600 Cal. BP, and again around 1900 Cal. BP and 1500 Cal. BP (Griffey, 1975, 1976), whilst there is evidence for a number of intervening, short-lived warm periods (Figure 2.6). Karlén (1979) in a study of the Svartisen, Okstindan, and Saltfjell areas of northern Norway, suggested glacial advances around 2800 Cal. BP, 1900 Cal. BP, 1500 to 1300 Cal. BP, 1100 Cal. BP, and throughout the last 600 years (Figure 2.6).

The Greenland ice sheet has particular significance since it is the largest ice mass in the northern Hemisphere (Kelly, 1980). Ice core data from the GISP-2 core has already been mentioned (Section 2.4.1.1) but records of advance and retreat are also available for the ice sheet's peripheral glaciers. Reviews by Weidick (1972) and Kelly (1980) suggested that for a long period between *c.* 6000 Cal. BP and *c.* 3000 Cal. BP, conditions were generally warmer than those currently experienced in the region and the ice sheet was diminished in extent. Between 3500 Cal. BP to 3000 Cal. BP, glacier advance appears to have commenced in response to the climate deterioration noted in the Camp Century ice core (Weidick *et al.*, 1990). This apparently culminated *c.* 2000 Cal. BP before the climate warmed slightly between *c.* 1500 Cal. BP and 1000 Cal. BP (Kelly, 1980). A return to cold conditions accompanied by glacier advance is recorded after 1000 Cal. BP (Fredskild, 1973; Kelly, 1980) which became particularly pronounced after *c.* 700 Cal.

BP to 400 Cal. BP (Weidick, 1972; Kelly, 1980). The last 100 years have witnessed the end of this 'Little Ice Age' period and the onset of glacier retreat (Weidick *et al.*, 1990).

In North America, glacier data are abundant on the western coast (the Rockies and Cascades), moving northward into Alaska (Brooks Range). Data are less abundant on the Atlantic coast, with much of the information being restricted to the Cumberland Peninsula, occupying the southernmost region of Baffin Island and extending into the Labrador Sea. Here, Miller (1973) suggested a number of glacial advances which are shown in Figure 2.8 along with the palynologically-based temperature curve of Nichols (1967). Davis (1985) noted a similar chronology of advances (Figure 2.6), whilst Andrews & Barnett (1979) recorded advances in the northern part of Baffin Island at 3100 Cal. BP to 2800 Cal. BP, 2100 Cal. BP, 1500 Cal. BP, 1000 Cal. BP and 750 Cal. BP, and suggested that these provide evidence for Island-wide, synchronous changes in glacier extent.

Röthlisberger (1986) compared the Holocene fluctuations of glaciers from both the Northern and Southern Hemispheres and suggested that they exhibited broadly synchronous changes (Figure 2.8). Inspection of Figure 2.8 reveals that during the late Holocene period a greater variety of glacier responses is apparent. Röthlisberger (1986) attributed these differences to variations in sample material, carbon content and from the greater body of data. The latter fact however, may indicate that apparent global synchroneity is simply an artefact of poor data coverage. Alternatively, the more detailed records from this period may reveal the modifying influence of other controls such as precipitation which is known to exhibit marked spatial variability. Indeed, Grove (1988) suggested that whilst global temperature was the dominant control on glacier extent, variations in precipitation could give rise to departures from the general pattern.

2.4.1.3 The Tree Ring Record

Tree ring data are routinely used to reconstruct temperature and precipitation (e.g. Graybill & Shiyatov, 1992; Briffa *et al.*, 1995). Factors other than climate variables also affect tree growth, such as site location and ecology, competition, interference or damage, and disturbance (Briffa & Atkinson, 1997). Composite records are therefore constructed using many trees from different sites to produce a more reliable chronology for the area or region under investigation. The compilation and interpretation of such chronologies is discussed in detail by Cook & Kairiukstis (1990).

Briffa *et al.* (1995) constructed a 1000 year record of mean summer temperature (May to September) for the northern Urals (Figure 2.9). The time series exhibits variations that persist for 100 to 200 years, upon which are superimposed decadal to annual scale fluctuations. In general terms, the interval from 1100 Cal. BP to 800 Cal. BP was cooler than the modern reference period (1951-1970). Between 800 Cal. BP and 500 Cal. BP, temperatures were similar to the present day, but after 500 Cal. BP there was a return to cold conditions which extended until the recent warming phase at the turn of the century.

Briffa *et al.* (1992) also reconstructed summer temperatures for Fennoscandia. These show significant annual to decadal scale variability making it difficult to discern any longer period changes. In general, temperatures were cooler between 1600 Cal. BP and *c.* 1100 Cal. BP, and then warmed slightly from 1100 Cal. BP to 800 Cal. BP. The following four centuries showed temperatures oscillating around those similar to today, before cooling quite dramatically around 400 Cal. BP. The last 400 years have seen a progressive warming trend, although this is most pronounced at the beginning of this century.

Serre-Bachet (1994) examined the record from a number of locations in southwestern Europe and the western Mediterranean. These reveal that from 1000 Cal. BP to 800 Cal. BP it was predominantly cold. These cold condition were succeeded by a phase of oscillating temperatures around the mean, with these fluctuations becoming increasingly large and frequent between 400 Cal. BP and 200 Cal. BP. The record then suggests warming after *c.* 1850 AD.

2.4.2 Precipitation

Variations in precipitation are both indicators of changing atmospheric circulation patterns (Section 2.2.2) and potential causes of oceanic circulation changes (Section 2.2.1). In the same way that some of the temperature proxies mentioned above are influenced by precipitation, climate records from lakes and bogs reveal an integrated signal of wet and/or cool, or warm and/or dry conditions.

Some of the most detailed records of changing precipitation have come from raised peat bogs or mires. Ombrotrophic bogs obtain their moisture from the atmosphere, and climate change results in variations in moisture supply. As the bog shifts between wet and dry phases, the plant species growing on its surface alter in sympathy. Changes in water table also cause variations in peat humification, manifest as darker or lighter accumulations. Across Europe many raised bogs show distinct transitions from dark, well-humified peat into overlying light, less humified accumulations reflecting changes in precipitation regime (Svensson, 1988). In southern Sweden these transitions, termed 'recurrence surfaces', show shifts to wetter conditions *c.* 3200 Cal. BP, 2600 Cal. BP, 1600 Cal. BP and 800 Cal. BP. A transition *c.* 2500 Cal. BP is also recorded in many British raised bogs (Godwin, 1975, 1981) and is associated with the shift to wetter and cooler conditions at the Sub-Boreal/Sub-Atlantic transition of the Blytt-Sernander climate period (see Chapter One).

Recent research indicates that the age of some of these recurrence surfaces varies by several hundred years (e.g. Blackford, 1993), and is probably influenced in part by local changes in climate and possibly by site specific factors such as mire hydrology (Moore, 1986, 1993). Nevertheless, the evidence for a general deterioration in climate *c.* 3000 Cal. BP remains compelling (e.g. Godwin, 1975). For example, in a study of Bolton Fell Moss, Cumbria, Barber (1981) developed a surface wetness curve reflecting changes in climate from *c.* 2880 Cal. BP (Figure 2.10). This demonstrates that the last 3000 years have been predominantly wet and/or cool with particularly pronounced periods between 700 Cal. BP and 500 Cal. BP, and 425 Cal. BP to 100 Cal. BP. The period between 1900 Cal. BP and 1400 Cal. BP reflects a prolonged period of drier and/or warmer climate, after which conditions deteriorate again. Blackford & Chambers (1991) studied five upland blanket mires ranging from western Ireland to North Yorkshire, and noted that all appeared to show a similar shift to wetter conditions around 1400 Cal. BP.

A similar sequence of events is recorded in Burnfoothill Moss, eastern Dumfriesshire, by Tipping (1995), who used a multi-proxy approach to identify wet or dry periods. Tipping (1995) noted a shift from wetter/cooler conditions between 3000 and 1900 Cal. BP to a warmer/drier climate between 1900 and 1200 Cal. BP. The remainder of the period indicates a return to predominantly wet/cool conditions, although some proxies appear to indicate a possible brief dry/warm period between 600 and 400 Cal. BP.

Anderson *et al.* (1998) using a combination of data from mires and lakes, noted a pronounced shift to wetter conditions between 3900 Cal. BP and 3500 Cal. BP. This is supported by evidence of rising lake levels in Scotland between 4400 Cal. BP and 3200 Cal. BP (Yu & Harrison, 1995).

Van Geel *et al.* (1996) noted an abrupt climate deterioration around 2850 to 2760 Cal. BP in the Netherlands, where many settlements were abandoned due to waterlogging of the soil. This period coincides with a sharp rise in the C^{14} content of the atmosphere, perhaps indicating a change in solar activity, the earth's magnetic field, or the exchange between carbon reservoirs (van Geel *et al.*, 1996). There is evidence that at this time a number of settlements were established around the coastal marshes of the Netherlands (Boersma, 1983), and that this may have been related to migration from the flooding river areas (Waterbolk 1959, 1966). This led van Geel *et al.* (1996) to postulate that exploitation of the saltmarshes was made possible because of a slowing in sea-level rise that was responsible for the 'Holland-VI' regression (Roeleveld, 1976; Griede, 1978). They further proposed that climate deterioration lead to thermal contraction of the ocean and/or a reduction in Gulf Stream strength, coupled with an increase in water stored on land in glaciers, fens, bogs, and as groundwater.

2.4.3 Changes in Ocean Temperature and Salinity

The previous sections have considered changing climate in terms of atmospheric variations detected on land. Of perhaps greater significance to the issue of climate and sea level are climate related variations in oceanic variables such as temperature and salinity. As well as providing information on the lag between atmospheric and oceanic change, these data can also provide evidence of variations in oceanic circulation. This type of information is commonly obtained by examining the lithofacies of ocean sediments and the foraminifera contained within them. In light of the central role the North Atlantic Conveyor plays in determining the climate of northwest Europe (Section 2.2.1), information on changing conditions in the areas most strongly related to its behaviour, namely the Greenland, Norwegian, Labrador, and Sargasso Seas, is of paramount interest.

Koç *et al.* (1993) investigated changes in the temperature and salinity of the Greenland, Iceland, and Norwegian Seas using diatoms. Their reconstructions indicate that there was a general cooling *c.* 3000 Cal. BP reflecting a reduction in the inflow of Atlantic waters. In the Norwegian basin, this temperature reduction may have been around 5 °C as the arctic front moved to its current position (Koç *et al.*, 1993).

In a study of planktonic foraminifera from the Norwegian Sea, Bauch & Weinelt (1997) identified a pronounced depletion in ^{13}C and $\delta^{18}O$ indicative of surface water cooling or increased ventilation around 3000 Cal. BP. Bauch & Weinelt (1997) suggested that the ^{13}C variation may indicate a more widespread stratification of the Norwegian Sea at this time induced by a freshening of upper surface water in the northern North Atlantic. A freshening at this time is consistent with palaeosalinity estimates from the Northeast Atlantic (Maslin *et al.*, 1995).

Hass (1996) examined lithostratigraphic and oxygen isotope evidence from the Skagerrak, situated on the other side of the Greenland-Norwegian system to Nansen Fjord (Figure 2.11). This region is sensitive to changes in the Jutland Current which is strengthened by increased westerly winds, and exports water into the Norwegian Sea *via* the Norwegian Coastal Current. Hass (1996) suggested that between *c.* 2400 Cal. BP and 1600 Cal. BP there is evidence for relatively warm conditions, but that these were replaced by a cold period extending from 1600 Cal. BP to 1300 Cal. BP. During this latter period, the sedimentology suggests an intensification of the currents, possibly in response to alterations in the atmospheric circulation. From 1300 Cal. BP to *c.* 600 Cal. BP, conditions seem to have been fairly constant, and Hass (1996) equated this with the Medieval Warm Period. After 600 Cal. BP, energy conditions increased within the Skagerrak, and there is evidence for repeated changes in current strength and water mass which persisted up to the beginning of the 20th century.

Jennings & Weiner (1996) used a combination of lithofacies and foraminiferal data to investigate changes in oceanography within Nansen Fjord, East Greenland. This site is well situated to identify changes in the strength of the East Greenland Current and the Irminger Current (Figure 2.11). The East Greenland Front occurs at the intersection of these two current systems (Johannessen, 1986), and its southerly migration brings Polar Water, sea ice, and lower temperatures to Iceland (Lamb, 1979). Jennings & Weiner (1996) suggested that between *c.* 1200 Cal. BP and 800 Cal. BP, conditions were relatively warm and stable, indicating that the Irminger Current (and therefore the Gulf

Stream/North Atlantic Drift) was relatively strong. After 800 Cal. BP, there is evidence for a shift to more variable conditions with periods dominated by cool Polar Water. These periodic shifts to cold conditions became more intense after *c.* 300 Cal. BP, with increasing evidence of polar ice around Iceland.

Lamb (1977) used a variety of historical sources to reconstruct the changing extent of sea ice around Iceland, relying heavily on the work of Koch (1945) (Figure 2.12). During the early days of Viking colonisation, between around 1020 Cal. BP to 1080 Cal. BP, sea ice is rarely mentioned in historical accounts. In fact records of sea ice are rare until *c.* 750 Cal. BP after which there appears to have been an increase in extent that finally forced sailing routes to be changed after 700 Cal. BP. From *c.* 540 Cal. BP onward there appears to have been no regular communication between Europe and any part of Greenland until around 230 Cal. BP, by which time the Norse colony had died out. Ogilvie (1992) has raised doubts about the quality of the data used by Lamb and suggested that prior to *c.* 350 Cal. BP there are insufficient data to provide anything more than a general indication of climate change. The more detailed record of Ogilivie (1992) shows that sea ice extent was at it greatest around 350 Cal. BP, 250 Cal. BP and increased again after *c.* 190 Cal. BP (Figure 2.12). Within this cold period, distinct but brief phases of warmth were apparent around 300 Cal. BP and from *c.* 210 Cal. BP to 230 Cal. BP.

Away from the northern seas, Keigwin (1996) examined changes in sedimentation and foraminiferal $\delta^{18}O$ from cores recovered beneath the northern Sargasso Sea. It is suggested that reductions in the abundance of carbonate in the sediments indicate an increase in the terrigenous flux transported by deep currents (Bacon & Rosholt, 1982; Suman & Bacon, 1989). Keigwin (1996) noted that these correlate well with $\delta^{18}O$ values from planktonic foraminifera, indicating that these deeper currents were more active during cold periods, perhaps reflecting an increase in the vigour of the Gulf Stream resulting from elevated levels of atmospheric storminess. This evidence from the Sargasso Sea indicates that there was an abrupt decline in SST between *c.* 5000 Cal. BP and *c.* 3500 Cal. BP, before a recovery to relative warm, quiescent conditions from *c.* 3000 to 2000 Cal. BP (Figure 2.13). These were interrupted a second time by a pulse of cold around 1500 Cal. BP, which Keigwin (1996) suggested precedes the Medieval Warm Period. A warming phase followed from *c.* 1500 Cal. BP to 900 Cal. BP after which the record exhibits a general cooling trend, although between 600 Cal. BP and 500 Cal. BP the $\delta^{18}O$ record indicates a brief pulse of warmth.

2.4.4 Storms

Whilst the temperature and precipitation records above provide a general picture of long term variations in climate, they do not give any indication of changes in the frequency of extreme events. The occurrence of storms is significant since they can result in major changes of coastal configuration and sedimentary character (Lamb, 1984b). Phases of enhanced storminess may result in the deposition of large sand layers or increased erosion of coastal deposits. These features serve to complicate interpretation of the stratigraphic sequences used in sea-level research and may remove portions of the record (Sections 2.5 & 3.1).

Reconstruction of storm frequencies through time is complicated by the changing availability of documentary evidence (Lamb, 1979). Nevertheless, there appears to have been a general increase in storminess associated with the deterioration in climate *c.* 800 Cal. BP noted in the terrestrial records presented above. A period of extensive dune formation in North West Germany, Belgium and the Netherlands indicates the existence of considerable bodies of blowing sand from *c.* 800 Cal. BP onward (Lamb, 1984b). There is a corpus of documentary evidence reporting the engulfing of land and settlements by blowing sand from *c.* 700 Cal. BP and damage or loss of life resulting from sea floods (Lamb ,1979). Whilst there is some indication that conditions may have improved briefly between *c.* 450 Cal. BP and 400 Cal. BP (Lamb, 1979), severe storms returned to characterise the following 200 years (Lamb 1984a, 1984b). Lamb (1979) speculated that the enhanced storminess that appears to have accompanied the deterioration in climate after 800 Cal. BP might have be caused by increased thermal gradients arising from a southward migration of the polar front.

2.4.5 Summary of Late Holocene Climate Data

Inspection of Figure 2.14 reveals that even the simplified climate records exhibit differing evidence for climate change during the last 5000 years. Additionally, records derived from the different proxies are not distributed uniformly through time. For example, tree ring records only become widely available after 1500 Cal. BP and consequently changes in climate may in part, reflect variations in data source. With these limitations in mind, a general trend in Northern Hemisphere climate is proposed on the basis of these data. There is evidence for deteriorating climate *c.* 3000 Cal. BP with a shift to wetter conditions and a period of glacier advance that is broadly consistent with the Sub-Boreal/Sub-Atlantic transition. This persists until *c.* 1900 Cal. BP when the glacier records show shorter periods of glacial advance, and the peat bog records suggest a warmer and drier phase persisting until *c.*1200 Cal. BP. A number of the terrestrial records indicate that this warmth continued in some areas until *c.* 800 Cal. BP, after which climate became much more variable, oscillating around a common value. Some evidence of a warmer phase around *c.* 400 Cal. BP to 300 Cal. BP is apparent in ice core, tree ring, and peat bog data. Finally, there is evidence from a number of proxy records supported by instrumental and documentary evidence, that the last 150 to 200 years have seen a general warming of the climate (Section 2.3).

A composite terrestrial temperature record is compiled from the data above, placing particular emphasis on variations indicated by a number of different climate proxies. This record is presented in Figure 2.15. along with a summary of the precipitation changes and the ocean records. The imprecision of the dating of transitions in these systems is further increased by the compilation of records into groups showing broadly consistent trends. This is unsurprising since there is evidence that individual areas have responded to climate change at different times (Jones & Bradley, 1992b). This results in a temporal smearing of transitions and it is therefore unwise to place too much emphasis on apparent synchroneity of change.

Perhaps the most reliable 'correlation' on the basis of timing is the compensatory changes occurring between the Greenland, Iceland, Norwegian Seas in the north, and the Sargasso Sea in the south at around 3000 Cal. BP. In Section 2.2.1, evidence is presented linking the Greenland, Iceland, Norwegian and Sargasso Seas together. It is proposed that increased stratification in the northern seas associated with a reduction in convective overturning, is linked to enhanced deepwater production in the Labrador Sea and an increase in the output of 18° mode water from the Sargasso Sea (Figure 2.16). Delworth *et al.* (1997) in a simulation of the Great Salinity Anomaly reported by Dickson *et al.* (1988), propose that negative SSS and SST anomalies indicative of a freshening in the arctic seas are transported *via* an enhanced East Greenland Current to the Labrador Sea, and that this is associated with warming off the coast of North America as a result of a reduction in the strength of the thermohaline circulation. The evidence of Bauch & Weinelt (1997) suggests that around 3000 Cal. BP there was an increase in the stratification of the Norwegian Sea, related to a fall in SST with evidence for an accompanying reduction in SSS (Maslin *et al.*, 1993). At the same time, Keigwin (1996) reports a potential reduction in the vigour of the Gulf Stream and a rise in temperature in the Sargasso Sea. From Section 2.2.1, a weakening of the Gulf Stream is expected to cause cooling in western Europe. It is therefore interesting to note that the Alpine glaciers started a major period of advance around 3000 Cal. BP, shortly followed by their Scandinavian counterparts, whilst evidence from the Netherlands suggests that there was a dramatic deterioration in climate around this time.

This relationship can be taken further in that there is a close correlation between the increased vigour of circulation in the Sargasso Sea *c.* 2000 Cal. BP and the onset of warmer and drier conditions indicated by the terrestrial temperature and precipitation records. Once again, this is entirely consistent with an increase in the strength of the Gulf Stream, and enhanced northerly transport of warm water. Unfortunately no data are available regarding stratification in the northern seas at this time, and data from the wind-sensitive Jutland Current are inconclusive.

The combined climate record indicates a transition to colder and wetter conditions between around 1200 Cal. BP and 800 Cal. BP, and evidence from Nansen Fjord suggests that around 800 Cal. BP there was a shift toward periods dominated by polar water indicative of an enhanced East Greenland Current. This corresponds well to a reduction in Gulf Stream strength and increase in temperature inferred from the Sargasso Sea. At the same time, the Skagerrak record shows a strong pulse in bottom current activity related to a change in atmospheric circulation. It is possible that these shifts in oceanic circulation are related to the southward migration of polar waters and sea ice that became more prevalent after 700 Cal. BP and ended communication with the Norse colonies of Greenland after 540 Cal. BP.

The most recent period exhibits an increase in variability both within individual records, and between data from different sources, hindering reliable correlation of records. Nevertheless, brief periods of change have been inferred from the oceanic and climate records. Whilst the last 600 years appear to have been predominantly cold there is some evidence for warming between 400 Cal. BP and 300 Cal. BP. Furthermore, an abrupt but pronounced change in temperature also occurred in the Sargasso Sea around 600 Cal. BP to 500 Cal. BP, perhaps indicating a variation in oceanic circulation at this time.

2.5 THE EVIDENCE FOR LATE HOLOCENE SEA-LEVEL CHANGE

Sea-level change can be measured directly using instruments such as tide gauges (e.g. Shennan & Woodworth, 1992) or satellite altimeters (e.g. Nerem, 1995); indirectly utilising palaeoenvironmental evidence (e.g. Devoy, 1979); or simulated using geophysical models (e.g. Lambeck, 1993). The indirect method of palaeoenvironmental reconstruction is the only viable approach to the study of late Holocene sea-level change since direct methods do not possess long enough records and geophysical models are too imprecise.

The palaeoenvironmental approach utilises physical features and biological organisms whose distributions are related to sea level in a consistent manner. Such phenomena include ecosystems like saltmarshes with their associated flora and fauna, or morphological features such as wave cut platforms and tidal notches. By establishing the vertical relationship between these **sea-level indicators** and the water levels which influenced their formation, it is possible to use preserved, relict features to constrain the position of former relative sea-level, and by dating them, produce a history of relative sea-level change. This palaeoenvironmental approach has the problem of determining whether observed variations are the result of relative sea-level change or a response to other forcing factors. For example, a transition from mudflat to saltmarsh may indicate a fall in relative sea-level, but could equally arise from an increased sedimentation rate resulting in marsh progradation. For this reason, a multi-proxy approach is strongly advocated, combining data from more than one site. In addition, the record of change will only consist of those variations that are registered by an indicator, and subsequently preserved. This will depend on a number of factors including the magnitude and rate of relative sea-level change, the sensitivity and resolution of the indicator used, and the successive environmental conditions experienced after its formation (Pirazzoli, 1996).

A high proportion of Holocene sea-level data are derived from analyses of saltmarsh sedimentary sequences. Saltmarshes exist at the interface between land and sea, and their vertical range is closely related to the tidal frame (Pethick, 1980, 1981; Allen, 1990a, 1990b, 1990c). Their usefulness as sea-level indicators is enhanced by a pronounced vertical zonation of flora and fauna, related to elevation above mean tidal level (Chapman, 1960; Scott & Medioli, 1980a; Gray, 1992). This vertical differentiation between saltmarsh subzones and adjacent freshwater or mudflat environments can be identified in fossil deposits, enabling the relationship of these sediments to former tidal levels to be quantified (termed the **indicative meaning** (van de Plassche, 1982, 1986; Shennan, 1986)). Saltmarshes are commonly located in sheltered, low energy environments and are therefore better suited to the accumulation of an unbroken record of change than more exposed, high energy settings such as sand dunes and gravel beaches. In addition, the rich floral assemblages of the high marsh environment provide an excellent source of organic material for radiocarbon dating.

The principal limitations of the saltmarsh system as a source of sea-level information derive from the fact that sea level is only one of a number of factors that control its development (Allen & Pye, 1992). Other important influences on marsh evolution are sediment supply, tidal regime, and wind/wave climate. It is therefore vital to distinguish between positive or negative **sea-level tendencies** which may arise from any one of these influences, and changes in depositional environment produced by vertical movements of sea level. Relative sea-level changes recognised in a number of marshes from different locations are more likely to be a response to a regional forcing such as crustal movements and eustatic sea level change (Shennan *et al.*, 1983). For this reason it is important to compile composite records of relative sea-level change from a number of sites.

The possible climatic significance of processes operating in the North Atlantic oceanic and atmospheric systems has been discussed above (Section 2.2). In the following sections, the evidence for variations in relative sea-level on both sides of the Atlantic is presented in an attempt to identify patterns that may provide information on possible changes in this oceanic or atmospheric circulation.

2.5.1 Late Holocene Relative Sea-Level Data from the North Atlantic East Coast

Relative sea-level change can be assessed in terms of vertical movements of water level or lateral shifts in coastline and depositional environment. In the following sections, evidence of relative sea-level change derived from these two viewpoints in discussed with particular reference to southern Britain, supplemented with European data from studies explicitly seeking to link climate and relative sea-level change.

2.5.1.1 Vertical Changes in Relative Sea-Level

The mid Holocene coastal sedimentary sequences of southern Britain are characterised by extensive peat beds intercalated with minerogenic deposits. Numerous investigations of similar buried coastal peat beds from around the UK and Europe have been conducted by Godwin (1939a, 1939b, 1940a, 1940b, 1943). On the basis of these data, Godwin (1940b, 1943) proposed a classic sequence of relative sea-level change for the mid to late Holocene period. Falling relative sea-levels characterised the Bronze Age (*c.* 3800 to 2800 Cal. BP) but gave way to a period of rising relative sea-level during the Iron Age to early Roman times, termed the 'Romano-British Transgression' (*c.* 2800 to 2000 Cal. BP). This was succeeded by a fall in relative sea-level after 2000 Cal. BP which was in turn replaced by a renewed rise in relative sea-level around *c.* 1200 Cal. BP.

The sequence of relative sea-level movements suggested by Godwin (1940b, 1943) is largely based upon the Fenland record and much of the data comes from deposits with no quantifiable relationship to a tide level (**indicative meaning**). For this reason the graph of relative sea-level change is more strictly interpreted as showing variations in coastal sedimentation. However, the large amount of supporting data collected from a wide variety of locations suggests that the general pattern of relative sea-level change can be used as a framework against which more precise sea-level information from southern Britain can be compared.

The work of Devoy (1979) is a classic early example of the modern methods employed to produce precise sea-level records. In a study of the Thames Estuary, Devoy (1979) presents a lithostratigraphy consisting of five main peat beds interdigitated with minerogenic sediments (Figures 2.17 & 2.18). This type of sequence is commonly used to reconstruct vertical changes in relative sea-level *via* the use of **sea-level index points**. (Figure 2.19). Errors in the determination of age and altitude require these data to be plotted as a sea-level band commonly up to 4 m in width, and this significantly reduces the resolution of the sea-level record. The sources and magnitude of errors associated with sea-level index points are comprehensively discussed in Shennan (1980, 1982, 1986), Heyworth & Kidson (1982), and van de Plassche (1986).

The Thames stratigraphy exhibits a marked reduction in the abundance of organic deposits toward the present day, and this is a common feature of many UK records (Allen, 1991). The last major period of peat formation (Tilbury III) terminated around 4100 Cal. BP, and the succeeding organic deposits (Tilbury IV and V) were substantially thinner, and of localised extent. Formation of Tilbury IV ended around 2900 Cal. BP (comparable to the Romano-British Transgression) and Tilbury V has yet to be dated. This switch in the nature of sedimentation within the Thames estuary has prevented the establishment of sea-level index points during the last 2500 years. Consequently, the quality of the late Holocene sea-level record in this system is poor (Long, 1995). The fact that this reduction in the availability of datable organic material is a characteristic of many UK sites is one reason why there are so few sea-level index points from the last 2000 years.

Large error envelopes and an absence of suitable datable material commonly frustrate the reconstruction of late Holocene sea-level change in other parts of Europe as well. For example, De Jong (1984) dated a number of organic units preserved beneath the sand dunes of the Frisian Islands, situated off the north coast of the Netherlands (Figure 2.20). These dates could not be directly used in sea-level reconstruction however, as they possessed no definable relationship to sea-level (indicative meaning). De Groot *et al.* (1996) attempted to overcome this problem by examining the sedimentological characteristics of adjacent minerogenic units. They used a set of criteria developed by Roep (1986) to relocate the position of palaeo-mean high water, and from this developed a 'best estimate' indicative meaning for the dated deposits. This novel approach was ultimately unsuccessful since the reconstructed altitude of palaeo-mean high water in a core was frequently centimetres to decimetres below the dated organic unit, and in some cases this vertical offset was up to 1 m. Consequently, a variable time lag existed between deposition around mean high water and the subsequent formation of the dated organic horizon. De Groot *et al.* (1996) suggested this lag may be up to several centuries in length, and therefore the chronology of the observed changes in mean high water level is at best provisional. This poor temporal resolution was matched by very low altitudinal control. The error band associated with palaeo-mean high water was 1 m wide, similar in size to the overall sea-level rise suggested to have occurred during the study period (Figure 2.21). Despite these limitations, the authors proposed an acceleration in the rate of mean high water rise around 800 Cal. BP, and a reduction around 400 Cal. BP which they correlate with the Little Ice Age climate period. The large errors associated with this work clearly render the inferred climatic relationship somewhat equivocal. This study demonstrates that unless the techniques employed are of the highest possible precision, the magnitude of age and altitude errors will be comparable to the size of the changes under investigation.

Similar problems are associated with the work of Ters (1987), whose synthesis paper of sea-level change on the French Atlantic coast identifies five sea-level oscillations during the last 3400 years. No account was taken of the differences in indicative meaning of the wide range of material dated and as a result, significant and variable altitudinal errors are associated with much of the dataset. This brings into doubt the reliability of the proposed chronology, especially when some of the suggested oscillations are only of the order of 50 cm, a resolution not achieved by the more rigorous work of Devoy (1979) and Long (1995) in the Thames.

Ters (1987) compared the timing of these 'oscillations' with the (oscillating) 'global' sea-level curve of Fairbridge (1961), Mörner's (1969) regional 'eustatic' curve, Shepherd's (1963) curve from North America, and the New Zealand curve of Suggate (1968). It is apparent from the discussion in Section 2.1 that comparing such an eclectic set of data derived from differing spatial scales and geographic locations with a local chronology has no sound theoretical basis. Ters (1987) further speculated that observed submergence events correlated with warm periods determined from the $^{18}O/^{16}O$ ratios of Greenland ice cores, and that the primary driving force behind late Holocene sea-level change was therefore glacio-eustasy. However, no estimates of the volumes of ice required to produce the suggested oscillations were given to check whether such correlation really equated to causation. For example, the thermal maximum at 2200 Cal. BP is suggested to have produced a rise in mean sea level of at least 8.5 m in 800 years (Ters, 1987). Clearly, inferred climatic signals within sea-level records must be associated with viable explanations of the mechanisms by which the two are related (Pirazzoli, 1996).

2.5.1.2 Lateral Shifts in Marine Influence

Sea-level tendencies consider the timing of lateral translations in marine influence, thereby removing the problems associated with precise determination of altitude (Shennan *et al.*, 1983). This approach enables a variety of dated sea-level indicators to be used in combination, without the need to quantify their indicative meanings. Tendencies are applied in an hierarchical fashion, constructing dominant or regional tendencies from a number of local ones. There are a number of problems associated with tendency analysis however. Firstly, the technique assumes that regional processes, such as changes in sea level, will be reflected by a regional tendency (Shennan *et al.*, 1983). In reality, the local tendencies may be strongly influenced by other processes, such as changes in sediment supply, which will complicate production of a dominant tendency (Long, 1992). Secondly, delimiting the period of a tendency is strongly affected by the temporal distribution of data points. Consequently, tendency records require large numbers of dated sea-level indicators, and their chronology is strongly influenced by data distribution and sampling strategy (Long, 1992).

One classic application of tendency analysis is the Fenland chronology of Shennan (1986), derived from 112 radiocarbon dated sea-level index points (Figures 2.17 and 2.22). The post 3000 Cal. BP period is dominated by positive tendencies (Wash VI, VII & VIII), with two brief periods of negative tendency (Fenland VI lasting *c.*350 years, and Fenland VII lasting only around 200 years). Shennan (1994) produced a hypothetical local relative sea-level curve for the Fenland area by modifying the regional eustatic curve of Mörner (1976) to account for the estimated 1 mm a^{-1} subsidence experienced in the Fenland (Shennan, 1986). This curve (Figure 2.23) predicts a number of relative sea-level changes may have occurred during the last 3000 years. Shennan (1994) suggested that three periods of coastline advance, between 3200 to 2800 Cal. BP, *c.* 2400 Cal. BP, and after 1900 Cal. BP, may be related to falls either of relative sea-level or in the rate of relative sea-level rise.

Waller (1994) examined the timing and extent of marsh expansion and marine incursion using palaeogeographic maps and identified three distinct phases of change during the same period using 83 radiocarbon dated sea-level index points. The first period, from 3000 to 2600 Cal. BP was characterised by a fen-wide seaward extension of freshwater conditions. This was followed by a prolonged return to marine conditions between 2600 to 1700 Cal. BP. Reliable sea-level data after this time are sparse (10 radiocarbon dates), and human activities, such as embanking and drainage, have complicated the record of change. Nevertheless, Waller (1994) suggested that this interval was dominated by a gradually rising relative sea-level, although there was some evidence for a negative tendency *c.*1700 to 1550 Cal. BP, during the Roman period.

Clearly there is some disagreement between the Wash/Fenland chronology of Shennan (1986, 1994) and the palaeogeographic interpretation of Waller (1994). Shennan envisages three periods in the last 3000 years when the rate of relative sea-level rise was reduced, whilst Waller prefers a dominant rise in relative sea-level interrupted perhaps once during the Roman Period. Consequently, regardless of which scheme is the more reliable, it is difficult to accurately construct a chronology of relative sea-level change for the East Anglian Fenland which represents the most intensively studied area in the UK.

Changes in coastal configuration have been used by a number of authors to produce a general impression of late Holocene relative sea-level change around Spain and Portugal. This information was obtained by examining the abundant spit bar systems which typify the littoral zone in this region. Zazo *et al.* (1994) and Lario *et al.* (1995), suggested that periods typified by progradation reflect stillstands or gentle falls in sea level, whilst phases of minimal progradation or erosion indicate relative sea-level rise or increased storminess (Dabrio *et al.*, 1995). Goy *et al.* (1996) however, suggested that increases in progradation might also be associated with a greater influx of Atlantic Superficial Water into the Mediterranean during anticyclonic conditions, and on this basis inferred a major change in prevailing wind direction between 3200 and 2800 Cal. BP around Almería (Figure 2.20). Goy *et al.* (1996) proposed a shift from westerly to easterly circulation, causing a reversal in the direction of littoral drift resulting in increased progradation, and suggested that this situation was typical of conditions from the Little Ice Age to the present day. Prior to this, during the Medieval Warm Epoch, there is little evidence of progradation, probably as a consequence of low pressure conditions (Goy *et al.*, 1996). Changes in the circulation of the Mediterranean Sea may be significant to the Atlantic system since variations in the outflow of Mediterranean water has been suggested as a possible cause of climate change, deflecting the warm North Atlantic Drift away from the Norwegian Sea toward the Labrador Sea (Johnson, 1997).

An examination of Guadalete estuary in the Bay of Cadiz (Figure 2.20), suggests that a prolonged period of progradation came to an end around 2800 Cal. BP, where there was a 300 year break in sedimentation (Dabrio *et al.*, 1995). Progradation resumed between 2500 Cal. BP and 800 Cal. BP, before a second break in sedimentation, indicating increased relative sea-level rise or storminess, occurred between *c.* 800 Cal. BP and 700 Cal. BP. The last 700 years have been characterised by ongoing progradation.

The influence of storms on coastal sedimentation is also apparent in the record from Dungeness Foreland, a large gravel barrier on the southern coast of England (Figure 2.17). Its formation permitted the establishment of saltmarsh and freshwater communities in the protected back barrier environment, producing prolonged peat accumulation from *c.* 6800 Cal. BP to *c.* 1900 Cal. BP (Long & Innes, 1995). After 1900 Cal. BP, there is increased evidence of barrier breaching and for the progressive spread of marine conditions across the marsh (Long & Innes, 1995), coupled with a dramatic reduction in the occurrence of organic deposits. Long & Hughes (1995) suggested that the alternating gravel and marsh deposits typical of the period between 4400 Cal. BP and 1900 Cal. BP were formed as a result of variations in storm intensity and sediment supply, and not solely as a response to long-term fluctuating relative sea-level. The prominence of minerogenic sedimentation in Romney Marsh after 1900 Cal. BP has hampered the development of a relative sea-level curve for the late Holocene period. A further complicating factor has been the substantial land reclamation which became increasingly significant after *c.* 900 Cal. BP.

A combination of data indicating lateral shoreline movements and vertical changes in relative sea-level have been used to investigate the Severn Estuary system, located on the west coast of southern Britain (Figure 2.17). Whilst no detailed peat-based chronology comparable to the Thames exists, a number of extensive peat beds occur, mainly dating between *c.* 6800 Cal. BP and 3200 Cal. BP (Allen, 1990d). The last 2600 Cal. years have witnessed the replacement of organic-rich deposits by estuarine silt and clay. The period after 1900 Cal. BP witnessed much human activity within the estuary, and substantial embanking, drainage and reclamation of the alluvial wetlands was undertaken from the Romano-British period onward (Allen & Rae, 1987).

The influence of this human activity on the wetland environment of the Severn has been utilised by Allen and his co-workers to determine late Holocene coastal change. Allen (1987a), and Allen & Rae (1987), described four discrete lithostratigraphic units that occur throughout the estuary, and reclamation of parts of these surfaces *via* the construction of seabanks and drainage ditches at various times during the last 2000 years has provided a means of dating these formations. The age of seabanks can be determined by dating wood, organic matter or archaeological artefacts within or beneath them, or through the use of maps and other documentary sources. The reclaimed land surfaces can also be dated through the collection of artefacts, analysis of field patterns, architectural features or map and documentary evidence (Allen, 1991). For sediments of post industrial age, Allen (1987b, 1987c, 1988) and Allen & Rae (1986) have utilised metal and coal dust contents to determine their ages by comparison with the documented increase of industry and associated pollution. Allen & Rae (1988) measured the relative height difference of surfaces either side of a number of dated seabanks throughout the estuary to produce a series of sedimentation rates. Assuming equilibrium with sea level, these rates were used as proxies for the rate of relative sea-level rise (Allen, 1990d; 1991). Whilst considerable scatter was present in the values of elevation difference with age, this was suggested to reflect tidal creek distribution and uneven antecedent topography, and Allen (1991) suggested that these errors were no worse than those associated with sea-level curves derived from intercalated sequences.

These combined results indicate that relative sea-level has risen by at least 1.5 m during the last 2000 years. From *c.* 1800 Cal. BP to *c.* 700 Cal. BP, the average rate of relative sea-level rise was only around 0.4 mm a^{-1} (Allen, 1991) and at some point during this period an erosive transgression took place (Allen, 1987a). The rate of relative sea-level rise increased to around 0.8 mm a^{-1} between *c.* 700 Cal. BP and 150 Cal. BP, and the chemical evidence suggests that further increases occurred after this culminating in the current rate of 4.7 mm a^{-1} (Allen, 1991). The evidence derived from shoreline movements indicates that whilst, on average, relative sea-level has been rising at an increasing rate, a number of shoreline advances and retreats have occurred that appear to be largely synchronous throughout the estuary (Allen & Rae, 1987). Allen (1987a) proposed that these changes could indicate variations in the wind/wave climate, or an internal cycle within the marsh, promoting destabilisation, mass failure and cliff formation. Allen (1987a) suggested that small oscillations in relative sea-level may also have been implicated but were unlikely to have been able to produce the large cliffs apparent in the formations in their own right. It is evident therefore, that whilst lateral movements in marine influence may reflect changes in relative sea-level, they may also indicate variations in other phenomena such as changes in wind direction and strength.

2.5.1.3 Summary

Studies of late Holocene relative sea-level change in Europe are frustrated by a lack of suitable datable material from the last 2000 to 2500 years (Allen, 1991; Long, 1995). This is compounded by the errors associated with age and altitude that are often of comparable magnitude to the changes under investigation (e.g. De Groot *et al.*, 1996).

A widespread replacement of organic sediments with marine minerogenic facies appears to have occurred *c.* 2800 Cal. BP in much of southern Britain (e.g. Devoy, 1979; Shennan, 1994). At the same time changes in the coastal configuration of Spain and Portugal were also occurring (Dabrio *et al.*, 1995; Goy *et al.*, 1996). Furthermore, a second change in sedimentation was recorded by Dabrio *et al.* (1995) *c.* 800 Cal. BP at a similar time when Allen (1987a) notes an increase in the rate of relative sea-level rise in the Severn Estuary.

2.5.2 Data from the North Atlantic West Coast

2.5.2.1 Vertical Changes in Relative Sea-Level

The saltmarsh sediments of North America frequently contain a much larger proportion of organic material than European ones (Long *et al.*, in press) and so investigations of late Holocene relative sea-level change are not hampered by a lack of datable material to the same extent as in the UK. These organogenic marshes may be subject to greater rates of compaction however, resulting in larger altitudinal errors associated with sea-level index points from intercalated sequences (e.g. Pizutto & Schwendt, 1997).

The publication by Meyerson (1972), proposing a major increase in the rate of relative sea-level rise around 1700 Cal. BP has become a benchmark paper cited in many investigations of late Holocene relative sea-level change in North America (e.g. Fletcher *et al.*, 1991, 1993; John & Pizzuto, 1995). Meyerson (1972) examined the abundance and nature of organic material present in three cores from Dennis Creek Marsh, New Jersey (Figure 2.24). These data were complemented by pollen and palaeosalinity analyses, and a chronology of change was constructed from six radiocarbon dates. Inspection of these data reveals a number of inadequacies, casting some doubt on the reality of this proposed acceleration in relative sea-level rise. Firstly, at least one of the dates is derived from allochthonous material, and all three cores are situated adjacent to creeks systems which may have migrated in the past, erasing portions of the record. Secondly, the interpretations of relative sea-level change are based upon 'tentative' correlations between undated lithostratigraphic units, subject to compaction, that were not levelled relative to a tidal datum. Therefore, the 1700 Cal. BP 'event' is in reality, based upon a single date from a core nearly five kilometres away from one of the other dated deposits. Additionally, this date is from the base of a freshwater swamp which, as Meyerson (1972) points out, cannot be considered a reliable indicator of sea-level change since it formed in an environment an unknown distance above high tide level.

Fletcher *et al.* (1993), presented evidence for five, rapid, short-term episodes of accelerated relative sea-level rise from Wolfe Glade Marsh, Delaware (Figure 2.24), which they suggested may be related to climate-induced variations in the Gulf Stream (Figure 2.25). They suggested that only the accelerations at 4400 ± 200 Cal. BP and 1800 ± 200 Cal. BP were likely to reflect more widely recognised sea-level movements on the basis that they appeared to correlate most strongly with phases of warmer climate deduced from glacial data. Fletcher *et al.* (1993) pointed out that the apparent movements in relative sea-level of between 25 cm and 100 cm were hard to account for in terms of climate change since steric effects are too small, and it was unlikely that ice masses could respond to climate change so quickly. Instead they proposed that changes in ocean surface topography resulting from variations in Gulf Stream strength could provide an explanation for the observed variability. Fletcher *et al.* (1993) suggested that during warm periods, the atmospheric thermal gradient between equator and pole is reduced, the Gulf Stream flow is diminished and its dynamic height decreases, raising sea levels at the adjacent US coast. Conversely, cooler conditions would equate to enhanced Gulf Stream flow and lower sea-level. This interpretation is at odds with the data presented in Sections 2.2 which suggests that cooling is associated with reduced Gulf Stream vigour, and it is possible that Fletcher *et al.* (1993) are underestimating the significance of thermohaline circulation in controlling the Gulf Stream.

There are also some doubts concerning the reliability of the accelerations in the rate of sea-level rise themselves, owing to the fact that the study site is situated a considerable distance from the open coast, and has been subject to isolation from the sea by prograding barriers and spits. Fletcher *et al.* (1993), equated 'true' sea-level movements to synchronous, site-wide changes in sedimentation, but such sequences could equally be produced by the emplacement and breakdown of a barrier across the mouth of the system. The spatial distribution of cores is not sufficient to reject the latter as a viable cause of at least some of the possible submergence/emergence events recorded.

John & Pizzuto (1995) proposed a 'dramatic transgression' took place in Delaware Bay around 2000 Cal. BP (Figure 2.24). Fifty-five cores were taken along a 15 km stretch of the Leipsig River and revealed two persistent saltmarsh peat beds interleaved by tidal river mud. Nine radiocarbon dates were presented from various parts of the system, which John & Pizzuto (1995) interpreted as indicating a regressive event at 2300 Cal. BP followed by the transgression at 2000 Cal. BP, and culminating in a regressive tidal wetland facies at around 900 Cal. BP. The evidence for the 'dramatic' nature of the transgression is supplied by the synchroneity of the Leipsig River event with similar transgressions recorded at Wolfe Glade (Fletcher *et al.*, 1991, 1993) and Dennis Creek Marsh (Meyerson, 1972). The transgressive overlap at Leipsig River is only dated in one place however, despite being over 10 km in length, varying in altitude by nearly 2 metres, and being overlain down-river by the regressive overlap of the '900 Cal. BP' wetland which was dated to between 2089 and 2329 Cal. BP. Additionally as mentioned above, the supporting transgressions of Meyerson (1972) and Fletcher *et al.* (1993) are suspect.

Patton & Horne (1991), investigating the Connecticut River Estuary (Figure 2.24), collected around 90 vibracores and produced a chronology for sedimentation from 25 radiocarbon dates. These were used to define two submergence curves, the first derived from inundated soils and drowned trees, and the second taken from intertidal peats (Figure 2.26). Patton & Horne (1991) proposed that these curves delimit the possible range of submergence curves. They suggested that the upper curve corresponds to the approximate trend of high water, but give no reference water level for the lower curve which was used in the subsequent analyses. Patton & Horne (1991) suggested that this lower curve indicated three periods of differing rates of relative sea-level rise during the last 4000 years. They recorded a deceleration in the rate of relative sea-level rise around 1700 Cal. BP which was followed by an acceleration *c.* 300 Cal. BP. These rates are equivocal since they are based on a single curve which does not take into account the errors associated with each data point (despite the fact that error bars are present). Additionally, whilst many basal peat deposits were used to reduce the significance of compaction, only one of these was described as being of saltmarsh origin, and no altitudinal data were given.

De Rijk (1995a) in a study of Great Marshes, Barnstable, Massachusetts (Figure 2.24), used foraminifera (see Section 3.4.1) to determine palaeoecological changes in two cores. Dating control was excellent, comprising 46 AMS radiocarbon dates and ^{210}Pb measurements from the upper core sections. Using the hydrological model of Schellekens (1994), De Rijk (1995a) suggested that accelerations in relative sea-level rise would result in increased porewater salinity, which in turn would be reflected by changing foraminiferal assemblages. On this basis, accelerations were proposed at 1750, 850, 600, and 400 Cal. BP. It was noted however, that changes were not synchronous between cores, which were situated about 2.5 km apart, and it is therefore impossible to gauge the spatial extent of these variations. The possibility that they reflect localised changes in salinity regime and not rising relative sea-level cannot be excluded.

Scott *et al.* (1995a, 1995b) also used precise, foraminiferal-based sea-level index points to investigate changes in Nova Scotia, and compared these with a record from South Carolina (Figure 2.24). They identified a marked acceleration in the rate of relative sea-level rise between 5000 Cal. BP and 4000 Cal. BP, followed by a slower rise to the present day (Figure 2.27). The evidence for this is somewhat equivocal since, in order to combine the two datasets, Scott *et al.* (1995a, 1995b) assume that the event is eustatic and therefore occurred at both sites at the same time.

Kelley *et al.* (1995) collected 50 cores from the saltmarshes of Wells, Maine, and constructed a sea-level curve from 43 radiocarbon dates (Figure 2.24). A rigorous selection procedure was used to screen the dates for potential errors, and a curve of change during the last 5000 years fitted to the ten best dates (Figure 2.28). Their smoothed curve shows a decreasing rate of relative sea-level rise during the late Holocene with no evidence of sudden accelerations or decelerations. The scatter in the graph is considerable, even if dates from only basal contexts are considered, and this is attributed to humic acid contamination (Belknap *et al.*, 1989; Gehrels & Belknap, 1992).

Gehrels *et al.* (1996) presented a synthesis of sea-level data from four sites along the coast of Maine, including the marsh at Wells described by Kelley *et al.* (1995), to which they added some new data. As before, a rigorous consideration of errors was presented, but this time saltmarsh foraminifera were used to determine the indicative meaning of the deposits, corrected for changes in tidal range by the tidal model of Sucsy (1990) and Sucsy *et al.* (1993). The general form of the resultant curves (Figure 2.29) was remarkably similar, although slight differences in the timings of the transitions was apparent perhaps due to differing rates of isostatic movement (Gehrels *et al.*, 1996). Consistent features were a deceleration in the rate of relative sea-level rise between 4000 and 3000 Cal. BP, and a possible acceleration in the last 1000 years.

2.5.2.2 Lateral Shifts in Marine Influence

A number of north American studies have employed approaches similar to those of tendency analysis used in the UK. An early example of this is the work by Rampino & Sanders (1981) which focused on the timing of peat inception and termination along the northeastern coast of the US. Using a database of 19 radiocarbon dates, Rampino & Sanders (1981) endeavoured to relate periods during the Holocene dominated by peat growth to the 'eustatic' sea level curve of Fairbridge & Hillaire-Marcel (1977). These results were interpreted as indicating a phase of marsh establishment between 4400 to 3200 Cal. BP, perhaps relating to a slowing in submergence rate also noted at this time by Kraft (1971) and Belknap & Kraft (1977). The number of dates is clearly a limiting factor in the accuracy of this approach and the authors suggested that these results would be confirmed in the future as more data became available.

Van de Plassche (1991) conducted a detailed stratigraphic examination of the Hammock River Marsh in Connecticut, comprising over 400 boreholes (Figure 2.24). After a thorough discussion of possible errors relating to compaction and diachronous contacts, van de Plassche (1991) suggested that the horizontality and depth consistency of many transgressive and regressive overlaps permitted depth to be used as a proxy for age. In this way he constructed a tendency histogram from transgressive and regressive overlaps, identifying five periods of increased and decreased rates of relative sea-level rise. After considering a number of factors, van de Plassche (1991) suggested that these marsh-wide, synchronous stratigraphic changes were best explained in terms of changes in the rate of mean high water rise, and that this in turn was best accounted for by changes in the level of the northwest Atlantic. These variations were placed in a temporal context *via* six radiocarbon dates to produce an age-depth diagram of mean high water rise (Figure 2.30). The large error band precluded detailed analysis of relative sea-level change, and so the stratigraphic tendency data were inserted to produce a fluctuating second order estimate of mean high water rise. Finally, these fluctuations were compared with glacier advances and retreats from around the globe (Röthlisberger, 1986), and a northern hemisphere temperature index for the last 1500 years derived from the central England temperature record, Greenland oxygen isotope data, and Californian tree lines (Hammer *et al.*, 1980). These displayed the same number of fluctuations as the sea-level record from Hammock River (Figure 2.31), and van de Plassche (1991) suggested that the timings could also be made to correlate with 'acceptable shifts in time'. He concluded that a higher age and depth resolution was required before the precise nature of the relationship between low amplitude sea-level changes and climate could be determined.

At the same time, Thomas & Varekamp (1991) were developing a combined faunal and chemical methodology to investigate the late Holocene saltmarsh development in Hammock River. Representative cores were selected from the extensive database compiled by van de Plassche (1991), and detailed foraminiferal, and trace metal analysis conducted on this material. Previous work by Scott & Medioli (1978, 1980a) had demonstrated that the distribution of saltmarsh foraminifera was controlled by elevation above mean tidal level, and that the highest marsh environment was characterised by a dominance of *Trochammina macrescens* (see Section 3.4.1.1). Thomas & Varekamp (1991) used the percentage frequencies of *T. macrescens* relative to the percentage frequencies of other species as a proxy for flooding frequency. Levels of iron, zinc and sulphur were found to be related to sediment grain size, and were inversely correlated with percentage frequencies of *T. macrescens*. The results from a number of cores showed the same sequence of events, indicating three periods of marsh submergence that the authors suggested may be related to 'true' eustatic accelerations.

This work was continued by Varekamp *et al.* (1992) who used fossil foraminiferal data to determine the indicative meanings of sediments which were expressed as marsh palaeoenvironmental curves, plotted relative to mean high water. These foraminiferal data were also used to calibrate geochemical data, enabling marsh palaeoenvironmental curves to be constructed from the iron content of sediments. An age-depth relationship was derived using seven AMS radiocarbon dates and the historical increase in copper pollution. This was then used to construct a curve of changes in mean high water (Figure 2.32). Varekamp *et al.* (1992) identified accelerations in the rate of relative sea-level rise at *c.* 800 Cal. BP, 300 Cal. BP, and 200 Cal. BP, whilst a deceleration was noted at 550 Cal. BP Additionally, rates of relative sea-level rise were very low between 1250 Cal. BP and 800 Cal. BP, and between 250 Cal. BP and 120 Cal. BP.

The main limitation of this method is obtaining precise age estimates for the submergence and emergence events. The work of Varekamp *et al.* (1992) possesses age uncertainties of ±150 ^{14}C years inherent to the radiocarbon technique, which render it impossible to precisely relate observed changes to documented climate events. Additionally, no radiocarbon dates were obtained for the interval between 1300 Cal. BP and 600 Cal. BP, and ages had to be estimated by interpolation. Consequently, Varekamp *et al.* (1992) were forced to conclude that there was no unequivocal correlation between intervals with high rates of relative sea-level rise and warm periods in the climate record.

Nydick *et al.* (1995), used the same marsh palaeoenvironmental curve technique in a study of a nearby marsh at Guilford (Figure 2.24). They examined 70 cores and used 17 radiocarbon dates to subdivide the record into three periods on the basis of inferred rate of relative sea-level rise (Figure 2.33). The only similarity with the Clinton record is the acceleration in relative sea-level rise around 400 to 300 Cal. BP, although this clearly pre-dates any change due to human induced global warming. Once again, the resulting error envelope is sufficient large not to preclude a constant rate of relative sea-level rise during the study period.

2.5.2.3 Summary

The age-altitude analyses conducted in North America have produced more suggested variations in relative sea-level during the late Holocene than their European counterparts. Many of these studies however, fail to employ the same rigorous evaluation of errors in age and altitude that is strongly advocated in the UK research methodology (e.g. John & Pizzuto, 1995). The application of marsh palaeoenvironmental curves has produced the most detailed picture of accelerations\decelerations in the rate of relative sea-level rise during the late Holocene period, although these records do not provide precise information on vertical changes owing to sediment compaction.

The precise timing of changes in relative sea-level differ between areas, as would be expected *a priori* given the differences in crustal movements experienced along the

American Atlantic coast. Nevertheless, a number of periods where similar changes in relative sea-level occur are evident. There is an indication of a general deceleration in the rate of relative sea-level rise between 4000 and 3000 Cal. BP in a number of the study marshes (Rampino & Sanders, 1981; Kelley *et al.*, 1995; Scott *et al.*, 1995a, 1995b; Scott & Medioli, 1995). An acceleration around 1800 Cal. BP is also noted by Meyerson (1971), Fletcher *et al.* (1991, 1993), and De Rijk (1995), after which the rate of relative sea-level rise appears to have slowed. A final acceleration or series of accelerations are noted in the last 1000 years, notably around 800 Cal. BP (Varekamp *et al.*, 1992; De Rijk, 1995; Kelley *et al.*, 1995), and also possibly between *c.* 400 and 300 Cal. BP (Patton & Horne, 1991; Varekamp *et al.*, 1992; De Rijk, 1995; Kelley *et al.*, 1995; Nydick *et al.*, 1995).

2.5.3 Methodological Evaluation

There are two fundamental methodologies evident in the studies outlined above: those that consider vertical movements in relative sea-level typically employing sea-level index points to construct sea-level curves; and those that use some form of tendency analysis to examine lateral shifts in marine influence.

The examination of vertical changes in late Holocene relative sea-level is primarily hampered by the fact that age-altitude errors are commonly of comparable magnitude to the scale of variation under consideration (Section 2.5.1.3). This is compounded in the UK by a lack of material suitable for dating.

A common pitfall within the sea-level literature is the attribution of a sea-level signal to a feature that may be explained in other ways. Two sea-level index points with different ages and altitudes are required to demonstrate a rise in relative sea-level and yet there are many examples of a rise, or acceleration in the rate of rise, being inferred from a single lithostratigraphy boundary. For example, Fletcher *et al.* (1993) infer an increase in the rate of relative sea-level rise from a **transgressive overlap**. Shennan (1982) and Tooley (1982) have emphasised that transgressive and regressive overlaps should be used as purely descriptive terms in which no process is implied. Whilst the occurrence of a widespread change in stratigraphy may indeed reflect rising sea level, it may also result from changes in sediment supply, site exposure, coastal configuration, wind/wave climate or tidal range. It is therefore necessary to demonstrate that relative sea-level rise is the most viable explanation of the observed lithostratigraphic change, either by providing supporting evidence from other systems or by eliminating other factors as potential causes. This cannot be achieved by Fletcher *et al.* (1993) since their study marsh is situated behind a spit known to have prograded in the past, suggesting that factors other than sea level rise may be implicated in causing the observed change. The high resolution and precision required by late Holocene sea-level research necessitates the careful selection of sensitive sites. Where possible, these should be relatively simple systems unlikely to have been significantly influenced by other factors such as changing coastal configuration or tidal wave deformation.

In more extreme cases the inundation of a basal freshwater peat unit has been cited as evidence for an acceleration in the rate of relative sea-level rise. This is clearly an invalid statement since freshwater peats have no quantifiable relationship to sea-level. Dating the transgressive contact of such peats can provide a reliable sea-level index point, fixing the altitude of sea-level in time. It does not however, provide any evidence of an acceleration in sea-level since the deposit was unrelated to the tidal frame until its inundation. Only when a sequence of basal peat deposits are presented can accelerations or decelerations be identified. Even these sequences are of limited value in high resolution studies of late Holocene relative sea-level change and climate owing to the weaknesses inherent to the radiocarbon technique. The stochastic nature of radioactive decay means that even the most precise age estimates are associated with an error term of at least ± 50 ^{14}C years. In order to detect a change in the rate of relative sea-level rise from basal deposits, a minimum of three radiocarbon dates is required, which will at best span a period of 200 ^{14}C years. Changes will not be discernible at timescales shorter than this, and when converted to calibrated years BP, this age window will increase.

Also, implicit within such rate calculations is the assumption that saltmarshes accrete in equilibrium with sea level. In this way, the calculation of sedimentation rates between sea-level index points provides a reliable proxy for the rate of relative sea-level rise. The saltmarsh models of Allen (1990a, 1990b, 1990c) suggest that this situation is not the case. Instead, rates of sedimentation vary greatly between different depositional environments, and interpolation between sea-level index points may mask many changing rates of relative sea-level rise within a single average figure. When searching for a climate signature within a sea-level record, these changes in rates of relative sea-level rise are of fundamental significance, and therefore assumed equilibrium is as unhelpful as it is misleading.

Reconstruction of relative sea-level change from sea-level tendencies is still complicated by the fact that an increase or decrease in marine influence at a single site cannot be used to infer a vertical movement of relative sea-level. The approach rests upon the assumption that changes in relative sea-level will be documented at a number of sites and therefore identification of periods dominated by similar local or regional tendencies can be used to infer vertical movements in relative sea-level.

The marsh palaeoenvironmental curve technique offers the potential to continuously monitor changes in water depth, and thereby has the ability to detect changing rates of relative sea-level rise. Its greatest limitation remains dating of emergence or submergence events, particularly if these occur within minerogenic sequences. This means that whilst marsh palaeoenvironmental curves may detect variations in relative sea-level, they may not be able to define the timing of these events accurately enough to allow a climate relationship to be inferred. The altitude of these events will also be influenced by sediment consolidation and compaction, leading to errors in the calculation of rates of relative sea-level rise and the former altitude of sea level.

A possible solution to these problems lies in an integration of age-altitude methods with those of marsh palaeoenvironmental curves. Continuous monitoring of water level changes within a core provides information on lateral shifts in depositional environment related to times when sedimentation and sea-level rise are not in equilibrium. The combination of such records from a number of cores across a saltmarsh can reveal phases of submergence or emergence. Age-altitude analysis can provide complimentary information on vertical movements in relative sea-level permitting the submergence and emergence events to be more reliably interpreted in terms of accelerations or decelerations in the rate of relative sea-level rise. This integration of the two approaches means that sea-level index points can also be used to place the accelerations and decelerations of relative sea-level rise inferred from marsh palaeoenvironmental curves into a vertical and temporal context. The limitations inherent with the radiocarbon technique mean that these changes in relative sea-level will never be unequivocally related to dated climatic events.

Until a new dating method is developed, the aim must be to constrain the possible period of change, and within these limits investigate their relationship to climate variations.

2.5.4 *Summary of Late Holocene Sea-Level Data*

The changes in late Holocene relative sea-level identified on both sides of the Atlantic are summarised in Figure 2.34. This diagram is not a traditional sea-level tendency plot since it portrays accelerations and decelerations in the rate of relative sea-level rise. There is a large degree of variability between the records and this is best accounted for by the influence of site specific factors and limitations in the methodologies employed to construct them. In spite of this variability however, there are three periods in the record that exhibit some coherency in the timing and direction of change.

The first period is the generally low rate of relative sea-level rise between *c.* 4000 and 3000 Cal. BP noted in the UK Fenland data (e.g. Shennan, 1994) and from some studies in North America (e.g. Gehrels *et al.*, 1996). This appears to have been a significant event owing to the widespread expansion of peat forming communities on both sides of the Atlantic, and possibly indicates a fall in relative sea-level as suggested by Godwin (1940a). At this time a major shift in the depositional regime along the Mediterranean coasts was taking place consistent with a reduced rate of relative sea-level rise but possibly also indicating a change in atmospheric and oceanic circulation (Dabrio *et al.*, 1995; Goy *et al.*, 1996).

An acceleration in the rate of relative sea-level rise *c.* 1800 Cal. BP is recorded in a number of sites in North America, although as mentioned in Section 2.5.2 this has an equivocal basis in at least two of the records. The fact that similar shifts are recognised by the more precise studies of Varekamp *et al.* (1992) and, within the error range associated with radiocarbon dating, by Nydick *et al.* (1995) and de Rijk (1995a), means that there is sufficient supporting evidence to accept this acceleration. It should be noted however, that Patton & Horne (1991) detect a deceleration in the rate of relative sea-level rise at this time, and a similar deceleration is inferred in the Fenland record of the UK, although this is short-lived (*c.* 150 years) and terminates *c.* 1550 Cal. BP (Waller, 1994).

Finally, records from a number of the North American sites show evidence for accelerations in the rate of relative sea-level rise around 800 Cal. BP and 400 Cal. BP

(Varekamp *et al.*, 1991; van de Plassche, 1991; De Rijk, 1995a; Nydick *et al.*, 1995). Whilst no records of comparable quality exist in the UK, data from the Severn Estuary indicates an acceleration in the rate of relative sea-level rise *c.* 750 Cal. BP (Allen, 1991), and there is further evidence of changing circulation in the Mediterranean (Goy *et al.*, 1996).

2.6 EXISTING EVIDENCE FOR A CLIMATE SIGNAL IN LATE HOLOCENE SEA-LEVEL RECORDS

The strength of evidence for a climate signal in late Holocene sea-level records is assessed by comparing the summary of late Holocene relative sea-level change presented in Section 2.5.4 with the summary of climate variations presented in Section 2.4.4. These data are summarised diagramatically in Figure 2.35.

It is striking how changes in the rate of relative sea-level rise coincide with variations in the climate record from the Sargasso Sea which has been related to fluctuations in the strength of the Gulf Stream (Section 2.4.4). The quiescent phase in the Sargasso Sea at *c.* 3000 Cal. BP coincides with a cooling and freshening of the Greenland and Norwegian Seas, an apparent climate deterioration in the Netherlands, and a reduction in the rate of relative sea-level rise on both sides of the Atlantic. However, it should be noted that the deceleration is recognised earlier along the western Atlantic margin where it appears to lead the oceanic changes.

The 1800 Cal. BP acceleration is also coincident with a change in the Sargasso Sea, although this time it indicates increased Gulf Stream vigour. At the same time, the terrestrial climate records show a shift to warmer and drier conditions in Europe, and a reduction in terrestrial ice. It is possible that steric or glacio-eustatic changes may be implicated in this acceleration, although both sea level and climate may be responding to the same oceanic forcing.

The final accelerations in the rate of relative sea-level rise at *c.* 800 Cal. BP and *c.* 400 Cal. BP are less clearly associated with a change in climate. There are certainly shifts in the oceanic circulation at these times, but at *c.* 800 Cal. BP there appears to have been a reduction in Gulf Stream strength and a deterioration in climate that intuitively should produce a deceleration. The acceleration at *c.* 400 Cal. BP however, is consistent with the changes inferred from the Sargasso Sea record and coincides with the brief phase of warmth inferred from terrestrial records.

2.7 A CLIMATE-BASED PREDICTION OF LATE HOLOCENE RELATIVE SEA-LEVEL CHANGE IN SOUTHERN BRITAIN

On the basis of the evidence presented in Sections 2.4.4 and 2.6, a qualitative prediction of expected relative sea-level change in southern Britain is proposed in Figure 2.36. This will serve as a benchmark against which the relative sea-level records derived in Chapter 6 are tested for a climate signal. The predicted relative sea-level scenario consists of the following five phases:

RSLC-I. A reduction in the rate of relative sea-level rise around 3000 Cal. BP in response to a deterioration in climate and a reduction in Gulf Stream strength;

RSLC-II. A renewed increase in the rate of relative sea-level rise after *c.* 2000 to 1800 Cal. BP on the basis of a warming in climate and increased Gulf Stream strength;

RSLC-III. A transition to a period of diminished relative sea-level rise around *c.* 1000 to 800 Cal. BP in response to a general deterioration of climate, perhaps with evidence of increased storm activity resulting from climate extremes;

RSLC-IV. A possible acceleration in the rate of relative sea-level rise around *c.* 500 to 300 Cal. BP in response to a shift in oceanic circulation and a brief warming indicated by terrestrial records. It should be noted that the duration of these climate and circulation changes could be too short to produce an identifiable sea-level signal. This phase is still included since an acceleration in relative sea-level rise is evident in the American records;

RSLC-V. A possible further acceleration in the last 150 years in response to the increased rate of temperature rise noted in the instrumental records although this change is also of short duration and has failed to be detected from tide gauge data.

This scenario of change is tested against the sea-level record from southern Britain in Chapter Seven.

2.8 SUMMARY

- The modern climate of the British Isles is strongly influenced by oceanic circulation associated with the Gulf Stream (Section 2.2.1) and atmospheric circulation associated with the North Atlantic Oscillation (Section 2.2.2).

- There is growing evidence that these two systems are closely related and that the oceans may provide the mechanism by which short period (days to months) atmospheric variability is translated into longer period (years to centuries) climate changes (Section 2.2.3).

- Instrumental records show that the northern hemisphere has warmed by around 0.5 °C since the mid 19th Century but there is no clear evidence for an acceleration in the rate of relative sea-level rise from tide gauge data (Section 2.3). This may indicate that sea level changes lag behind associated variations in climate.

- A variety of climate proxies from the northern hemisphere (ice core data, glacial evidence, tree-ring chronologies, peat bog records) and changing oceanic conditions indicate that the last 4000 Cal. years have witnessed significant climate change (Section 2.4). There is evidence for a deterioration in climate *c.* 3000 Cal. BP with a shift to wetter conditions and a period of glacier advance which is broadly consistent with the Sub-Boreal/Sub-Atlantic transition (Section 2.4.2). This persists until *c.* 1900 Cal. BP when the glacier records show shorter periods of glacial advance (Section 2.4.1.2), and the peat bog records suggest a warmer and drier phase persisting until *c.*1200 Cal. BP (Section 2.4.2). A number of the terrestrial records indicate that this warmth continued in some areas until *c.* 800 Cal. BP, after which climate became much more variable, oscillating around a common value before the final warming phase recorded by instrumental data. The apparent synchroneity between inferred changes in oceanic circulation and the terrestrial climate records provides further evidence for a close link between these two systems (Section 2.4.4).

- Studies of late Holocene relative sea-level change in Europe are frustrated by a lack of suitable datable material from the last 2000 to 2500 years (Section 2.5.1.1). This is compounded by the errors associated with age and altitude that are often of comparable magnitude to the changes under investigation (Section 2.5.1.1). The

European sea-level records indicate that the periods around *c.* 2800 Cal. BP and *c.* 800 Cal. BP were times of widespread changes in coastal deposition (Section 2.5.1.3).

- Coastal deposits from the Atlantic coast of North America are much richer in organic material than their European counterparts. This makes them better suited to the production of high resolution, radiocarbon-based chronologies of relative sea-level change (Section 2.5.2.1). A number of high resolution, foraminiferal-based sea-level studies have indicated a series of changes in the rate of late Holocene relative sea-level rise (Sections 2.5.2.1 & 2.5.2.2). A general deceleration is apparent between *c.* 4000 Cal. BP and *c.* 3000 Cal. BP, whilst accelerations at *c.*1800 Cal. BP, *c.* 800 Cal. BP and possibly *c.* 400 Cal. BP have been recorded from a number of sites (Section 2.5.2.3).

- A comparison of late Holocene climate change and relative sea-level records from American and European North Atlantic coasts indicates that the principal phases of widely recognised relative sea-level change (*c.* 3000 Cal. BP, *c.* 1800 Cal. BP, *c.* 800 Cal. BP) are closely correlated with shifts in climate and alterations in oceanic circulation (Section 2.6).

- A simplified climate record is used to generate a scenario of expected late Holocene relative sea-level change against which the UK records produced in Chapter Six can be compared (Section 2.7). This suggests five phases of predicted relative sea-level change (*RSLC-I* to *RSLC-V*) consisting of two possible decelerations at *c.* 3000 Cal. BP and *c.* 1000 to 800 Cal. BP, and three possible accelerations at *c.* 2000 to 1800 Cal. BP, *c.* 500 to 300 Cal. BP, and in the last 150 years.

Thesis Methodology

This research combines the age-altitude methodology commonly employed in UK sea-level studies with the recently developed foraminiferal-based techniques pioneered in North America (Section 2.5.2.2) to develop a high resolution record of late Holocene relative sea-level change in southern Britain. The biggest problem apparent in existing studies of late Holocene sea-level change is the size of age and altitude errors since these are commonly comparable to the period and magnitude of the variations under investigation (Section 2.5.1.3). The methodology employed in this thesis seeks to minimise these errors by careful site selection and the adoption of the highest resolution techniques currently available. This chapter consists of the following:

- A description of the site selection process employed in this thesis;
- The sampling design used to investigate the chosen sites;
- A description of the lithostratigraphic techniques employed and their associated limitations;
- A review of saltmarsh foraminifera and their application to sea-level reconstructions, followed by a description of the methodology employed in this research;
- A discussion of the chronostratigraphic techniques used, and their associated limitations and errors.

3.1 SITE SELECTION

Late Holocene sea-level studies require sensitive sites that possess high resolution records of relative sea-level changes from the last 3000 years. It is also desirable to limit the influence of factors such as changes in tidal range, coastal configuration, sediment supply, sediment compaction, and human interference that complicate the record of relative sea-level change produced. For these reasons a preliminary search of the UK was undertaken for saltmarsh sites possessing the following characteristics:

1. Thin sequences of shallow (late Holocene), intercalated organic and minerogenic sediments, preferably including thin basal peats overlying a bedrock surface at a range of altitudes;

2. Located in an low energy environment, with low tidal range and relatively open communication to the sea;

3. An absence of disruptive human activities such as embanking and reclamation.

The first characteristic is desirable since it will maximise the potential for producing a range of uncompacted, high resolution sea-level index points that can be used to precisely document vertical changes in relative sea-level.

The second feature was selected to minimise changes in marsh configuration and tidal range. The indicative meaning of a sea-level indicator is expressed with reference to the tidal frame. Consequently changes in tidal range during the time period under consideration have the potential of introducing altitudinal errors into sea-level reconstructions. Such changes are likely to be less significant in areas with low tidal ranges. This is particularly important when using sea-level indicators related to the upper limits of tidal influence (Orford, 1987), such as saltmarsh foraminifera (see Section 3.4). Reed (1995) suggests that increased tidal flooding resulting from sea-level rise will be most pronounced in areas with small tidal ranges. It is possible therefore that such areas may be more sensitive to low magnitude sea-level changes. High tidal ranges produce much larger marshes with sizeable creek systems. This leads to greater heterogeneity in surface conditions, complicating the distribution of sea-level indicators which may in turn hinder accurate determination of their indicative meaning. In addition, the higher energy of these environments means that they will be more susceptible to erosion and changes in coastal configuration.

The third attribute was selected to minimise the occurrence of 'artificial' changes in relative sea-level produced by modification of the coastal environment. Whilst human activity in the coastal zone can be used to study late Holocene relative sea-level change (e.g. Allen, 1987; Allen & Rae, 1987) it may also remove portions of the sedimentary record, thereby destroying the natural sequence of relative sea-level change.

It was recognised that whilst an ideal site would possess all these characteristics, in reality certain criteria might have to be compromised. Maps, tide tables and published data were used to generate a list of potential sites worth further investigation. The map in Figure 3.1 shows the areas where reconnaissance coring was undertaken. The sites in Scotland were rejected as there was no evidence of sedimentary sequences with the desired characteristics. The east coast sites were either rejected on stratigraphic grounds, or owing to the considerable evidence for human disturbance of the saltmarsh environment. Reconnaissance coring of sites along the south coast and in south Wales produced promising results and the following systems were selected for full-scale investigation:

1. Poole Harbour, Dorset;

2. Southampton Water, Hampshire;

3. Loughor Estuary, South Glamorgan.

Poole Harbour is primarily recommended by its low tidal range which, at 1.5 m on spring tides, is one of the smallest in the country. The system contains many low energy, fine grained saltmarsh deposits with evidence of high altitude basal peats overlain by intercalated sequences of organic and minerogenic sediments. Full details of this site are presented in Section 4.1.

Southampton Water possesses a number of saltmarsh sites with high altitude organic deposits, intercalated with minerogenic sediments (Long & Scaife, 1996). The tidal range is moderate (4.0 m on spring tides) resulting in small, fairly uniform marshes characterised by fine grained sediments. Sites in the upper estuary may have experienced some change in tidal conditions as a consequence of variations in estuary configuration, although there is no evidence to suggest this will have been significant in the last 3000 years. Full details of this site are presented in Section 4.2.

The Loughor Estuary contains wide expanses of saltmarshes, particularly along its southern shore. Whilst the tidal range is the highest of all areas included in this study (8.1 m on spring tides), it also contains the most organic rich sequences of all the sites investigated and therefore provides the opportunity to generate a number of late Holocene sea-level index points.

3.2 SAMPLING DESIGN

Whilst selection of study sites was primarily dictated by the presence of suitable lithostratigraphy, the geographical distribution and varying site character provides additional opportunities for evaluating both the methodological approach adopted, and the quality of the sea-level data produced.

Two study marshes were examined from within Poole Harbour in an attempt to determine the response of the system as a whole to relative sea-level change. Intuitively, similar sequences of events from different marsh systems are likely to provide more reliable information on relative sea-level changes than records from single locations alone, which may be strongly biased by site specific processes. The sea-level record produced from Southampton Water, which lies 50 km east of Poole Harbour, allows the responses of two neighbouring systems to be compared. In this way, the spatial scale of significant events can be expanded further. This hierarchical approach can be extended to comparing records from the Solent region with the record from the Loughor Estuary which experiences the more forceful influence of the North Atlantic. In addition, the larger tidal range and marsh area within Loughor Estuary will provide information on how suitable these systems are to the construction of high resolution sea-level records.

3.3 LITHOSTRATIGRAPHY

The lithostratigraphy of each site was investigated *via* a series of gouge augered transects. Sediments were described in the field using the Troels-Smith scheme of stratigraphic notation (Troels-Smith, 1955). A Leica TC 1010 total station was used to level the ground surface altitude of all cores to the closest Ordnance Survey benchmark.

Possible sources of altitudinal error have been considered in detail by Shennan (1980, 1982, 1986). Errors in the measurement of stratigraphic contacts are considered to be no more than ± 0.02 m OD, and levelling to the benchmark was closed to within ± 0.02 m OD, giving a combined local altitudinal error of ± 0.04 m OD. When comparing between sites, the altitudinal errors inherent in the Ordnance Survey benchmark network itself must be considered. This gives a maximum altitudinal error for inter-site comparison of ± 0.19 m OD.

Full lithostratigraphic descriptions are presented in Appendix One along with simplified stratigraphic diagrams of each transect. Diagrams were drawn using a modified symbol set designed to emphasise transitions from minerogenic to organic sedimentation (Figure A1.1).

Selected cores were recovered for biostratigraphic analysis and dating. Material was removed using a Russian-type coring device and checked in the field for contamination, before being placed in plastic tubing and sealed in polythene. On return to the laboratory, all cores were stored below 4 °C to retard biological and chemical processes.

3.4 BIOSTRATIGRAPHY

The most commonly used biological indicators of UK sea-level change are floral. Pollen from saltmarsh and terrestrial taxa are frequently used to infer the palaeoecology of a study marsh and the area surrounding it (e.g. Godwin, 1940a; Tooley, 1978). This approach may be complemented by diatom analysis, which can be used to reconstruct sediment palaeosalinity, thereby providing additional information on the depositional environment (e.g. Shennan *et al.*, 1995). In the high resolution sea-level investigations pioneered in North America, saltmarsh foraminifera are commonly employed since it is suggested they can be used to relocate former sea-levels more accurately than other floral or faunal indicators (Scott, 1976; Scott & Medioli, 1978, 1980a, 1986). The use of saltmarsh foraminifera has been largely overlooked in UK sea-level studies but recent work indicates that marsh assemblages are closely related to tidal levels (Horton, 1997; Horton *et al.*, submitted). This study is the first to employ saltmarsh foraminifera as high resolution indicators of late Holocene relative sea-level change in the UK.

3.4.1 Saltmarsh Foraminifera and Sea-Level Research

Foraminifera are single-celled marine organisms that construct shells or 'tests' from detrital sediment grains, or through the secretion of calcium carbonate derived from seawater. Some foraminifera live in the water column (planktonic), whilst others reside on or in marine sediments, or attached to plants and rocks (benthic). Benthic foraminifera are also present in the marginal marine setting of saltmarshes, where their incorporation into, and preservation by, these sediments can be used to identify fossil marsh deposits. Useful reviews are provided by Boltovskoy & Wright (1978), Brasier (1980), and Murray (1991).

3.4.1.1 The Distribution of Saltmarsh Foraminifera

Saltmarsh foraminifera are present in temperate environments from around the globe (Scott, 1976; Scott & Medioli, 1978, 1980a; Scott *et al.*, 1980; Scott & Martini, 1982; Petrucci *et al.*, 1983), and despite some climatic influence (Scott *et al.*, 1990), the general composition of assemblages is remarkably consistent (Scott & Medioli, 1980a). Most early work was concentrated in North America, with notable contributions by Phleger (1965a, 1965b, 1967, 1970). The zonation of saltmarsh foraminifera, and its similarity to observed floral zonations, led early workers to consider salinity as the dominant controlling environmental variable (Parker & Athearn, 1959). Indeed, Murray (1971) classified foraminiferal assemblages on the basis of the marsh salinity regime noting that, whilst distinct, the assemblages possessed many species common to all.

Scott (1976) observed that foraminiferal assemblages from four different study marshes in California displayed pronounced variations with elevation. Scott & Medioli (1978) demonstrated that this was not a local phenomenon by comparing the Californian data with saltmarsh assemblages observed in the cool and humid environment of Nova Scotia. This led to the conclusion that the foraminifera characterising the high marsh environment, and restricted to it, were controlled by absolute elevation above mean sea level. Thus the presence of these foraminifera in fossil saltmarsh sediments indicated deposition in the upper portion of the tidal range, regardless of marsh conditions. Of greatest significance was the marked reduction in foraminiferal abundance towards the upper limit of marine influence. Scott & Medioli (1978) proposed that this boundary could locate highest high water with a precision of ± 5 cm. This seminal work led to a more detailed treatment where five sites from different locations within Chezzetcook Inlet were studied to determine the reliability of this relationship (Scott & Medioli, 1980a). This demonstrated that the foraminiferal assemblages were remarkably consistent despite different salinities, tidal ranges and climates. The high marsh fauna showed little variation, and the dramatic decline in foraminiferal abundance towards the upper limit of marine influence was observed in all study marshes. The influence of salinity was most pronounced on the low marsh and mudflat fauna, that were found to be more closely related to conditions within the estuary itself. Hence whilst the highest high marsh zone IA was consistent in composition and elevation with respect to the tidal frame, the lower assemblages that characterised zone II were more variable (Figure 3.2).

Recently, De Rijk (1995a, 1995b), and De Rijk & Troelstra (1997) have provided a caveat to the relationship between foraminiferal assemblages and elevation with respect to the tidal frame. A survey of the Great Marshes, Barnstable, Massachusetts, revealed that salinity did not vary consistently with altitude, and as a consequence foraminiferal assemblages were not vertically zoned. Instead the large spatial scale of the marsh system (13.6 km^2) coupled with its irregular surface topography, resulted in substantial variability in the balance between freshwater supply, tidal inundation, seepage and evaporation, producing an inconsistent relationship between elevation and salinity.

3.4.1.2 The Application of Foraminiferal Zonation to Sea-Level Research

The tenet underpinning the use of saltmarsh foraminifera in sea-level research is that, whilst salinity may alter the composition of the foraminiferal assemblage, an unambiguous and quantifiable vertical zonation will still be present. Foraminifera clearly have great potential as precise sea-level indicators, but their vertical relationship to sea-level is indirect. The results of De Rijk (1995a, 1995b) and De Rijk & Troelstra (1997) show salinity to be the dominant environmental control, and any vertical zonation remains an artefact of elevational changes in salinity. It is important therefore that the modern distribution of foraminifera at a study site is investigated to confirm the presence of a vertical zonation before interpretations from fossil material are made.

The work of Horton (1997) provides evidence that vertically zoned saltmarsh foraminiferal assemblages do exist in the UK (Figure 3.3). At Cowpen Marsh (Tees Estuary), fortnightly samples were taken during the course of a year to investigate the influence of seasonality on foraminiferal assemblages (Horton, 1997). When averaged over the course of a year, altitude can account for 87% of the total variation in the death assemblage. However, this study reveals that the distribution of the death foraminiferal assemblage varies seasonally so that the results of single transects may under-estimate or over-estimate the altitudinal range of a zone when averaged over the course of a year. Spatial variability is noted by Gehrels (1994) in a study of modern foraminiferal distributions at four saltmarshes in Maine, USA. He notes that the vertical range of similar assemblage zones, and the elevations of faunal boundaries relative to the tidal frame, vary between sites and even between transects from the same site.

Such inter-site and intra-site variability can introduce vertical errors into sea-level reconstructions. Intuitively, spatial variability as described by Gehrels (1994) will be greater in marshes covering large areas. Also, these large marshes are likely to experience higher energy regimes, therefore increasing the chances of substantial changes in marsh configuration. This adds further support to the selection of small saltmarshes from areas of low tidal range as suggested in Section 3.1. Whilst it is beyond the scope of most sea-level investigations to sample multiple transects at various times during the year, one potential solution to the problem of spatial and temporal variability is to combine contemporary data derived from a number of different marshes each displaying a vertical zonation. Incorporation of this combined dataset with samples from the vertically zoned study marsh will reduce errors associated with unquantified seasonal and spatial variability. This approach is in direct contrast to that advocated by van de Plassche (1986) and Gehrels (1994) who suggest that modern, site specific studies close to the source of fossil material will improve the resolution of palaeoenvironmental reconstructions. Whilst such an approach increases the precision of sea-level reconstructions, it may significantly reduce their accuracy. For this reason, the contemporary foraminiferal data collected as part of this study are combined with the results of similar investigations conducted on other saltmarshes in the UK to produce a large modern training set (Chapter Five).

3.4.1.3 A Question of Life or Death

The effect of seasonal changes on foraminiferal abundance and distribution is not well understood. It is suggested that calcareous species display greater seasonal variability, although this may be an artefact of test dissolution (Parker & Athearn, 1959). It is recognised that the number of living foraminifera recorded at a site will vary spatially to a greater extent than the dead component of the assemblage (Scott & Medioli, 1980b;

Horton, in press). Scott (1976) observes significant seasonal variation in the living populations of saltmarsh foraminifera, but suggests that the total percentage occurrence do not change significantly (Scott & Medioli, 1980b).

This variability in living foraminiferal abundance has led to some controversy over which component of the assemblage should be used in palaeoecological reconstruction. Murray (1973) maintains that only living populations should be related to observed environmental parameters, and ideally long term seasonal studies should be used (Buzas, 1968). The variable nature of the life assemblage however, has led others to suggest that the total assemblage is a better indicator of general environmental conditions (Albani & Johnson, 1975; Scott & Medioli, 1980b).

De Rijk (1995a) emphasises the different requirements of investigations seeking to study the ecology and taphonomy of foraminifera compared with those that utilise foraminifera as palaeoenvironmental indicators. Fossil foraminiferal assemblages differ fundamentally from their contemporary counterparts in that they are, by definition, death assemblages and are also subject to post depositional modification. Such changes comprise the removal of tests by erosion, corrosion or dissolution, and the addition of tests by transport and deposition or contamination. The relative importance of these factors will vary according to the composition of the original assemblage and the environmental conditions encountered by the foraminiferal tests after their deposition. In general, the low energy environment of a saltmarsh has little capacity to remove or deposit foraminifera (De Rijk, 1995a), but dissolution of calcareous foraminifera will occur where pH falls (Jonasson & Patterson, 1992; Alve & Murray, 1994). Conversely, more marine settings will preserve calcareous forms but these assemblages are more prone to reworking by wave and tidal forces.

In his twelve month study of Cowpen Marsh, Horton (1997) compared contemporary foraminiferal assemblages with sub-surface assemblages below the zone of infaunal activity. He concludes that the modern death assemblage accurately represents the sub-surface assemblage and is less susceptible to seasonal variability. On this basis Horton (1997) recommends the use of death foraminiferal assemblages in palaeoenvironmental reconstructions, and consequently these assemblages are used throughout this thesis unless specifically mentioned to the contrary.

3.4.1.4 UK Saltmarsh Foraminifera

Whilst a number of studies have been conducted on inter-tidal foraminifera in the UK (e.g. Murray, 1968, 1980; Alve & Murray, 1994), there is a paucity of detailed information regarding saltmarsh distribution. As noted above, the use of saltmarsh foraminifera in UK sea-level investigations is notably rare. Adams & Haynes (1965) used foraminifera to identify periods of marine deposition in sediments from beneath Borth Bog, Wales, and Huddart (1992) applies a similar approach in his analysis of Downholland Moss, Lancashire. No contemporary surveys were made in either case, and the occurrence of foraminifera was merely used to confirm the marine origin of sediments. No detailed interpretation of water level changes indicated by the variations in foraminiferal assemblage were attempted. Similar general studies have been conducted in Romney Marsh where foraminifera were used to distinguish saltmarsh, inter-tidal mudflat, and sub-tidal deposits (Waller *et al.*, 1988; Wass, 1995). The saltmarsh deposits were not sub-divided on the basis of their foraminiferal contents however, and inferred changes were restricted to broad scale transitions between these different littoral environments.

Coles (1977) and Coles & Funnell (1981) conducted a contemporary study in Broadland, UK and identified a similar vertical zonation to Scott & Medioli (1978, 1980a). The high marsh zone was dominated by *J. macrescens*, *T. inflata* and some *Haplophragmoides* species, whilst the tidal flat fauna consisted of a more diverse calcareous assemblage including *A. beccarii*, *E. excavatum*, and *H. germanica*. These data were later employed by Godwin (1993) in a semi-quantitative palaeoenvironmental reconstruction although the chronology of change was derived from existing radiocarbon dates.

Horton (1997) collected detailed foraminiferal data from a number of sites in the UK that were used to develop a quantitative technique for determining indicative meanings from fossil assemblages (Section 3.4.1.2). This was then applied to 35 previously collected Holocene sea-level index points from the east coast of the UK.

3.4.2 Sampling Design

The aims of the foraminiferal analyses conducted in this thesis are to identify, where possible, vertically zoned contemporary assemblages, and to quantify their relationship to the tidal frame. Application of this relationship to fossil foraminiferal assemblages will then permit accurate determination of the sample's indicative meaning.

To realise these aims, two types of foraminiferal analysis are conducted. The first focuses on quantifying the relationship between modern foraminifera assemblages and a reference tidal level. This is achieved by the construction of a transfer function, and this process is described in Chapter Five. The second set of analyses is directed to investigate fossil foraminiferal assemblages and employs three sampling strategies. The first endeavours to construct a broad picture of depositional changes across the site as a whole, in order to identify potential marsh-wide submergence or emergence events, reflecting changes in marine influence. This is achieved by analysing complete cores, sampling at intervals of 12 cm or less. The second aim is to assign indicative meanings to material removed for dating, and thereby enable development of high resolution sea-level index points to reconstruct vertical movements is relative sea-level. This is achieved by examining contiguous samples across the dated level. The final strategy aims to identify the first occurrence of marine conditions recorded in cores possessing basal peat. Contiguous samples are taken across the lithostratigraphic transgressive contact to pinpoint the arrival of high marsh foraminifera. This material can then be sampled for dating purposes on the basis of its foraminiferal content. The methods of sample collection and preparation are described in Appendix Two. The results of fossil foraminiferal analysis are described in Chapter Four and Appendix Two.

3.4.3. Foraminiferal Sample Size

It is generally agreed that the optimum count for statistical significance is between 300 to 400 specimens (Moore *et al.*, 1991; Murray, 1991). British saltmarsh foraminifera are typically found in extremely low numbers particularly toward the upper limit of marine influence, and yet it is these assemblage zones which frequently possess the most precise relationship with tidal levels. Physical limitations in sample size (particular when dealing with fossil material that is also required for dating) and time restrictions mean that whilst 300 specimens are counted wherever possible, there are many samples which have far fewer test numbers. Rejection of such samples removes a large proportion of the data and severely restricts the resolution of the sea-level reconstructions. Clearly a balance must be reached between the removal of potentially useful data and the inclusion of erroneous and misleading results.

Taxa	Aliquot								
	I	II	III	IV	V	VI	VII	VIII	Mean
Jm	17	21	35	28	32	27	34	27	28
Mf	6	10	9	11	15	11	11	13	11
HA	1	1	3	2	4	5	3	2	3
Ti	1	0	0	0	0	0	0	0	0
Total	25	32	47	41	51	43	48	42	41
Percentage Frequencies									
Jm	68%	66%	74%	68%	63%	63%	71%	64%	67%
Mf	24%	31%	19%	27%	29%	26%	23%	31%	26%
HA	4%	3%	6%	5%	8%	12%	6%	5%	6%
Ti	4%	0%	0%	0%	0%	0%	0%	0%	0%

Table 3.1 *Results of the Wetsplitter Test.* Jm = Jadammina macrescens, Mf = Miliammina fusca, HA = Haplophragmoides species A, Ti = Trochammina inflata.

To investigate the reliability of low counts, a typical low abundance high marsh sample was divided into eight aliquots using a wetsplitter (Scott & Hermelin, 1993). These were analysed and the results are presented in Table 3.1. It is apparent that whilst the exact percentage frequencies of individual taxa vary, there is no significant change in the dominant species present. Low abundance fossil counts are therefore used in this thesis, although these are treated with caution in the absence of supporting evidence. A detailed discussion of such evidence is presented in Section 5.8.1.

3.5 CHRONOSTRATIGRAPHY

Two types of dating control are used in this research. Primary dating is provided by the use of radiocarbon assays. Conventional samples were taken as range finders where possible, whilst detailed investigation used AMS dating methods. Sample material consisted of freshwater peat or organic saltmarsh sediments. Foraminifera were used to determine the indicative meaning of these sediment samples, but were not themselves the subject of dating. Supplementary age information is provided by identification of the dated rise in *Pinus* pollen from the sediments of Poole Harbour and Southampton Water, and where possible from the occurrence of *Spartina* macrofossils (Section 3.5.2).

3.5.1 Radiocarbon Dates

Twenty-one new radiocarbon dates are presented in Appendix Three, in association with a further sixty-five used in developing site chronostratigraphy.

A comprehensive review of the sources of error associated with radiocarbon dating can be found in Lowe & Walker (1984), Mook & van de Plassche (1986), and Grove & Switsur (1994). There are two major sources of error which may be significant in this research. A 'hard-water' effect can arise when water containing dissolved, old carbon is utilised by saltmarsh plants, engendering an apparent age to the living organism. The headwaters of the rivers draining into Poole Harbour and Southampton Water are located within Cretaceous chalk uplands and may therefore instil a hard-water effect on samples from these sites.

Contamination may also result from incorporation of fossil organic material during sediment deposition, or *via* rootlet penetration and organic infiltration. The former will be most significant in low marsh to mudflat samples where the organic content is low and the proportion of allochthonous material is greater than in a higher marsh context. Incorporation of this type of material will increase the apparent age of a sample. In this way, it is possible that the lignitic clays of the Poole Formation found within Poole Harbour, and the Coal Measures beneath the Loughor estuary, may be sources of old carbon. Conversely, rootlet penetration and organic infiltration will be most prominent in high marsh contexts and will have the effect of rejuvenating sample material. At Keysworth Marsh (Poole Harbour, Dorset), *Spartina* rhizomes have been reported from over a metre below the surface (Hubbard & Stebbings, 1968), and Kaye & Barghoorn (1964) noted the rootlets of *Spartina* growing in Boston, Massachusetts, penetrated over 30 cm below its rhizomes. Material used to investigate late Holocene relative sea-level in southern Britain rarely comes from more than 2 m below the surface, and so most samples used in this research are susceptible to contamination from rootlets. This means that all samples must be carefully inspected for the presence of rootlets which can then be removed before submission for dating. Organic infiltration is most pronounced in soils, and therefore is only likely to be significant in the highest marsh samples where they reside above the water table for prolonged periods.

Three types of radiocarbon date are utilised in this study.

Category A : AMS dating of basal transgressive contacts established from foraminiferal analysis to produce precise sea-level index points for time-altitude analysis.

Category B : AMS dating of selected intra-core variations in depositional environment revealed by combined lithostratigraphy and biostratigraphy to produce a chronology of sea-level tendencies.

Category C : Conventional dating of the deepest basal peats to constrain the time period of deposition at the site.

The age control for sea-level graphs is routinely achieved through the use of conventional radiocarbon dating. Conventional dating requires comparatively large volumes of organic material ($\geq$ 90 gm) which can equate to a sample thickness of 5 cm or more from narrow bore coring devices. Thick samples have correspondingly longer periods of formation, and therefore age estimates represent an integration of the time elapsed during sediment deposition. In high marsh contexts, sedimentation rates are usually low, and therefore substantial age errors may be introduced in this way. In addition the greater volume of material required demands collection of secondary cores for microfossil analysis, and thereby introduces errors associated with cross-correlation. Finally, the dating of sediments with low organic contents, such as those typically found in the late-Holocene sequences of the UK, is precluded (Section 2.5.1.1).

The development of AMS radiocarbon dating allows much smaller quantities of carbon to be used (< 1 gm). This reduces the errors outlined above and expands the range of environments from which dated material can be taken.

All radiocarbon data used within this thesis is presented in Chapter Four and Appendix Three, along with the calibration procedures adopted. All data are reported in Cal. yrs. BP.

3.5.2 Pollen

The pollen signature of historically introduced flora has been used to date late Holocene sediments (e.g. Schafer & Mudie, 1980; Clark & Patterson, 1984, 1985). The recent expansion of coniferous plantations across southern England produced a pronounced increase in the frequencies of *Pinus* pollen, which is evident in the data from a number of sources (e.g. Flower, 1977; Haskins, 1978; Scaife, 1980; Waton, 1983). The onset of tree planting is well documented and, although some debate exists regarding the length of time before a pollen signature was recognisable in the sedimentary record, the start of the *Pinus* rise can be confidently dated between 1700 and 1800 AD. This biological chronohorizon has been successfully employed in a study of sedimentation at Poole Harbour, Dorset (Long *et al.*, in press), and in Southampton Water, Hampshire (Long & Scaife, 1996). These pollen data are used to supplement age estimates provided by radiocarbon dating in the last 200 years of this study.

3.6 SUMMARY

- Site selection criteria and reconnaissance coring recommend the areas of Poole Harbour in Dorset, Southampton Water in Hampshire, and the Loughor Estuary in West Glamorgan for further analysis.

- By employing a hierarchical approach of site scale, local scale, and regional scale analysis, the robustness of the derived relative sea-level record can be assessed.

- The vertical zonation of saltmarsh foraminifera provides a method for precisely reconstructing the indicative meaning of fossil deposits.

- AMS radiocarbon dating permits smaller sediment samples to be dated, and together with foraminiferal analysis facilitates the construction of sea-level index points with small age and altitude errors.

The Study Sites

Chapter Three describes the criteria used to select the general areas of study in this thesis. More precise details regarding individual study sites are presented below, where the following points are considered:

- The location and character of each study area with a summary of its geological history and existing sea-level information;
- The lithostratigraphy of the individual study marshes and the fossil foraminifera contained within their sediments;
- The results of radiocarbon assays from each site.

4.1 POOLE HARBOUR

Poole Harbour is a small estuary situated at the western end of Poole Bay, Dorset (Figure 4.1). Its shallow waters, commonly less than 4 m deep at high tide (Dyrynda, 1987), cover an area of 35 km^2 from which project a number of small islands. The largest island, Brownsea, is situated at the mouth of the estuary where tidal waters from Poole Harbour exchange with those in Poole Bay *via* the 350 m wide 'Swash Channel' that interposes the double spits of South Haven Peninsula and Sandbanks. At low tide, water withdraws to a network of channels to reveal large expanses of mudflat and fringing saltmarsh.

The Harbour possesses a complex tidal character due to the proximity of an M_2 amphidrome that removes the dominance of the lunar and solar semi-diurnal tides whilst increasing the significance of the quarter-diurnal tides (M_4, MS_4). Consequently the tidal curve displays a double high water on springs and a triple high water during neaps (Green, 1940). Deformation of the tidal wave also means that water levels remain elevated for nearly 16 hours every day and at standing high water the Harbour resembles a lagoon (Bray *et al.*, 1991). The low tidal range (Table 4.1), coupled with the Harbour's enclosed nature, means that meteorological conditions can have a significant effect on its water levels. For example, strong south-westerly winds coupled with a low pressure system can elevate water levels by 0.5 m (Burkmar, 1995).

Site	Mean High Water Spring Tide (m OD)	Mean High Water Neap Tide (m OD)	Mean Low Water Neap Tide (m OD)	Mean Low Water Spring Tide (m OD)	Mean Sea Level (m OD)	Mean Tide Level (m OD)	Tidal Range Springs (m)	Tidal Range Neaps (m)
Arne Peninsula	0.60	0.30	-0.10	-0.60	0.10	0.05	1.20	0.40
Newton Bay	0.70	0.10	-0.20	-0.90	0.10	-0.08	1.60	0.30
Bury Farm	1.76	0.96	-0.94	-2.24	0.16	-0.12	4.00	1.90
Llanrhidian Marsh	4.14	2.14	-1.16	-2.96	1.39	0.54	7.10	3.30

Table 4.1 *Tide data for the four study marshes (from Admiralty Tide Tables, 1998)*

Freshwater is supplied to the system by four main rivers, the largest of these being the Frome and Piddle located at the western limit of the Harbour. Fluvial input is estimated to contribute around 2 % of the total low water volume of the harbour (Bray *et al.*, 1991), and salinities in the main body of the estuary range from *c.* 35 ‰ by the entrance to *c.* 30 ‰ north of Arne Peninsula (Green, 1940).

4.1.1 Geological History

Poole Harbour is located within an asymmetrical syncline, termed the Hampshire basin, that is filled with Upper Palaeocene to Lower Oligocene sediments, and into which a number of rivers drain from the Upper Cretaceous chalk to the north (Allen & Gibbard, 1993). The Harbour itself is underlain by Tertiary sediments of the Bracklesham Group, comprising the fine to coarse grained, cross-bedded sands and carbonaceous and lignitic clays of the Poole Formation, overlain by the Branksome Sand Formation (Bristow *et al.*, 1991).

During the Pleistocene period a major river system (termed the 'Solent River') flowed eastward from Poole Harbour through the modern day Solent area before discharging into the English Channel (Darwin-Fox, 1862; Everard, 1954; Allen & Gibbard, 1993). Extensive spreads of fluvial gravel found throughout the region are the legacy of this ancient river system, and are categorised as 'Plateau' or 'Valley' gravels depending on their altitude and relationship to the extant drainage system (Reid, 1892, 1902; Reid & Whitaker, 1902; White, 1917, 1921; Edwards & Freshney, 1987).

The entrance to Poole Harbour is believed to have been over 3.5 km in width at the start of the Holocene (Robinson, 1955). May (1969) presents the 'probable' shoreline of Poole Harbour at *c.* 6000 Cal. BP suggesting that the Sandbanks had yet to form. The accuracy of these data is equivocal since May (1969) assumes no sea-level rise since 6000 Cal. BP (see Section 4.1.3) and Diver (1933) suggests that South Haven Peninsula formed around a pre-existing outcrop of the underlying Tertiary deposits. In the last 400 years some changes in the morphology of the spits is apparent from cartographic evidence (Diver, 1933; Robinson, 1955) but the width of the harbour entrance appears to have been stable.

4.1.2 Sediment Sources

The coastal sediments of Poole Harbour vary from coarse grained sand, pebble, and cobble beaches found in more exposed areas, to fine grained silt-clay mudflats and saltmarshes located along the southern, western and northwestern shores (Bird & Ranwell, 1964). May (1969) describes a 'low bluff' *c.* 1.5 m to 6.0 m in height, often associated with beach material, that backs many areas of marshland. In some areas exposed to the influence of wave action, such as in the vicinity of Shipstal Point and northward around Arne Peninsula, erosion is still occurring. Sediment samples from the Harbour bottom consist of sand-sized material that progressively fine to the west (Green *et al.*, 1952). These deposits are similar in composition to the Tertiary and Pleistocene sediments exposed in cliff sections (Green, 1940), as indeed is much of the beach material (May, 1976).

The fine-grained sediments that comprise the extensive mudflats and saltmarshes fringing the Harbour are predominantly of fluvial origin (May, 1969). Gray (1985) suggests that fluvial sediments are deposited in the upper estuary due to the poor flushing characteristics of the system, but Bray *et al.*, (1991) note the quantitative contribution of these fluvial sources is still poorly understood.

4.1.2.1 The Significance of **Spartina anglica**

The sedimentary regime of Poole Harbour was dramatically altered at the turn of the century by the arrival of *Spartina anglica*, the newly evolved fertile form of *Spartina townsendii*. *S. townsendii*, the hybrid progeny of European *Spartina maritima* and North American *Spartina alterniflora*, is believed to have first appeared in Southampton Water around 1870 AD (Hubbard, 1965; Gray *et al.*, 1990). After its inception *c.* 1890 AD, *S. anglica* spread rapidly, reaching Poole Harbour by 1899 AD (Linton, 1906; Hubbard, 1957) and invading every estuary eastward to Chichester by 1913 AD (Goodman *et al.*, 1959).

During this initial expansion, new occurrences of *Spartina* were most evident where previously unvegetated mudflats had been colonised (Hubbard, 1965). *Spartina* appears able to exploit these zones as a consequence of its ability to withstand longer periods of tidal submergence than other marsh plants, such as *Zostera* that previously dominated the flora of Poole Harbour (Ranwell *et al.*, 1964). The colonisation of large areas of previously unvegetated mudflat resulted in a phase of saltmarsh expansion and increased rates of sediment accretion (Ranwell *et al.*, 1964). At Keysworth Marsh, a subsurface investigation of *Spartina* remains indicated that at least 160 cm of sediment had accreted in the 50 years since it first became established (Hubbard & Stebbings, 1968). This enhanced capacity for sediment storage was reflected in a deepening of the main channels within the Harbour until the 1920s AD (Gray *et al.*, 1990).

The first evidence of a decline in the spread of *Spartina* came in the early 1920s AD when drifting plants and marsh erosion were noted (Oliver, 1920; Oliver 1923-29). The acreage of *Spartina* was reduced by around 20 % between the period 1924 AD and 1952 AD (Hubbard, 1965), and from the 1930s AD onward, the deepened channels began to fill up with sediment once more (Gray *et al.*, 1990). The degeneration of *Spartina*, termed 'die-back' (Goodman, 1957), is now a widespread phenomena around Poole Harbour, and has been noted in Arne Bay (Goodman, *et al.*, 1959), located close to the first study marsh, and at the second study marsh in Newton Bay (Hubbard, 1965). Die-back although widespread, is not ubiquitous in Poole Harbour, and in some areas *Spartina* stands are still spreading (Bray *et al.*, 1991).

4.1.3 Existing Sea-level Data

Considering the abundance and quality of saltmarsh sequences found in Poole Harbour, there is a remarkable absence of any detailed information on sea-level change (Long, in press). The majority of data have been collected in association with botanical or archaeological evidence. Godwin *et al.* (1958) present a single radiocarbon date, taken from a freshwater peat at -12.8 m OD, overlain by marine sediments at -11.58 m OD. The age of 8410 to 7960 Cal. BP predates the subsequent marine incursion of the system.

Pollen evidence of changing plant communities has been used to infer relative sea-level movements. Haskins (1978) speculates that the replacement of alder peat with *Phragmites* peat at two separate locations in Poole Harbour could be evidence of rising relative sea-level during the Neolithic (*c.* 5200 Cal. BP to *c.* 4000 Cal. BP). Also, Haskins (1978) and Sutherland (1984) infer a transgression during the middle to late Iron Age (*c.* 2400 to 2000 Cal. BP), at a time when similar inundations of freshwater peat are recorded from many coastal sites around the UK (Fulford & Champion, 1997).

Long *et al.*, (in press) report the only reliable sea-level index point available from Poole Harbour. This comes from the transgressive contact of a basal freshwater peat found at -1.15 m OD on Arne Peninsula, and dated to between 3460 and 2810 Cal. BP. In addition, Long *et al.*, (in press) use a dated rise in *Pinus* pollen in association with the presence of *Spartina* macrofossils to infer variation in the rate of late Holocene relative sea-level rise during the last 200 years. On this basis, the authors suggest that since 3460 to 2810 Cal. BP, relative sea-level at this site has risen by around 1.99 m, at a long-term rate of *c.* 0.59 mm a^{-1}. Between *c.* 200 Cal. BP and *c.* 60 Cal. BP sediment accretion rates rose from 0.28 mm a^{-1} to 1.14 mm a^{-1}, and increased further to 7.17 mm a^{-1} after *c.* 60 Cal. BP, following the establishment of *Spartina anglica.*

Variations in relative sea-level during the Roman Period (*c.* 2000 to 1600 Cal. BP) have been inferred from archaeological evidence. Jarvis (1992) suggests a rise in relative sea-level of at least 2.67 m since *c.* 1700 to 1600 Cal. BP based on the excavation of a freshwater peat-filled ditch capped by sand, and currently resting in the inter-tidal zone of Brownsea Island. A similar magnitude of relative sea-level rise is implied by excavations at Hengitsbury Head (Cunliffe, 1987). If these data are correct, it indicates relative sea-level must have fallen by around 0.7 m at some time between *c*. 3000 Cal. BP and *c*. 1700 Cal. BP.

Short-lived radionuclide investigations (^{210}Pb, ^{137}Cs, $^{239,240}Pu$, ^{241}Am, ^{60}Co) of various marshes in the Solent region, including Poole Harbour, indicate a general accretion rate of 4 to 5 mm a^{-1} (Cundy & Croudace, 1996). This compares to a long term-rate of relative sea-level rise of between 1 and 1.5 mm a^{-1}, inferred from geological evidence (Long & Tooley, 1995, Long *et al.*, in press). Cundy & Croudace (1996) suggest this is evidence of a recent acceleration in the rate of relative sea-level rise, which is corroborated by the Portsmouth tide gauge record, that shows a rate of 5 ± 0.5 mm a^{-1} (Woodworth, 1987). The cited tide gauge record is based on a time series of only 25 years however, and indicates elevated rates of sea-level rise relative to other tide gauge stations (Emery & Aubery, 1985). In addition, measured rates of sediment accretion show considerable variability depending on the radionuclide used, with values ranging from less than 1 mm a^{-1} to around 10 mm a^{-1} (Cundy & Croudace, 1996). The biggest problem with inferences about relative sea-level based using short-lived radionuclide data is that these studies are recording the dramatic acceleration in sedimentation rates that occurred after the establishment of *S. anglica*. The use of sedimentation rate as a proxy for the rate of relative sea-level rise is only valid when the sedimentary system is in equilibrium with rising water levels. The dramatic expansion of saltmarshes throughout the Solent region in response to the arrival of *S. anglica* is clear evidence of disequilibrium during this time.

4.1.4 Arne Peninsula

The study marsh forms part of the Arne Nature Reserve, managed by the Royal Society for the Protection of Birds, and as such has remained largely undisturbed by recent human activities (Plate 4.1, Figure 4.2). A well developed floral succession is apparent, grading from *Alnus*, *Quercus* and *Salix* woodland with *Sphagnum* communities at the rear of the marsh, through a dense sward of *Scirpus maritimus*, down to the main marsh area colonised by *Halimione portulascoides*, *Spartina anglica, Spergularia marina, Suaeda maritima* and *Aster tripolium* (Long *et al.*, in press). The study marsh is protected from wave action by Round and Long Islands to the west and Shipstal Point to the north, where a small sand beach is fed by cliff erosion (May, 1976). The saltmarsh is around 130 m wide, and slopes gently seaward from around +1.0 m OD to +0.4 m OD, where a small cliff around *c.* 0.2 m high marks the transition from marsh to mudflat (Figure 4.2). The marsh surface is occasionally dissected by shallow tidal creeks not more than 0.5 m deep.

4.1.4.1 Lithostratigraphy

The lithostratigraphic investigation at Arne consists of three transects comprising a total of 40 boreholes. Tabulated sediment descriptions and accompanying figures can be found in Appendix One. A summary transect is presented in Figure 4.3, showing all the major stratigraphic units recorded. The pre-Holocene surface slopes gradually seaward from +0.6 m OD to -0.5 m OD whereupon there is a stepped drop to -1.2 m OD, after which the progressive seaward slope resumes. Well humified basal peat rests upon Tertiary bedrock seaward of this break in slope, below *c.* -1.0 m OD. This unit is capped by sand, silt and clay, the nature of the upper contact varying from gradational to erosive. The sediments between -0.5 m OD and around 0 m OD are more uniformly distributed across the marsh, comprising silt-clay intercalated with sediments containing *turfa* and some humified organic matter. A laterally persistent, organic clay with some detrital herbaceous material is found between -0.5 m OD and -0.2 m OD, and grades into the overlying inorganic silt-clay. The upper component of the stratigraphy is up to 80 cm in thickness and comprises a wet clay with abundant *Spartina* macrofossils.

4.1.4.2 Biostratigraphy

The fossil foraminiferal investigation at Arne comprises 192 samples from 13 boreholes, and a total count of *c.* 32000 specimens. Full results, which are presented diagramatically and in tabular form, can be found in Appendix Two. A summary foraminiferal diagram from core ARN1-95-90 is shown in Figure 4.4, and possesses all the major biostratigraphic units identified. This is visually divided into four assemblage zones to aid description.

Zone I is an almost monospecific assemblage of *Jadammina macrescens*, with lesser numbers of *Balticammina pseudomacrescens*, *Haplophragmoides Sp. A*, *Miliammina* *fusca*, and *Trochammina inflata*, individually contributing less than 10 % of the assemblage.

Zone II possesses an assemblage dominated by *M. fusca*, with low abundances of *Ammoscalaria runiana*, and *J. macrescens*, and large numbers of foraminiferal test linings.

Zone III contains a low abundance assemblage of *Reophax* species, some *M. fusca* (up to 50 %), a low abundance of *J. macrescens*, and high numbers of foraminiferal test linings.

Zone IV is characterised by a dominance of *J. macrescens* with some *M. fusca* and *T. inflata*. At the base of this zone, *Textularia* species are also present, contributing up to 14 % of the assemblage.

4.1.4.3 Chronostratigraphy

The chronostratigraphy at Arne consists of 4 AMS and 1 conventional radiocarbon date collected as part of this study in addition to the single conventional radiocarbon date described in Section 4.1.3. The results of these radiocarbon assays are summarised in Table 4.2, whilst full details are presented in Appendix Three. The ages of all available radiocarbon dates are given in Cal. BP on the summary stratigraphy of Figure 4.3.

Sample	Publication Code	Source Core	Altitude (m OD)	^{14}C Enrichment (% Modern ± 1σ)	Carbon Content (% by Weight)	$\delta^{13}C_{PDB}$ ± 0.1%	^{14}C yr BP ± 1σ	Cal. Year Range (2σ)
ARN#1	CAMS-41300	ARN1-96-8	+0.48	96.25 ± 0.45	18.8	-25.7	310 ± 40	290 - 470
ARN#2	CAMS-40117	ARN1-95-80	-0.29	72.10 ± 0.48	1.2	-20.4	2630 ± 60	2550 - 2849
ARN#3	CAMS-40116	ARN1-95-80	-0.57	85.10 ± 0.52	6.3	-17.4	1300 ± 50	1080 - 1300
ARN#4	CAMS-40115	ARN1-95-80	-1.04	83.21 ± 0.51	12.0	-28.7	1480 ± 50	1290 - 1500
ARN#5	BETA-87925	ARN1-95-90	-1.09	-	-	-29.9	3130 ± 60	2810 - 3460
ARN#6	BETA-81279	ARN1-95-130	-2.30 to -2.35	-	-	-28.0	3720 ± 50	3940 - 4280

Table 4.2 ***A summary of the radiocarbon data from Arne Peninsula***

4.1.5 Newton Bay

Newton Bay is located in the southeastern part of Poole Harbour, between Cleavel and Goathorn Points, and is *c.* 3 km from Arne Peninsula (Figure 4.1). The area is largely protected by Brownsea, Furzey, and Green Islands although a sandy beach is evident on the more exposed, western shore. The study marsh at Newton Bay (Figure 4.5) forms part of the privately owned Rempstone Estate, and public access to the site is prohibited. The Bay is fringed by saltmarsh which is currently experiencing erosion as a consequence of *Spartina* die-back (Plate 4.2). The saltmarsh vegetation is similar in composition to that at Arne Peninsula but the vegetated marsh is not more than 50 m in width. The marsh resides between +0.9 m OD and +0.3 m OD and is backed by a small bluff about 1.5 m in height, similar to those described by May (1969). This area is wooded, and the sediments in the back marsh are dominated by detrital wood and leaf litter.

4.1.5.1 Lithostratigraphy

The lithostratigraphic investigation at Newton Bay consists of three transects comprising a total of 27 boreholes. Tabulated sediment descriptions and accompanying figures can be found in Appendix One. A summary transect is presented in Figure 4.6, showing all the major stratigraphic units recorded. Whilst the slope of the pre-Holocene surface is steeper than at Arne it can be seen that similarities in the general form exist. The Tertiary bedrock slopes seaward from +0.2 m OD to -0.6 m OD where a vertical displacement of *c.* 60 cm occurs, before the progressive seaward slope returns below -1.0 m OD. Once again, a well-humified basal peat is observed seaward of this step below -1.0 m OD and is capped, unconformably in places, by sand and clay-silt. Between -1.0 m OD and 0 m OD, the stratigraphy is dominated by inorganic sediments, including substantial sand layers. The composition of the upper 60 to 80 cm of sediment varies laterally across the marsh. A thick, dark black peat occurs in back marsh areas, although this is very different in nature to the basal peat referred to previously: it is dominated by detrital wood and leaf litter, possesses little *turfa*, and is similar in composition to the modern backmarsh sediments. This unit is restricted to a 10 m wide zone before grading laterally into an organic clay-silt similar in composition to the *Spartina* sediments of Arne Peninsula.

4.1.5.2 Biostratigraphy

The fossil foraminiferal investigation at Newton Bay comprises 142 samples from 8 boreholes, and a total count of *c.* 12500 specimens. Full results, which are presented diagramatically and in tabular form, can be found in Appendix Two. A summary foraminiferal diagram from core NEB2-96-60 is shown in Figure 4.7, and possesses all the major biostratigraphic units identified. This is visually divided into five assemblage zones to aid description.

Zone I is characterised by a very low abundance assemblage dominated by *Jadammina macrescens*, with occasional *Haplophragmoides* species and *Miliammina fusca*.

Zone II is also dominated by *J. macrescens*, although *M. fusca* rises to prominence above 130 cm depth. Fairly consistent abundances of *Ammoscalaria runiana* and *Trochammina inflata* occur throughout, although these never contribute more than 30 % of the total assemblage. Large numbers of foraminiferal test linings are also present.

Zone III is virtually devoid of foraminifera, and no counts above 2 were recorded.

Zone IV comprises a *M. fusca* dominated assemblage with low numbers of *A. runiana*, *Eggerelloides scaber*, *J. macrescens* and *Reophax* species.

Zone V contains a mixed assemblage of *J. macrescens*, *M. fusca*, and *T. inflata*, with some *Textularia* species present at the base of the zone.

4.1.5.3 Chronostratigraphy

The chronostratigraphy at Newton Bay consists of 6 AMS radiocarbon dates. The results of these radiocarbon assays are summarised in Table 4.3, whilst full details are presented in Appendix Three. The ages of all available radiocarbon dates are given in Cal. BP on the summary stratigraphy of Figure 4.6.

Sample	Publication Code	Source Core	Altitude (m OD)	^{14}C Enrichment (% Modern ± 1σ)	Carbon Content (% by Weight)	$\delta^{13}C_{PDB}$ ± 0.1%	^{14}C yr BP ± 1σ	Cal. Year Range (2σ)
NEB#1	CAMS-41301	NEB2-96-0	+0.46	97.90 ± 0.56	33.1	-28.7	170 ± 50	0 - 300
NEB#2	CAMS-41303	NEB1-96-10	+0.14	88.39 ± 0.50	2.5	-26.8	990 ± 50	780 - 970
NEB#3	CAMS-41304	NEB3-96-70	-0.15	82.58 ± 0.50	5.1	-16.2	1540 ± 50	1310 - 1530
NEB#4	CAMS-41302	NEB1-96-10	-0.24	74.85 ± 0.44	12.5	-28.5	2330 ± 50	2190 - 2430
NEB#5	CAMS-40113	NEB2-96-20	-0.69	65.67 ± 0.46	12.5	-28.3	3380 ± 60	3470 - 3820
NEB#6	CAMS-40112	NEB2-96-40	-1.15	58.97 ± 0.48	20.9	-28.0	4240 ± 70	4560 - 4970

Table 4.3 ***A summary of the radiocarbon data from Newton Bay***

4.2 SOUTHAMPTON WATER

Southampton Water is a large estuary forming part of the Solent estuarine system (Figure 4.8). It is *c.* 10 km in length and 2 km wide at its confluence with the Solent. The estuary progressively narrows up stream until it bifurcates into the tidal reaches of the rivers Test and Itchen. These tributaries, along with the rivers Hamble and Meon, supply the estuary with freshwater draining from the chalk uplands to the north, and the mean surface salinity within the main body of Southampton Water is just below 34 ‰ (Webber, 1980).

The tidal characteristics of Southampton Water are also influenced by its proximity to the M_2 amphidrome (Section 4.1) resulting in a characteristic double high water. A summary of the tidal conditions is presented in Table 4.1.

4.2.1 Geological History

Southampton Water is located within the Hampshire Basin, and therefore possesses many similarities to Poole Harbour (Section 4.1.1). The main body of the estuary is underlain by the mid to late Eocene sands and clays of the Bracklesham Formation (West, 1980). During the Pleistocene, Southampton Water was a major tributary of the Solent River (Section 4.1.1), and extensive gravel deposits cover its floor and form terraces at its margins (Everard, 1954, Dyer, 1975; West, 1980; Nicholls, 1987, Allen & Gibbard, 1993). These are commonly overlain by terrestrial peat and estuarine silt and clay deposits which formed during the Holocene (Godwin & Godwin, 1940; West, 1980; Long & Scaife, 1996).

4.2.2 Sediment Sources

A variety of depositional regimes co-exist within the estuary reflecting the prevailing energy conditions. The estuary is ebb dominated and the enhanced tidal currents along the eastern shore result in coarse grained sediments and active erosion. Coarse gravel beaches such as those at Calshot Spit are found at the entrance to the system, whilst the western shore of the estuary and the reaches of the rivers are commonly flanked by cohesive muds and saltmarshes. The area is an important port and considerable sections of the shoreline have been subject to defence and reclamation (Coughlan, 1979).

The sea bed sediments of Southampton Water comprise silt clays with sand content increasing in the upper estuary (Algan *et al.*, 1994). This distribution is the inverse of that observed in Poole Harbour, and Dyer (1980) suggests that the majority of the fine-grained sediment is of marine origin.

4.2.3 Existing Sea-level Data

The dock excavations at the end of the 19th and beginning of the 20th century revealed the widespread distribution of buried peat beds in Southampton Water (Shore & Elwes, 1889; Anderson, 1933; Godwin & Godwin, 1940). Pollen anlaysis of one such exposure at George V Graving Dock, indicates that the base of the peat bed (residing at *c.* -6 m OD) formed between *c.* 9500 and 9000 Cal. BP (Godwin & Godwin, 1940). Rising sea-levels inundated this peat around 6000 Cal. BP which is replaced by marine silt and clays at -1.3 m OD. A corpus of borehole data is interpreted by Long & Scaife (1996) as indicative of widespread peat development in the upper estuary during the mid Holocene. There is also evidence for an expansion of peat forming communities in the lower estuary where freshwater peat is recorded overlying marine sediments at Fawley (Churchill, 1965; Hodson & West, 1972), and Hythe (Barton & Roche, 1984; Long & Scaife, 1996). It is suggested that this period of expansion commenced around 5500 Cal. BP and persisted until between 3500 Cal. BP and 3000 Cal. BP, after which minerogenic sedimentation prevailed (Long *et al.*, 1997).

An age-altitude graph of sea-level index points for Southampton Water has been proposed by Long *et al.* (1997), and is shown in Figure 4.9. This comprises sea-level index points derived from Hythe, Hamble, Dibden Bay and Bury Farm within Southampton Water, and Stansore Point just beyond the estuary mouth (Long & Tooley, 1995; Long & Scaife, 1996; Long *et al.*, 1997). These data are presented in Appendix Three.

4.2.4 Bury Farm

Bury Farm is located in the upper reaches of Southampton Water, situated opposite the Western Docks and the Bury Farm swinging ground (Figure 4.10). The marsh is around 100 m wide, although deeply dissected by saltmarsh creeks. The upper marine limit is well defined by a small break in slope associated with a strand line of flotsam and decaying vegetation (Plate 4.3). The saltmarsh resides between +2.00 m OD and +1.50 m OD, and is connected to the lower lying mudflat area by a mud ramp. The intertidal mudflat extends from +1.00 m OD to -1.70 m OD, and is underlain by a consolidated peat bed which outcrops below 0 m OD and is covered by at thin veneer of modern sediments. The roots systems of buried trees are clearly visible at low tide (Plate 4.4). The marsh is currently undergoing erosion at an estimated rate of between 0.50 and 1.00 m a^{-1} (ABP, 1995).

4.2.4.1 Lithostratigraphy

The lithostratigraphic investigation at Bury Farm consists of one transect comprising a total of 10 boreholes. Tabulated sediment descriptions and accompanying figures can be found in Appendix One. A summary transect is presented in Figure 4.11, showing all the major stratigraphic units recorded.

The pre-Holocene surface at Bury Farm exhibits a distinct stepped profile, the upper level being located at +1.8 m OD, the second level centred around +0.4 m OD, and the latter cores indicating an abrupt drop to -1.6 m OD, followed by a progressive deepening. Two well-developed peat beds are apparent below +0.4 m OD. The lower peat extends from around -2.0 m OD up to *c.* -1.0 m OD, and possesses a shell rich layer at its base. The upper woody peat extends from *c.* -0.5 m OD up to +0.4 m OD where it laps onto the middle 'step'. The sediments between and above these peats are predominantly grey silt and clay with occasional lenses of coarser material. A very thin, poorly developed and laterally discontinuous peat layer is occasionally recorded around +1.0 m OD.

4.2.4.2 Biostratigraphy

The fossil foraminiferal investigation at Bury Farm comprises 34 samples from two boreholes, and a total count of *c.* 7300 specimens. Full results, which are presented diagramatically and in tabular form, can be found in Appendix Two. A summary foraminiferal diagram from core BF-96-11 is shown in Figure 4.12, which possesses all the major biostratigraphic units identified. This is visually divided into five assemblage zones to aid description.

Zone I extends from 222 cm to 168 cm and is characterised by a mixed assemblage of *Jadammina macrescens*, *Miliammina fusca*, and *Trochammina inflata*, in association with the calcareous taxa *Ammonia beccarii*, *Haynesina germanica*, *Nonion depressulus*, and *Elphidium* species.

Zone II extends from 168 cm to 86 cm and is dominated by *J. macrescens*, with lesser numbers of *T. inflata*, and *Haplophragmoides species A*, and low counts of *M. fusca*.

Zone III extends from 86 cm to 80 cm and contains a low abundance, monospecific assemblage of *J. macrescens*.

Zone IV extends from 80 cm to 28 cm and is similar in composition to Zone 2, except that *M. fusca* is absent and *T. inflata* is present in much lower abundances.

Zone V extends from 28 cm to the present surface and possesses an assemblage dominated by *J. macrescens* with *T. inflata*.

4.2.4.3 Chronostratigraphy

The chronostratigraphy at Bury Farm consists of 6 AMS radiocarbon dates collected during the course of this study combined with a single conventional radiocarbon date presented in Long *et al.* (1997) (see Appendix Three). The results of these radiocarbon assays are summarised in Table 4.4, whilst full details are presented in Appendix Three. The ages of all available radiocarbon dates are indicated in Cal. BP on the summary stratigraphy of Figure 4.11.

Sample	Publication Code	Source Core	Altitude (m OD)	^{14}C Enrichment (% Modern ± 1σ)	Carbon Content (% by Weight)	$\delta^{13}C_{PDB}$ ± 0.1%	^{14}C yr BP ± 1σ	Cal. Year Range (2σ)
BF#1	BETA-105582	BF-96-11	1.40	-	-	-32.0	2550 ± 40	2460 - 2760
BF#2	BETA-105581	BF-96-11	1.11	-	-	-32.5	1530 ± 40	1320 - 1520
BF#3	BETA-105580	BF-96-11	1.00	-	-	-32.1	2900 ± 40	2300 - 2880
BF#4	BETA-105579	BF-96-11	0.71	-	-	-31.8	2980 ± 50	2970 - 3330
BF#5	BETA-105578	BF-96-11	0.25	-	-	-31.1	3960 ± 50	4260 - 4530
BF#6	CAMS-41306	BF-96-50	0.20	66.97 ± 0.40	9.3	-29.0	2870 ± 50	2860 - 3120

Table 4.4 ***A summary of the radiocarbon data from Bury Farm***

4.3 LOUGHOR ESTUARY

The Gower Peninsula is located to the west of Swansea in West Glamorgan, South Wales (Figure 4.13). The study area is situated on its northern coast which is characterised by large expanses of saltmarsh which have formed in the shelter of a protective sand spit. These marshes comprise around 35 % of the 45 km^2 Loughor Estuary (Burry Inlet) through which the rivers Loughor, Lliw, Llan, Gwili and Morlais discharge into Carmarthen Bay (Kay & Rojanaripart, 1977; Moore, 1977). The estuary is macro-tidal (see Table 4.1), and the tidal prism exceeds river discharge by several orders of magnitude, resulting in well mixed waters with salinities ranging from 29 ‰ to 33 ‰ (Moore, 1976).

4.3.1 Geological History

The Loughor Estuary is largely underlain by Coal Measures, although Millstone Grit and Carboniferous Limestone are found along the southern shores of the inlet. The presence of glacial striae and till deposits indicate that the Loughor Valley has been glaciated and, at its greatest extent, ice appears to have reached the end of the Gower Peninsula (Bridges, 1977). The sand dunes currently found at Whiteford Point have accumulated on a foundation of gravels suggested to be the remnants of the glaciers terminal moraine (Bridges, 1977).

Plummer (1960) investigated the lithostratigraphy of saltmarshes around Llanelli, on the northern shore of the Loughor Estuary. The sequences contained evidence for two humified peat beds, one below -2.5 m OD resting upon 'firm clay', and the other around 0 m OD intercalated between silt-clay sediments. The remaining sequence is dominated by silt-clays but with some organic intercalations. Carling (1978) recovered two cores from Llanrhidian Marsh to seaward of the area investigated in this study. The sequence reveals a single 'Flandrian peat' *c.* 2 m thick, centred around 0 m OD and intercalated between minerogenic sediments that is interpreted on the basis of pollen analysis as forming in an open fresh to brackish environment. This sequence is similar to the record from the northern shore reported by Plummer (1960) although Carling suggests that the 'higher peats' are absent at Llanrhidian.

4.3.2 Sediment Sources

Sediments within the Loughor Estuary mainly consist of reworked fluvioglacial outwash that was mobilised and moved onshore by rising Holocene sea level (Bridges, 1977). The sea bed deposits consist of a very well sorted, fine sand and the only coarser material comes from whole and fragmented shells (Elliott & Gardiner, 1981). A relatively large tidal range has permitted the development of extensive saltmarsh areas measuring up to 1.5 km in width (Goodwin, 1983). Large tidal creeks intersect the marsh, and the surface is further complicated by shallow infilled drainage features and salt pans (Plate 4.5). At the seaward margin, a pronounced cliff separates the saltmarsh from the extensive sand and mud flats beyond. The cliff appears to be currently eroding, and many blocks of saltmarsh sediments are to be found lying on the adjacent intertidal flats.

The main tidal channel is known to have migrated in the past, swinging southward at the beginning of the 18th Century, before slowly returning to its more northerly course between 1830 AD and 1877 AD (Goodwin, 1983). After 1877 AD, construction of a training wall enhanced deposition in the southerly margin (Plummer, 1960). The introduction of *Spartina anglica* in 1935 AD resulted in substantial marsh expansion (Plummer, 1960).

4.3.3 Existing Sea-level Data

There are no sea-level index points from the Loughor Estuary and the closest body of sea-level data comes from Bristol Channel. A large number of radiocarbon dates have been compiled by Kidson & Heyworth (1973, 1976, 1978) and Heyworth & Kidson (1982) who suggest that relative sea-level change in this area has been minimal during the last 3000 years. The reliability of this record is questionable however since much of the dated material does not possess a quantifiable vertical relationship to a tide level.

Figure 4.14 shows an age-altitude graph of validated sea-level index points held by the Sea-Level Research Unit at the University of Durham, and these data along with their sources are catalogued in Appendix Three. A recent investigation of the Somerset Levels indicates that the altitudes of many of these sea-level index points have been lowered by compaction (Haslett *et al.*, 1998). Whilst Haslett *et al.* (1998) present only one sea-level index point from the last 3000 years, they conclude that there is no support for the idea of 'stable' relative sea-level during this period proposed by Heyworth & Kidson (1982). Instead, they suggest relative sea-level was generally rising and add that their foraminiferal data provides support for the oscillatory nature of relative sea-level change inferred by Allen (1995).

4.3.4 Llanrhidian Marsh

Preliminary lithostratigraphical investigations were conducted at intervals around the coastline extending from Llanelli on the north shore, Gowerton at the eastern edge of the inlet, through Pen-clawdd, Crofty and Wernffrwd on the south shore to Landimore and Llanmadoc toward the west (Figure 4.13). The area around Llanrhidian revealed sedimentary sequences dominated by well-developed peat beds associated with buried trees, and was selected on this basis for further study (Figure 4.15). This area forms a slight embayment adjacent to the village of Llanrhidian and is backed by steep hills abruptly rising to over +70 m OD. Freshwater springs issue from the scarp slope and feed the Leason and Llanrhidian Pills that drain across the marsh and the intertidal Llanrhidian Sands. This significant input of freshwater led to the establishment of the wool mill at Stavel Hagar in whose grounds the two westernmost transects are located.

The vegetation is dominated by *Salicornia* species, *Puccinellia maritima*, *Sueda maritima*, *Plantago maritima*, *Halimione portulacoides*, *Aster tripolium*, *Festuca rubra*, *Armeria maritima*, *Juncus maritima*, and *Spartina anglica* (Goodwin, 1983). The marshes are grazed by sheep and ponies which tend to favour certain species and can significantly stunt the growth of others.

4.3.4.1 Lithostratigraphy

The site lithostratigraphy was investigated using 46 boreholes in five transects (Figure 4.15). Full stratigraphic descriptions and diagrams are included in Appendix One, and the main features are discussed below.

A summary lithostratigraphic diagram is presented in Figure 4.16. Most cores terminate in a stiff, grey clay-silt with sand and some larger clasts. This unit is sometimes iron-stained and may contain large roots. This unit deepens northward from around +4.0 m OD at the landward margins of the marsh to around +2.2 m OD toward the axis of the estuary.

Resting upon this basal unit is a thick, humified *turfa* with detrital wood, typically extending up to +3.6 m OD. This unit is intercalated with grey clay-silt layers of varying thickness, often iron stained and containing *turfa*, detrital wood and sometimes *Phragmites* remains. Above this main peat bed, organic-rich silt and clay extends across much of the marsh. The organic content of these sediments appears to increase to seaward. At around +4.2 m OD, a thin humified peat layer is commonly recorded, although this is most pronounced in the western transects.

The sequence is capped by around 50 cm of grey, sandy clay-silts, sometimes iron-stained, which grade into the modern surface sediments.

4.3.4.2 Biostratigraphy

The fossil foraminiferal investigation at Llanrhidian Marsh comprises 45 samples from four boreholes, and a total count of 1600 specimens. Full results, which are presented diagramatically and in tabular form, can be found in Appendix Two. A typical fossil foraminiferal sequence taken from core STH2-96-80 is shown in Figure 4.17.

The majority of fossil material is devoid of foraminifera. When foraminifera are recovered, abundances are frequently low, and assemblages consist almost entirely of *J. macrescens*. Rarely, small numbers of *Haplophragmoides* species, *Miliammina fusca*, and *Trochammina inflata* are also observed, along with foraminiferal test linings.

Thecamoebians are abundant in many samples not containing foraminifera, and sometimes occur in association with low abundance foraminiferal assemblages.

4.3.4.3 Chronostratigraphy

The chronostratigraphy at Llanrhidian Marsh consists of 5 AMS radiocarbon dates collected as part of this study. The results of these radiocarbon assays are summarise in Table 4.5, whilst full details are presented in Appendix Three. The ages of all available radiocarbon dates are presented in Cal. BP on the summary stratigraphy of Figure 4.16.

Sample	Publication Code	Source Core	Altitude (m OD)	^{14}C Enrichment (% Modern ± 1σ)	Carbon Content (% by Weight)	$\delta^{13}C_{PDB}$ ± 0.1%	^{14}C yr BP ± 1σ	Cal. Year Range (2σ)
LLN#1	CAMS-41307	LLN1-96-66	+4.21	91.16 ± 0.55	11.5	-29.4	740 ± 50	570 - 730
LLN#2	CAMS-40114	STH2-96-10	+3.75	62.24 ± 0.45	2.3	-28.5	3560 ± 60	3690 - 3990
LLN#3	CAMS-41310	STH2-96-80	+3.46	78.05 ± 0.48	45.0	-29.1	1990 ± 50	1820 - 2040
LLN#4	CAMS-41309	STH2-96-80	+2.82	72.90 ± 0.53	14.6	-29.0	2540 ± 60	2360 - 2760
LLN#5	CAMS-41308	STH2-96-80	+2.42	62.02 ± 0.39	11.4	-29.5	3840 ± 60	4000 - 4410

Table 4.5 ***A summary of the radiocarbon data from Llanrhidian Marsh***

Contemporary Saltmarsh Foraminifera

Distillation of precise sea-level information from the fossil foraminiferal assemblages outlined in Chapter Four depends upon a detailed understanding of the modern relationship between saltmarsh foraminifera and tide levels. These modern analogues should be collected from a wide range of saltmarshes since any study marsh may have varied significantly in character through time. In this chapter, the contemporary foraminiferal distributions at four study marshes are investigated and their relationship to elevation above mean tide level determined. These data are then combined with existing foraminiferal data from contrasting saltmarsh environments to produce an extensive modern training set. This training set is then used to develop a transfer function relating foraminiferal assemblages to elevation above mean tide level, which will be applied to fossil material in Chapter Six.

This chapter comprises the following:

- A description of the modern saltmarsh foraminiferal death assemblage distribution at each study area (Poole Harbour, Southampton Water, Loughor Estuary);
- Quantification of the relationship between these contemporary foraminiferal assemblages and elevation relative to mean tide level;
- Development of a transfer function from a large training set of modern UK saltmarsh foraminifera to permit reconstruction of past water levels from fossil foraminiferal assemblages (Chapter Six);
- An assessment of the transfer function's performance and applicability to the reconstruction of relative sea-level at each study site.

5.1 ZONATION METHOD

The use of a biological organism as a sea-level indicator rests upon the assumption that its distribution is related to sea level in a consistent manner and, that by establishing this vertical relationship with a specified tide level, the former position of relative sea-level, may be determined (Section 2.5). The seminal paper by Scott & Medioli (1980a) visually grouped contemporary saltmarsh foraminifera to define vertical zonations at a number of sites. More recent work has sought to group assemblages on statistical grounds which may then be related to distinct depositional environments (e.g. De Rijk, 1995a; Barbero *et al.*, 1997; Horton, 1997; Woo *et al.*, 1997). One statistical technique commonly employed is unconstrained cluster analysis which groups samples possessing similar characteristics regardless of their original order. These results can then be displayed in the form of a hierarchy or dendrogram (Kovach, 1995). Two variants of 'Unconstrained Incremental Sum of Squares Cluster Analysis' are used in this thesis: 'unweighted Euclidean distance' in which the data are not transformed or standardised; and 'unweighted Chord distance' in which the square root percentage foraminiferal abundances are used, thereby giving greater significance to the minor taxa. Identical zones discriminated by both variants are not dependent on the method used, and are therefore considered to be statistically reliable (Prentice, 1986; Horton, 1997).

5.2 ARNE PENINSULA

To investigate the contemporary foraminiferal distribution at Arne Peninsula, Poole Harbour (Figures 4.1 & 4.2), 18 surface samples were collected from a range of altitudes between +1.10 m OD and -0.16 m OD along a 130 m transect (Figure 5.1). Samples were collected in September 1995 and March 1997, during low water of spring tide (Table 5.1). The March 1997 survey extended above and below the lowest sample collected in September 1995 and, when combined, both surveys produced a consistent distribution. The total assemblage contains 26 species, whilst the life and death assemblages consist of 19 and 26 species respectively. The total number of specimens counted was 4351, with the death assemblage contributing 62% of these data. The dominant foraminiferal species in the death assemblage are *Jadammina macrescens*, *Haplophragmoides Sp. A*, and

Miliammina fusca, which contribute 25%, 21%, and 20% of the dead foraminiferal population respectively. The most abundant calcareous taxa are *Haynesina germanica* and *Ammonia beccarii* which constitute 8% and 7% of the dead foraminiferal population respectively. Full counts are presented in Appendix Two.

Site	Mean High Water Spring Tide (m OD)	Mean High Water Neap Tide (m OD)	Mean Low Water Neap Tide (m OD)	Mean Low Water Spring Tide (m OD)	Mean Sea Level (m OD)	Mean Tide Level (m OD)	Tidal Range Springs (m)	Tidal Range Neaps (m)
Arne Peninsula	0.60	0.30	-0.10	-0.60	0.10	0.05	1.20	0.40
Newton Bay	0.70	0.10	-0.20	-0.90	0.10	-0.08	1.60	0.30
Bury Farm	1.76	0.96	-0.94	-2.24	0.16	-0.12	4.00	1.90
Llanrhidian Marsh	4.14	2.14	-1.16	-2.96	1.39	0.54	7.10	3.30

Table 5.1 *Tidal characteristics of the four study marshes* [**Source:** Admiralty Tide Tables, 1998]

5.2.1 Vertical Distribution of Foraminifera

Figure 5.2 shows a plot of the contemporary death assemblages against altitude. The four stations above +0.80 m OD are characterised by low counts, typically less than 20 specimens. A pronounced change in the death assemblage occurs between +0.70 m OD and +0.60 m OD where the tests of calcareous species are first observed. These taxa are restricted to the marsh and mudflat below this level, as is the agglutinated species *Ammoscalaria runiana*.

The cluster analysis results presented in this thesis are generated within the TILIA program (version 2.0 b.0.5, Grimm, 1991 -1993) after removal of species contributing less than 2% to the death assemblage and counts of smaller than 40 specimens (Fritz *et al.* 1990; Gehrels, 1994). This screening process removes 16 of the minor taxa and excludes data from the four stations above +0.80 m OD. Cluster analysis results for the contemporary death assemblage at Arne Peninsula are displayed graphically in Figures 5.3 and 5.4. Both unweighted Euclidean distance and unweighted Chord distance variants distinguish 4 groups. Comparison of the two dendrograms reveals that there are 2 groups common to both analyses:

Group AR-I extends from +0.80 m OD to +0.05 m OD and is dominated by *J. macrescens, M. fusca*, and *Haplophragmoides species A*. The dominant calcareous taxa are *A. beccarii*, *Quinqueloculina* species, and *Elphidium williamsoni*.

Group AR-II extends from +0.25 m OD to -0.16 m OD and is dominated by the calcareous species *H. germanica* and *A. beccarii*, with *E. williamsoni* and *Nonion depressulus* locally contributing up to *c.* 20% of the death assemblage. Agglutinated taxa generally contribute less than 20% of the total death assemblage.

The altitudinal distribution of these groups is displayed graphically as a pair of box plots in Figure 5.5, indicating that foraminiferal assemblages are vertically zoned. Figure 5.6 shows changing percentage abundance with altitude of the 6 principal taxa at Arne Peninsula. These taxa demonstrate that saltmarsh foraminifera exhibit a range of responses to changes in altitude. *J. macrescens* possesses a positive relationship to altitude, whilst the distribution of *M. fusca* is unimodal with a pronounced peak between +0.70 m OD and +0.80 m OD (corresponding to Zone ARC-I). The calcareous taxa generally possess a negative relationship to altitude, whilst *Haplophragmoides Sp. A* and *Quinqueloculina* species show only a weak correlation.

5.3 NEWTON BAY

To investigate the contemporary foraminiferal distribution at Newton Bay, Poole Harbour (Figures 4.1 & 4.5), 9 surface samples were collected from a range of altitudes between +0.75 m OD and +0.14 m OD along a 65 m transect (Figure 5.7). Samples were collected in June 1996, on a rising neap tide (Table 5.1). The total assemblage contains 13 species, whilst the life and death assemblages consist of 10 and 13 species respectively. The total number of specimens counted was 2216, with the death assemblage contributing 84% of these data. The dominant foraminiferal species in the death assemblage are *Jadammina macrescens*, *Trochammina inflata*, and *Miliammina fusca* which contribute 41%, 27%, and 17% of the death assemblage respectively. The most abundant calcareous taxa are

Quinqueloculina species and *Elphidium williamsoni* which each constitute 1% of the dead foraminiferal population. Full counts are presented in Appendix Two.

5.3.1 Vertical Distribution of Foraminifera

Figure 5.8 shows a plot of the contemporary death assemblages against altitude. The stations below +0.56 m OD are characterised by low counts. This decrease in abundance corresponds closely to a decline in *Haplophragmoides Sp. B* and *Tiphotrocha comprimata*, and an increase in the proportion of calcareous taxa recorded. Calcareous species are present throughout the marsh although these frequently contribute less than 10% of the death assemblage.

Data screening removes 3 of the minor taxa and excludes data from the 3 stations below +0.51 m OD. Cluster analysis results of the contemporary death assemblage at Newton Bay are presented graphically in Figures 5.9 and 5.10. Unweighted Euclidean distance and unweighted chord distance both distinguish two groups:

Group NB-I extends from +0.75 m OD to +0.61 m OD and is characterised by a dominance of *J. macrescens*, with *M. fusca*, *T. inflata*, and low abundances of *Haplophragmoides Sp. B* and *T. comprimata*. Small numbers of calcareous species are also recorded although these never contribute more than 2% of the death assemblage.

Group NB-II extends from +0.56 m OD to +0.51 m OD and is distinguished from Group I by the absence of *Haplophragmoides Sp. B* and *T. comprimata*.

The vertical distribution of these groups is displayed graphically as a pair of box plots in Figure 5.11, and the relationship of the major taxa to altitude is shown in Figure 5.12. The boxplots demonstrate that the modern foraminiferal assemblages are vertically zoned. The behaviour of the principal taxa is similar to that described for Arne in Section 5.2.1, although the assemblages are dominated by *J. macrescens* and *M. fusca* to a much greater degree.

5.4 BURY FARM

To investigate the contemporary foraminiferal distribution at Bury Farm, Southampton Water (Figures 4.8 & 4.10), 18 surface samples were collected from a range of altitudes between +2.29 m OD and -1.76 m OD along a 190 m transect (Figure 5.13). The samples were collected in March 1997 at low water of spring tide (Table 5.1). The total assemblage contains 21 species, whilst the life and death assemblages consist of 14 and 21 species respectively. The total number of specimens counted was 8316, with the death assemblage contributing 80% of the data. The dominant foraminiferal species in the death assemblage are *Miliammina fusca* (33%), *Jadammina macrescens* (32%), and *Trochammina inflata* (12%). The most abundant calcareous taxa are *Ammonia beccarii* and *Haynesina germanica* which contribute 4% and 3% of the death assemblage respectively. Full counts are presented in Appendix Two.

5.4.1 Vertical Distribution of Foraminifera

Figure 5.14 shows a plot of the contemporary death assemblages against altitude. The stations above +2.08 m OD, at -0.57 m OD and -0.97 m OD, are all characterised by low counts. There is a pronounced increase in the abundance of calcareous foraminifera below +0.83 m OD which coincides with decreases in the abundance of *Haplophragmoides* species, and *M. fusca*.

Data screening removes 7 of the minor taxa and excludes data from 2 stations. Cluster analysis results of the contemporary death assemblage at Bury Farm are presented graphically in Figures 5.15 and 5.16. The unweighted Euclidean distance variant discriminates 5 groups whilst the unweighted Chord distance techniques identifies 4 clusters. Comparison of the two dendrograms reveals that there are 3 groups common to both analyses:

Group BF-I extends from +2.08 m OD to +1.64 m OD and is dominated by *J. macrescens* with lesser abundances of *Haplophragmoides Sp. A*, *T. inflata*, *Haplophragmoides Sp. B*, and *M. fusca*.

Group BF-II extends from +1.88 m OD to +0.83 m OD and is co-dominated by *M. fusca.* and *J. macrescens*, with lesser numbers of *T. inflata*, and *Haplophragmoides* species.

Group BF-III extends from +0.43 m OD to -1.76m OD and is typified by high abundances of *J. macrescens*, and the calcareous species *A. beccarii* and *H. germanica.* Lower abundances of *T. inflata* are also recorded along with low percentage frequencies of calcareous taxa.

The vertical distribution of these groups is displayed graphically as three box plots in Figure 5.17, and indicates that the modern foraminiferal assemblages are vertically zoned. Figure 5.18 shows that *J. macrescens* exhibits a positive relationship with altitude, whilst the distributions of *M. fusca* and *Haplophragmoides Sp. A* are unimodal. *A. beccarii* shows a negative distribution whilst the remaining taxa exhibit a weak relationship with altitude.

5.5 LLANRHIDIAN MARSH

To investigate the contemporary foraminiferal distribution at Llanrhidian Marsh, Loughor Estuary (Figures 4.13 & 4.15), 11 surface samples were collected from altitudes ranging between +5.56 m OD and +3.66 m OD along a 1100 m transect (Figure 5.19). The samples were collected in February 1997 after one of the highest tides of the year (Table 5.1). The total assemblage contains 33 species, whilst the life and death assemblages contain 12 and 33 species respectively. The total number of specimens counted was 5003, with the death assemblage contributing 65% of the data. The dominant foraminiferal species in the death assemblage is *Jadammina macrescens* which accounts for 57% of the dead specimens counted. The second most numerous taxon is the calcareous species *Haynesina germanica* although this only contributes 11% of the death assemblage. Full counts are presented in Appendix Two.

5.5.1 Vertical Distribution of Foraminifera

Figure 5.20 shows a plot of the contemporary death assemblages against altitude. The samples above +4.89 m OD are devoid of calcareous foraminifera, but there are two other samples at lower elevations which also possess no calcareous tests. *Haplophragmoides* species are largely restricted to elevations greater than +4.16 m OD. It is also interesting to note the low percentage frequencies of *Miliammina fusca* which never contribute more than 5% of the death assemblage.

Data screening removes 25 of the minor taxa but all stations have counts greater than 40. Cluster analysis results of the contemporary death assemblage at Llanrhidian Marsh are presented graphically in Figures 5.21 and 5.22. The unweighted Euclidean distance variant discriminates 2 groups whilst the unweighted Chord distance techniques identifies 3 clusters. Comparison of the two dendrograms reveals that none of the groups are common to both techniques and so are of limited statistical significance (Prentice, 1986).

5.6 DISCUSSION

The purpose of this survey of modern foraminiferal distributions is to improve the accuracy of indicative meanings derived from fossil foraminiferal assemblages, and thereby enable the development of a high resolution sea-level record. To this end, it is first necessary to assess the extent to which the contemporary distributions are related to altitude (see Section 3.4.1.2).

The cluster analysis results coupled with the vertical distribution of foraminiferal species provide strong evidence that the death assemblage found at Arne Peninsula is related to altitude (Figures 5.2, 5.5, and 5.6). The upper marsh is typified by an agglutinated assemblage dominated by *Jadammina macrescens* and *Miliammina fusca* which gives way to the low marsh assemblage dominated by calcareous species such as *Ammonia beccarii* and *Haynesina germanica*. The agglutinated species *Ammoscalaria runiana* possesses a similar distribution to the calcareous fraction, and is restricted to below +0.60 m OD.

A similar vertical differentiation between agglutinated and calcareous dominated assemblages is also apparent at Bury Farm, and is picked out by the cluster analysis results (Figures 5.14, 5.17, and 5.18). *A. runiana* is also present at Bury Farm, although in very low numbers, and occupies the same altitudinal range as the calcareous foraminifera.

There is no pronounced change between agglutinated and calcareous dominated assemblages at Newton Bay, despite the marsh being located close to Arne Peninsula.

Cluster analysis discriminates two zones on the basis of changes in the percentage frequencies of *Haplophragmoides* species, but only after a third of the data is screened out due to low counts. The marsh at Newton Bay is backed by a low bluff and as a consequence there is no high marsh to upland transition (see Section 4.1.4). In addition, the marsh is currently experiencing erosion, enhanced by the die-back of *Spartina*, which is releasing considerable volumes of previously trapped sediment back into the marine system (Section 4.1). Distinct depressions partially vegetated by sparse stands of *Spartina* are associated with very wet, unconsolidated, fine grained sediment, which may be recently deposited or mobile. The lack of any clear sequence of foraminiferal change is most likely a product of the small vertical range of samples analysed (26 cm), but given the observations above, may also be complicated by the erosion and transport of sediment within the marsh. In particular, the high percentage frequencies of *M. fusca* in the low abundance, low marsh samples, is in marked contrast to its distribution at Arne Peninsula and Bury Farm. This may be indicative of reworking, with incorporation of tests transported from higher marsh samples.

The situation at Llanrhidian Marsh is more complex than that of the Solent marshes, and a simple vertical bipartite division of calcareous and agglutinated foraminifera is not apparent. Instead, the distribution of foraminifera appears to be more closely related to distance from the seaward marsh edge (Figure 5.23). The marsh profile in Figure 5.19 reveals that the majority of the marsh surface analysed is above mean high water spring tides, and consequently is rarely inundated. The occurrence of such high numbers of calcareous taxa is therefore surprising. One explanation for this is that many of the tests are allochthonous, transported onto the marsh by the unusually high tide which preceded sample collection. Evidence supporting this conjecture is provided by an analysis of the life assemblage distribution (Figure 5.24) which shows that living calcareous foraminifera are rarely found beyond the marsh fringe and never in significant numbers. The allochthonous nature of the calcareous assemblage would account for the relative homogeneity of its composition and explain the lack of any defined altitudinal relationship.

The topographic high apparent in Figure 5.19 marks the transition between assemblages containing calcareous taxa, and those composed entirely of agglutinated species. Station 5 (+4.54 m OD), located on the seaward face of the mid marsh ridge, possesses anomalous foraminiferal death and life assemblages, both of which exhibit much higher proportions of calcareous foraminifera than adjacent stations, and include *Cibicides lobatulus* which is commonly associated with more fully marine environments. This marine signature is also apparent in the agglutinated death assemblage which possesses unusually high numbers of *Trochammina squamata*. It is possible that the ridge impedes the advection of marine waters across the marsh during the highest of tides, resulting in enhanced deposition of lower marsh species at this higher altitude.

De Rijk (1995a) and De Rijk and Troelstra (1997) have suggested that salinity is the dominant control on foraminiferal distributions and consequently a vertical zonation is only apparent when salinity exhibits a consistent relationship with altitude (Section 3.4.1.2). The marsh studied by De Rijk (1995a) where salinity was found to be the dominant control on surface foraminiferal distributions, was large and possessed an uneven surface topography similar to that observed at Llanrhidian. It is therefore possible that the foraminiferal distribution at Llanrhidian Marsh is being controlled by salinity, but may also be complicated by the presence of allochthonous material transported onto the marsh.

5.6.1 Summary

On the basis of the data presented above, it is proposed that the contemporary death assemblage distributions at Arne Peninsula and Bury Farm possess clear indications of a close relationship with altitude. A similar relationship is apparent in the distribution at Newton Bay, although the narrow vertical range of sample material means that evidence for this is more equivocal. These sampling limitations mean that high marsh assemblages at Newton Bay must be inferred from those of neighbouring Arne Peninsula.

The contemporary foraminiferal death assemblage distribution at Llanrhidian marsh does not exhibit a consistent relationship to a tide level, and is therefore being influenced by factors other than elevation. Consequently, the Llanrhidian samples are not included in the contemporary dataset used to produce estimates of indicative meaning, and sea-level reconstructions at this site are treated separately in Section 5.8.3.

5.7 THE DEVELOPMENT OF A FORAMINIFERAL-BASED TRANSFER FUNCTION

Whilst the cluster analysis results presented above are useful in determining the presence of a vertical zonation in modern foraminiferal assemblages, they are of limited value in reconstructing a defined tide level from fossil material owing to the large altitudinal ranges covered by individual zones. For example, zone AR-I (Figure 5.5) spans a vertical interval of almost 80 cm which equates to nearly three-quarters of the spring tidal range at Arne Peninsula. An alternative statistical approach is the use of biologically-based transfer functions which permit reconstruction of one or more environmental variables from preserved floral or faunal proxies (Birks, 1995). Predictions produced by transfer functions have much smaller vertical errors than those associated with the cluster analysis zones presented above and therefore offer the potential to reconstruct higher resolution records of relative sea-level change.

Horton (1997) demonstrated that the modern distributions of UK saltmarsh foraminifera from 8 sites around the British Isles are related to sample elevation above mean tide level. Using these data as a modern training set, he developed a foraminiferal-based transfer function capable of reconstructing the former elevation of saltmarsh sediments relative to mean tide level. This provides the potential for continuous calculation of water level changes throughout fossil cores, enabling high resolution records of relative sea-level change to be developed. In this section, modern foraminiferal distributions from the vertically zoned marshes in Poole Harbour and Southampton Water, are combined with the Horton (1997) data to produce a new training set. This is then used to develop a foraminiferal-based transfer function.

5.7.1 The Modern Training Set

The new foraminiferal training set consists of 165 samples collected from 11 marshes (see Table 5.2) located around the coast of the UK (Figure 5.25). The new data provided by this study contribute 22% of the total dataset, and increase the spatial extent of modern samples which were previously absent from the UK's southern coast. All counts were converted to a common taxonomic basis, and screened to remove species contributing less than 2% to the death assemblage and samples with counts of less than 40 specimens (Fritz *et al.*, 1990).

Study Marsh	Number of Samples	Author
Arne Peninsula, Dorset	12	R.J. Edwards
Brancaster Marsh, Norfolk	23	B.P. Horton
Bury Farm, Hampshire	15	R.J. Edwards
Cowpen Marsh, Teeside	28	B.P. Horton
Kentra Bay, Argyll	6	E.J. Twiddy
Newton Bay, Dorset	6	R.J. Edwards
Nith Estuary, Solway Firth	15	J.M. Lloyd
Roudsea Marsh, Morecombe Bay	13	J.M. Lloyd
Thornham Marsh, Norfolk	24	B.P. Horton
Tramaig Bay, Isle of Jura	3	J.M. Lloyd
Welwick Marsh, Humberside	20	B.P. Horton

Table 5.2 *The composition of the contemporary foraminiferal training set*

5.7.2 Standardised Water Levels

The altitude of modern foraminiferal samples is measured relative to UK Ordnance Datum. These altitudinal data must be standardised to enable comparison between sites with differing tidal ranges and marsh elevations. Since the indicative meaning is defined with respect to tidal levels the altitudinal data are converted to an index related to the tidal cycle. Horton (1997) defines a 'standardised water level index' (SWLI) in the following way:

Standardised Water Level Index Equation
$\chi_{ab} = [((A_{ab} - MTL_b) \div (MHWST_b - MTL_b) \times 100) + 100]$
Where:
χ_{ab} = the SWLI of sample *a* at site *b*;
A_{ab} = the altitude (m OD) of sample *a* at site *b*;
MTL_b = the mean tide level (m OD) at site *b*;
$MHWST_b$ = the mean high water spring tide level (m OD) at site *b*.
After Horton (1997)

For example, surface samples collected from around mean high water spring tide at Arne Peninsula and Bury Farm, whilst possessing differing altitudes (+0.60 m OD and +1.76 m OD respectively), have the same SWLI (200). This equation is applied to the combined contemporary foraminiferal dataset, producing a SWLI for every surface sample.

5.7.3 Model Construction

The vertical zonation concept is based upon the idea that individual foraminiferal taxon are found within a particular altitudinal range, and that their abundance rises to a maximum value close to an altitudinal optimum. This form of response, common in many environmental systems, is termed 'unimodal' or 'Gaussian' and is effectively modelled by weighted averaging regression and calibration (WA) (ter Braak, 1987b; Birks *et al.*, 1990; Birks, 1995). The WA technique assumes that the most abundant foraminiferal taxon at a given SWLI will be close to its SWLI optimum. By averaging the SWLIs at which individual foraminiferal taxon are found, weighted by the taxon's relative abundance, the weighted average of each taxon's SWLI optimum can be calculated. These estimated optima can then be used during WA calibration to infer a value of SWLI from a foraminiferal assemblage. Certain taxa may be more sensitive to changes in elevation, and accordingly occur across a smaller range of SWLIs. Such narrow tolerance species can provide a more precise indication of SWLI, and may be given greater significance in WA by using the tolerance-weighted estimate ($WA_{(Tol)}$) of SWLI.

The WA technique produces a transfer function for SWLI by relating the training set of foraminiferal samples to their standardised altitudes *via* the CALIBRATE program (version 0.70, Juggins and ter Braak, 1997). The transfer function is first produced by consideration of the total dataset, and is then cross-validated to assess the function's predictive ability by repeating the process with one sample omitted (ter Braak & Juggins, 1993). This procedure, termed 'jack-knifing', is repeated using each sample in turn, and subtraction of the predicted value from the observed value produces a prediction error for each sample.

It is inevitable that there will be some samples which possess a poor relationship to SWLI. This will arise where the foraminiferal assemblage is strongly influenced by another environmental variable, such as salinity; when the assemblage is not completely autochthonous; or if the assemblage has undergone any post-depositional modification, such as the dissolution of calcareous tests (Horton, 1997). These samples decrease the predictive ability of the transfer function and must be screened out (Gasse *et al.*, 1995; Jones & Juggins, 1995). This screening process is implemented after the initial transfer function is produced, by deleting samples with absolute residuals (observed minus predicted) greater than the standard deviation of the SWLI (Jones & Juggins, 1995).

5.7.3.1 A Foraminiferal-Based Transfer Function for SWLI

Screening of the new training set described in Section 5.7.1 removes a total of 44 samples, 30 % of which are from the south coast marshes. These screened samples are largely from low marsh environments and dominated by calcareous taxa. Such low elevation environments are more prone to regional variability, reflecting the local characteristics of the estuary within which they are found (Scott & Medioli, 1980a). It is possible therefore, that the removal of these samples indicates that different conditions are found in southern coast marshes relative to their counterparts located elsewhere in the UK. A more likely explanation however, is that the difference is a consequence of variation in sampling design, since the new surveys from Arne Peninsula and Bury Farm extend much lower in the tidal frame than most of the pre-existing data.

The performance of this new transfer function is summarised in Table 5.3, and presented graphically in Figure 5.26. The root mean square error of prediction ($RMSEP_{(jack)}$) assesses the overall predictive abilities of the training set (Birks, 1995), whilst the coefficient of determination ($r^2_{(jack)}$) gauges the strength of the relationship between the predicted and observed values (Gasse *et al.*, 1995).

Method	Deshrinking Method	Apparent RMSE	r^2	RMSEP$_{(jack)}$	$r^2_{(jack)}$
WA	Inverse	10.45	0.61	11.60	0.52
WA	Classical	13.40	0.61	14.38	0.53
WA$_{(Tol)}$	Inverse	10.51	0.60	12.59	0.44
WA$_{(Tol)}$	Classical	13.52	0.60	15.86	0.45

Table 5.3 *Summary Statistics of WA and WA-Tol transfer functions for SWLI using the new, screened training set*

Figure 5.26 shows the pattern of observed *versus* predicted SWLIs and the absolute residuals for two deshrinking methods. Deshrinking is required to correct for the tendency of inferred SWLIs to shift towards the overall mean value of SWLI. This arises because averages are taken during WA regression to determine species optima and tolerance, and again during WA calibration to infer a value of SWLI. Classical regression deshrinks more than inverse regression (Martinelle, 1970) and pulls inferred values further away from the mean of the training set (Birks, 1995). It therefore performs better at extremes of the environmental gradient, and this is reflected by the small residuals below SWLIs of 170 and above SWLIs of 210. Conversely, inverse regression is more appropriate when considering mid-range values of SWLI (Gasse *et al.*, 1995), where the residual scatter is significantly less than for classical regression.

Regardless of the deshrinking method used however, the WA technique by its very nature, has limited predictive resolution at the upper and lower extremes of the environmental gradient, and both plots of observed against predicted SWLI possess a characteristic sigmoid shape (Figure 5.26). The propensity for high marsh predictions to plot toward the lower limits of the error range defined by RMSEP$_{(jack)}$ has particular significance for foraminiferal-based sea-level reconstructions, since the most sensitive assemblages are located in this high marsh environment. The plateau in predicted values of SWLIs above

220, mean that, whilst the assemblages in this region may be most sensitive to change, the transfer function's ability to detect this variation is at its lowest.

5.7.3.2 Model Assessment

Before the transfer function predictions can be employed to reconstruct SWLIs from fossil material, their applicability to these preserved sequences must be quantified. This is achieved *via* the modern analogue technique (MAT) in which the similarity (or dissimilarity) between fossil assemblages and the modern training set is quantified (Birks *et al.*, 1990). Whilst MAT can be used to predict values of SWLI, in this instance it is simply employed to assess the similarity between fossil samples and the modern training set, and in this way, provide an independent assessment of the reliability of WA predictions.

MAT is performed using the MODERN ANALOGUE TECHNIQUE program (release 1.1, 1997; Juggins & ter Braak, 1992), with SWLI reconstruction based on the weighted average of the 10 most similar modern samples. Squared chord distance was selected as the dissimilarity coefficient (Prentice, 1980; Overpeck *et al.*, 1985; Birks *et al.*, 1990) since it maximises the signal to noise ratio when used with percentage data (Birks, 1995). The performance of MAT for the training set is summarised in Table 5.4.

RMSEP $_{(jack)}$	r^2	**1st Percentile**	**2nd Percentile**	**5th Percentile**	**10th Percentile**	**20th Percentile**
8.90	0.73	0.05	0.07	0.12	0.19	0.31

Table 5.4 *Summary statistics for MAT predictions of SWLI and dissimilarity percentiles*

Dissimilarities below the 20th percentile threshold are indicative of good analogues (Horton, 1997). Consequently, samples with dissimilarity coefficients (DC) less than or equal to 0.31 have good analogues in the training set, whilst those with DCs greater than 0.31 have no close analogue, indicating that WA predictions may be in error.

On the basis of these criteria, MAT is used to investigate the applicability of the modern training set to core ARN1-95-90, recovered from the marsh at Arne Peninsula (see Figure 4.4). The results are shown in Table 5.5.

Sample Depth (cm)	Visual Classification	WA (Inv.) $SWLI_{(pred)}$	WA (Class.) $SWLI_{(pred)}$	Min. DC	Analogue
10	Mid Marsh	204.86	215.50	0.05	Good
20	Mid Marsh	196.70	202.07	0.07	Good
30	Mid Marsh	199.17	206.13	0.05	Good
40	Mid to Low Marsh	198.48	205.00	0.18	Good
50	Mid to Low Marsh	199.32	206.39	0.15	Good
60	**Mudflat**	**195.44**	**200.00**	**0.34**	**No Close**
70	**Mudflat**	**184.57**	**182.13**	**0.87**	**No Close**
80	**Mudflat**	**181.09**	**176.39**	**0.97**	**No Close**
85	Mid to Low Marsh	193.93	197.51	0.24	Good
90	Mid to Low Marsh	198.93	205.74	0.24	Good
95	**Mid to Low Marsh**	**199.82**	**207.21**	**0.43**	**No Close**
100	**Mudflat**	**198.74**	**205.43**	**0.42**	**No Close**
110	**Mudflat**	**194.98**	**199.23**	**0.31**	**No Close**
120	**Mid to Low Marsh**	**162.86**	**146.42**	**0.75**	**No Close**
130	**Mid to Low Marsh**	**189.65**	**190.47**	**0.32**	**No Close**
140	**Mid to Low Marsh**	**186.63**	**185.51**	**0.39**	**No Close**
150	Mid to Low Marsh	197.69	203.70	0.21	Good
160	**Mid to Low Marsh**	**183.63**	**180.58**	**0.43**	**No Close**
165	Mid to Low Marsh	190.16	191.31	0.14	Good
170	High Marsh	197.97	204.16	0.09	Good
175	High Marsh	199.56	206.77	0.09	Good
180	High Marsh	199.70	207.00	0.09	Good
181	High Marsh	200.90	208.98	0.07	Good
182	High Marsh	201.09	209.29	0.09	Good
183	High Marsh	200.23	207.88	0.03	Good
184	High Marsh	189.68	190.51	0.22	Good

Table 5.5 *A MAT assessment of WA predictions for SWLI for samples from core ARN1-95-90*

The MAT results indicate that 10 out of the 26 fossil samples possess no modern analogue. It is evident that the applicability of the modern training set to fossil samples decreases with decreasing marsh elevation. The most plausible explanation for this is that dissolution of fossil calcareous tests has occurred in the lower marsh environments, producing an agglutinated assemblage that does not reflect the conditions at the time of deposition. This post depositional modification violates one of the fundamental tenets of palaeoenvironmental reconstruction, which assumes fossil assemblages are directly comparable to their modern counterparts (Birks, 1995).

This transfer function is not without merit, in that failure to identify a suitable modern analogue serves to identify samples which may have experienced post-depositional modification, and can potentially highlight assemblages that have been decalcified. It also performs well in higher marsh, agglutinate-dominated settings, and has the potential to do so in low intertidal to subtidal sediments where calcareous taxa are preserved (Horton, 1997). This latter area is of particular interest since abundant calcareous assemblages can be used to furnish material for AMS radiocarbon dating, thereby permitting construction of sea-level index points from minerogenic sediments.

In the future, this may open up possibilities for constructing high resolution sea-level records from environments currently considered unsuitable due to a lack of organic material. At present however, there is a distinct lack of modern samples from these inaccessible environments, and therefore no modern analogues are available.

5.7.4 An Agglutinated Foraminiferal-Based Transfer Function

Development of a transfer function based on agglutinated foraminifera would circumvent the problems associated with the dissolution of fossil material. The success of such an approach depends on the ability to discern vertical changes in the agglutinated fraction of the modern assemblages. Whilst there are some minor taxa at Arne Peninsula and Bury Farm, such as *Ammoscalaria runiana*, which possess similar distributions to calcareous species (Figures 5.2 and 5.14), these are absent from most of the other marshes in the modern training set. In these cases, the low marsh agglutinated assemblages are similar in composition to their high marsh counterparts (Horton, 1997). Consequently, a training set of agglutinated taxa from all 11 marshes will be incapable of distinguishing between high and low marsh fossil assemblages.

The test linings of calcareous foraminifera are frequently preserved within the fine-grained sediments of the study marshes and, whilst not permitting differentiation of species, form a calcareous 'finger-print'. These test linings offer a means to discriminate between high and low marsh dissolution assemblages, and can be used in association with an agglutinate-based transfer function to improve the range of good modern analogues. To this end, the training set can be 'dissolved' by combining all calcareous taxa into a single group, which can then be used to construct a new transfer function. The data are screened as before, and this time a total of 42 samples are removed, 31% are from the south coast marshes. Once again, the majority of these samples are from the lowest elevations where the assemblages are less sensitive to changes in altitude. The results of this combined agglutinate-test lining foraminiferal (ALF) based transfer function are summarised in Table 5.6, and displayed graphically in Figure 5.27. Full details of this ALF-based transfer function are included in Appendix Four.

From the $r^2_{(jack)}$ values it is apparent that the ALF-based $WA_{(Tol)}$ transfer function performs better than the previous transfer function developed from the combined modern training set data described in Section 5.7.3.1. Scatter in the data generally increases with decreasing elevation, and reflects the fact that assemblages dominated by a single calcareous component are less sensitive to changes in elevation than the more diverse, agglutinated assemblages higher in the tidal frame.

Method	Deshrinking Method	Apparent RMSE	r^2	$RMSEP_{(jack)}$	$r^2_{(jack)}$
WA	Inverse	13.15	0.59	13.89	0.54
WA	Classical	17.16	0.59	17.85	0.55
$WA_{(Tol)}$	Inverse	12.53	0.62	13.46	0.57
$WA_{(Tol)}$	Classical	15.86	0.62	16.88	0.58

Table 5.6 *Summary statistics of the ALF-based WA and WA-Tol transfer functions for SWLI.*

Inverse deshrinking performs best between SWLIs of 160 to 205, whilst classical deshrinking is more appropriate at SWLIs less than 160, and greater than 205. The tendency for WA predicted SWLIs ($SWLI_{(pred)}$) at high elevations to plot below their true values remains. Figure 5.28 shows the residuals from both deshrinking methods plotted against $SWLI_{(Pred)}$ (classical deshrinking) in order to determine where the transition from inverse-based to classical-based reconstructions should be made. These results are summarised in Table 5.7.

$SWLI_{(Pred)}$ (Classical Deshrinking)	Most Appropriate Deshrinking Method
<146	Classical
146 - 209	Inverse
> 209	Classical

Table 5.7 *Summary of most suitable deshrinking method for specified ranges of* $SWLI_{(Pred)}$

5.7.4.1 Assessing the Performance of the ALF-based Transfer Function

The performance of the ALF-based transfer function is assessed, as before, using the MAT program. Any calcareous taxa remaining in the fossil samples were combined with test linings in order to be comparable with the dissolution training set. Table 5.8 shows the MAT results, and from this, it is evident that good analogues are present for DCs less than or equal to 0.19.

$RMSEP_{(jack)}$	r^2	1st Percentile	2nd Percentile	5th Percentile	10th Percentile	20th Percentile
11.18	0.70	0.01	0.02	0.05	0.10	0.19

Table 5.8 *Summary statistics for MAT predictions of SWLI and dissimilarity percentiles for the dissolution-based training set*

The applicability of the ALF-based transfer function is tested using the fossil core ARN1-95-90 as before, and the results summarised in Table 5.9.

Sample Depth (cm)	Visual Classification	$WA_{(Tol)}$ (Inv.) $SWLI_{(pred)}$	$WA_{(Tol)}$ (Class.) $SWLI_{(Pred)}$	Min. DC	Analogue
10	Mid Marsh	204.84	215.18	0.05	Good
20	Mid Marsh	192.68	195.69	0.03	Good
30	Mid Marsh	202.47	211.38	0.05	Good
40	Mid to Low Marsh	192.85	195.98	0.14	Good
50	Mid to Low Marsh	179.42	174.47	0.11	Good
60	Mudflat	170.11	159.56	0.04	Good
70	Mudflat	177.34	171.13	0.21	No Close
80	Mudflat	177.23	170.97	0.34	No Close
85	Mid to Low Marsh	159.83	143.10	0.04	Good
90	Mid to Low Marsh	163.58	149.10	0.04	Good
95	Mid to Low Marsh	182.77	179.83	0.12	Good
100	Mudflat	193.22	196.56	0.18	Good
110	Mudflat	189.77	191.04	0.11	Good
120	Mid to Low Marsh	163.25	148.57	0.24	No Close
130	Mid to Low Marsh	181.83	178.32	0.10	Good
140	Mid to Low Marsh	181.19	177.31	0.14	Good
150	Mid to Low Marsh	180.45	176.12	0.06	Good
160	Mid to Low Marsh	178.24	172.58	0.16	Good
165	Mid to Low Marsh	181.65	178.04	0.03	Good
170	High Marsh	196.63	202.02	0.09	Good
175	High Marsh	202.17	210.90	0.06	Good
180	High Marsh	201.54	209.89	0.07	Good
181	High Marsh	202.98	212.19	0.07	Good
182	High Marsh	201.86	210.40	0.13	Good
183	High Marsh	200.61	208.40	0.06	Good
184	High Marsh	194.96	199.35	0.18	Good

Table 5.9 *A MAT assessment of $WA_{(tol)}$ predictions for ALF-based $SWLI_{(Pred)}$ for samples from core ARN1-95-90*

The use of the ALF-based transfer function increases the range of matching analogues, so that only three samples are now without a modern equivalent. A second improvement is the ability of the ALF-based transfer function to produce lower $SWLI_{(Pred)}$ values than the previous transfer function, which are more consistent with the minerogenic nature of the sediments.

Three samples remain without modern analogues, and this may be caused by a number of factors. In some cases, fossil taxa are absent in the modern training set, which can arise when an environment is encountered in the fossil core that has not been sampled in the modern setting. This is most likely to be a low intertidal to subtidal situation since the training set possesses few samples from these environments. Alternatively, the fossil assemblage may reflect an environmental state which is not currently observed in the saltmarshes of the UK, perhaps as a result of changes in the rates of sea-level rise, sedimentation processes, tidal characteristics or wind-wave climate. A further cause may be inherent to the use of foraminiferal test linings. The exceptionally low settling velocity of test linings means that some are inevitably lost in the wetsplitting stage of sample preparation (see Appendix Two). This, coupled with any destruction of test linings prior to sample collection, will result in a tendency to underestimate the contribution of calcareous taxa. Nevertheless, this approach expands the range of environments which can be reliably reconstructed, and offers the potential for producing continuous records of changing water levels from fossil assemblages.

5.8 FORAMINIFERA AND SEA-LEVEL RECONSTRUCTION

The biostratigraphic analysis of fossil material within this thesis has two distinct aims:

1. To reconstruct water level variations from changing fossil foraminiferal assemblages in order to investigate lateral shifts in marsh palaeoenvironment;

2. To assign precise indicative meanings to radiocarbon dated sea-level index points in order to constrain vertical movements of relative sea-level.

To realise the first aim, the method employed to reconstruct changing water levels from fossil assemblages must be able to recognise and quantify a wide altitudinal range of depositional environments. In this instance, recognition of a sequence of change, even if only relative in nature, is of paramount importance. Conversely, the organic material suitable for producing precise sea-level index points is altitudinally restricted in its distribution, and so there is no need to consider such a range of altitudes. In this instance, high precision water level reconstructions are requisite.

The transfer function developed in Section 5.7.4 is a reliable method for reconstructing SWLIs between 130 to 220. In most marshes of study, this equates to a vertical range encompassing the whole vegetated marsh area and extending down onto the intertidal flats above mean low water neap tides. It therefore covers a large proportion of the marsh environment, including the transition from minerogenic to organogenic sedimentation, recommending it for use when reconstructing past water levels from changing fossil foraminiferal assemblages. From Figure 5.26a and c, it is apparent that scatter in the data increases with decreasing SWLI, and this decrease in predictive precision at lower elevations is concomitant with a reduction in the altitudinal sensitivity of foraminiferal assemblages. This arises because the low marsh to mudflat environment experiences smaller variations in environmental extremes, and conditions are more akin to the fully marine environment. Consequently, low marsh to mudflat assemblages are composed of species commonly observed subtidally, and their composition is strongly influenced by the neighbouring marine communities. This accounts for the fact that the lowest marsh to mudflat foraminiferal assemblages exhibit considerable geographic variability, whilst high marsh assemblages from different areas remain remarkably consistent (Scott & Medioli, 1980a). It would therefore be unwise to use the lowest marsh and mudflat species as precise proxies for water level, regardless of transfer function performance.

5.8.1 The High Marsh Environment

In Section 5.7.3.1 limitations in the performance of WA-type transfer functions in the highest marsh environment were outlined. These limitations are compounded by the nature of the foraminiferal assemblages present. The precise vertical zonation of the highest saltmarsh foraminiferal assemblages results from the fact that few species can tolerate the extremely hostile environment at the upper limit of marine influence. Whilst being of great potential use in the reconstruction of past water levels, the low abundance, low diversity nature of these assemblages renders them poorly suited to statistical analysis. In fact, all these high marsh samples were screened from the training set before development of the transfer function. This, coupled with the general behaviour of WA-type models at the extremes of the environmental gradient, means that the $SWLI_{(Pred)}$ values for the highest marsh samples are consistently minimum estimates.

Whilst these limitations are of little significance to general investigations of relative water level change through an entire core, they become potentially serious when attempting to produce precise sea-level index points from the highest marsh contexts. One way to address this problem is to utilise a combined approach employing statistical-based methods in association with a more subjective consideration of lithostratigraphy and the well-documented behaviour of saltmarsh foraminifera at the upper extremes of tidal influence.

The consistency in the highest marsh foraminiferal assemblages can be seen by examining the vertical distributions present at Arne Peninsula and Bury Farm (Figures 5.2 and 5.14). There is a clear interplay between *Jadammina macrescens* and *Miliammina fusca*, with the former dominating the highest marsh before being replaced by the latter at lower elevations. Variability in the exact percentage frequencies of each species present is inevitable, since the counts are so low that differences of only one or two tests will have large effects. The changing pattern of relative dominance from *J. macrescens* to *M. fusca* remains consistent however. This is a refinement of the procedure used by Thomas & Varekamp (1991), Varekamp *et al.*, (1992) and Nydick *et al.*, (1995), in their marsh palaeoenvironmental curves, where the percentage abundance of *J. macrescens* relative to 'other species' is considered.

Substantial errors may occur if interpretations are based on single samples with low counts, but the probability of such errors arising is reduced if interpretations are based upon a number of samples displaying a recognised trend. Identification of a low abundance fossil assemblage dominated by *J. macrescens* is in itself inconclusive, but if this forms part of a sequence from samples devoid of foraminifera, through a *J. macrescens* dominated zone into a *M. fusca* dominated zone, associated with lithostratigraphic changes indicating increased minerogenic sedimentation, a high marsh interpretation becomes more cogent. Analysis of the vertical interval over which such transitions occur today permits the errors associated with the transfer function to be minimised.

5.8.1.1 Application at Bury Farm

Figure 5.29 shows the interplay between *M. fusca* and *J. macrescens* on the present marsh surface. The decline in *M. fusca* numbers occurs gradually within a range of SWLIs that can be accurately predicted by the ALF-based transfer function. In fact the only assemblage which occurs at SWLIs above 220 is the low abundance, mononspecific assemblage of *J. macrescens*. Assuming that no changes in tidal regime have occurred through time, the transition between this low abundance assemblage and the barren freshwater deposits can be assigned a SWLI of 225±3.

5.8.1.2 Application at Arne Peninsula

Figure 5.30 shows the interplay between *M. fusca* and *J. macrescens* on the modern marsh at Arne Peninsula. Once again, the peak in *M. fusca* abundance occurs within the range of SWLIs accurately predicted by the ALF-based transfer function. The transition to a *J. macrescens* dominated assemblage above SWLIs of 236 is less abrupt than at Bury Farm and probably reflects the low tidal range of this system which imparts a greater significance to meteorological conditions (see Section 4.1). No living foraminifera are found above a SWLI of 254, and the exceptionally low abundance, mixed death assemblage found at these elevations must have been washed in during meteorologically enhanced high water. This increased vertical range reduces the precision to which predictions of SWLIs greater than 220 can be derived, but still provides an improvement to the persistent underestimates of the ALF-based transfer function which will be even more pronounced in this setting. Consequently, the mixed *J. macrescens* and *M. fusca* assemblage, where *J. macrescens* dominates, is assigned a SWLI of 246 with a SWLI error range of ± 26. From Figure 5.30, this range extends from the decline in *Miliammina fusca* through to the upper limit of marine influence.

5.8.2 Reconstruction Procedure

Figure 5.31 summarises the stages employed in the water level reconstructions that will be presented in Chapter Six. In the first stage, the ALF-based transfer function is applied to the whole core, using the classical deshrinking method. Samples producing $SWLI_{(Pred)}$ values between 146 and 209 are recalibrated using the inverse deshrinking method. Samples interpreted as being of highest marsh origin on the basis of combined lithostratigraphic and biostratigraphic evidence are re-examined using the site specific criteria described in Section 5.8.1.

This procedure is illustrated by applying it to the fossil core ARN1-95-90 and the results summarised graphically in Figure 5.32. The diagram shows the major species present followed by the ALF-based transfer function output, the MAT results, a summary of the reconstruction methods used, and a tendency diagram of relative sea-level movement. It should be noted that the tendencies are unaffected by the choice of reconstruction method except for the samples corrected using the high marsh visual zonation technique.

At the base of the core, the transfer function indicates a reduction in water depth followed by a submergence. This is clearly at odds with the lithostratigraphy and visual inspection of the foraminiferal assemblages which show a classic transition from a barren, freshwater peat into an overlying high marsh deposit dominated by *J. macrescens*. In these circumstances, the $SWLI_{(Pred)}$ would be assigned visually.

Between 160 cm and 95 cm the transfer function indicates relatively little change in water depth although a slight increase is intimated around 120 cm. This could be unreliable however, since MAT indicates that the sample is without a modern analogue.

A well defined submergence is identified between 95 cm and 90 cm, based largely on an increase in the percentage abundance of foraminiferal test linings. This inferred increase in water depth is immediately succeeded by an apparent shallowing. This is equivocal however, since the foraminiferal assemblages responsible for the increased $SWLI_{(Pred)}$ values are dominated by *Reophax* species and have no modern analogue. In fact, the reduction in organic content concomitant with the arrival of *Reophax* may be indicative of marsh submergence resulting in the arrival of subtidal species not encountered in the modern surveys.

Foraminiferal assemblages recovered from the upper 70 cm of the sediment core all possess good modern analogues and appear to indicate a progressive reduction in water depth. This emergence may have been interrupted by a submergence event around 20 cm. Whilst it is possible that this submergence is merely an artefact arising from the transition between classical and inverse deshrinking methods, the increase in *Miliammina fusca* in association with the brief return of foraminiferal test linings are consistent with a deepening at this time.

From this example it is clear that whilst $SWLI_{(Pred)}$ produced by the ALF-based transfer function can provide useful information on changes in water depth, it is vital that these values are checked by consideration of supporting lithostratigraphic and biostratigraphic data. This is particularly important when dealing with samples that possess no modern analogues. Since the modern training set possesses no reliable data from subtidal environments, the magnitude of submergence events predicted by the ALF-based transfer function may be significantly underestimated or not detected at all. For this reason, the transfer function is employed in Chapter Six to provide qualitative information on changing water depths and no attempt is made to precisely quantify these variations.

5.8.3 Llanrhidian Marsh

The absence of an unequivocal vertical zonation at Llanrhidian Marsh restricts its potential for the production of high resolution sea-level index points. It is inappropriate to apply the transfer function developed above to an environment where the controls on foraminiferal distribution are clearly different.

The range of SWLI covered at Llanrhidian extends from 239 at the summit of the ridge, down to 187 on the tidal flat just seaward of the marsh fringe. At comparable SWLI values, the marshes at Arne Peninsula and Bury Farm are characterised by a dominance of *J. macrescens* and an absence of calcareous foraminifera. The assemblage at Llanrhidian is not dissimilar in that it too is dominated by *J. macrescens*. The major difference is the presence of calcareous foraminifera at much greater elevations than encountered in the other two marshes, and the low abundances of *M. fusca* which never contribute more than 5% of any sample.

The challenge remains to distil useful altitudinal information from a marsh essentially dominated by one taxon. Of primary use in this respect is the percentage abundance of *Haplophragmoides* species which rise from background levels of less than 7% above +4.68 m OD, up to 50% between +4.68 m OD and +4.44 m OD (Figure 5.20). Numbers must decline rapidly below this level because the next sample down (+4.16 m OD) is virtually devoid of *Haplophragmoides* species. It is interesting to note that the peak abundances of *M. fusca* are also found within this zone. The foraminiferal zonation at Arne sets a precedent for this where the same sequence is observed (Figure 5.2).

Another potentially useful indicator is the presence of thecamoebians within a sample. Thecamoebians are freshwater testate amoebae traditionally believed to be restricted to the supratidal environment (e.g. Scott & Medioli, 1980a). Whilst recent research indicates thecamoebians are abundant in the high marsh environment, they are dominantly smaller than 63μm and so are rarely apparent in foraminiferal analyses (Charman *et al.*, 1998). Identification of thecamoebians in a foraminiferal sample theoretically places it close to the upper limit of marine influence. This idea is supported by the limited occurrences of thecamoebians on the modern marsh, where they are restricted to altitudes of +4.89 m OD and above (Figure 5.20).

Combination of these two factors thereby presents a rudimentary altitudinal framework into which fossil foraminiferal samples can be placed. Once again, individual samples may not in themselves be diagnostic, but a sequence of change across a number of samples, coupled with changes in lithostratigraphy, add extra credence to the interpretations. A schematic fossil sequence is presented in Figure 5.33. The highest altitude environments will be characterised by thecamoebians and an absence of foraminifera (Zone A). Travelling down the marsh, these will be superseded by assemblages dominated by *J. macrescens* with low abundances of *Haplophragmoides* species, some *Trochammina inflata*, and rare thecamoebians, indicative of SWLIs ranging from 220 ± 1 to a minimum estimate of 240 (Zone B). This assemblage is replaced by one devoid of thecamoebians, but with elevated abundances of *Haplophragmoides* species and low numbers of *M. fusca*, indicative of SWLIs between 204 ± 4 and 220 ± 1 (Zone C). Finally, the sequence moves to an assemblage dominated once again by *J. macrescens* but with an absence of thecamoebians *T. inflata*, and *Haplophragmoides* species, indicative of SWLIs below 204 ± 4 (Zone D).

5.9 SUMMARY

- The modern foraminiferal death assemblages at Arne Peninsula, Newton Bay, and Bury Farm are strongly related to altitude;
- The elevation with respect to the tidal frame of all sample stations is standardised *via* the standardised water level index (SWLI) of Horton (1997);
- An ALF-based transfer function relating foraminiferal assemblages to a SWLI is developed from a dataset comprising 123 modern foraminiferal samples from 11 UK saltmarshes;
- This transfer function is tested by calibrating fossil foraminiferal samples to produce example $SWLI_{(Pred)}$ values from saltmarsh cores;
- A combined statistical and visual multi-proxy approach is developed to reconstruct relative sea-level change at Arne Peninsula, Newton Bay, and Bury Farm;
- The modern foraminiferal death assemblage distribution at Llanrhidian Marsh is strongly influenced by factors other than elevation relative to mean tide level;
- The transfer function cannot be reliably applied to fossil material from Llanrhidian Marsh, and so a rudimentary altitudinal framework is constructed using a visual, multi-proxy approach.

Late Holocene Relative Sea-Level Change in Southern Britain

In Chapter Five, a transfer function for reconstructing SWLIs was developed from surveys of modern saltmarsh foraminifera. In this chapter, the transfer function is used to calibrate the fossil foraminiferal assemblages presented in Chapter Four and Appendix Two. This provides information on changes in water depth through the saltmarsh sediments that, in association with the lithostratigraphic data, are used to infer phases of marsh inundation and emergence. These lateral shifts in depositional environment are placed in a vertical and temporal framework *via* the development of sea-level index points from AMS radiocarbon dated material (Appendix Three). Comparison of the sequences of change from a number of different sites permits identification of regional *versus* site specific signals which are used to construct a record of late Holocene sea-level change.

This chapter comprises the following:

- A description of water level changes inferred from fossil foraminifera at Arne Peninsula;
- Combination of these lateral shifts in environment with vertical movements of mean tide level derived from sea-level index points to produce a pattern of relative sea-level change at this site;
- Compilation of a similar record from Newton Bay against which the relative sea-level record derived from Arne Peninsula can be compared;
- Combination of the records from Newton Bay and Arne Peninsula to infer relative sea-level change within Poole Harbour as a whole;

- Comparison of the sequence of change recorded in Poole Harbour with that from Bury Farm, Southampton Water, to gauge the spatial extent and coherence of these variations;

- An analysis of vertical movements in relative sea-level inferred from sea-level index points at Llanrhidian Marsh, and a comparison with the record from the south coast sites.

6.1 ARNE PENINSULA

Variations in water level through a single core are inferred from a combination of lithostratigraphic and biostratigraphic data. The spatial extent of these changes is then investigated by comparing this record with those derived from other cores recovered at Arne Peninsula. These records are compiled to produce an indication of site-scale changes in relative sea-level which are dated using six radiocarbon dates and two biostratigraphic chronohorizons.

6.1.1 Water Level Changes Inferred from ARN1-95-90

The core ARN1-95-90 is presented in Figure 4.4 as a typical example of the lithostratigraphy and biostratigraphy of Arne Peninsula, and has been used to check the applicability of the ALF-based transfer function in Section 5.7.4.1. These factors recommend it as an ideal starting place to investigate water level changes at this site.

The lithostratigraphy (Appendix One) comprises a humified basal peat which, on the basis of its pollen content (Long *et al.*, in press) and the absence of foraminifera, is interpreted as being of freshwater origin. This grades into a marine silt-clay around 60 cm in thickness, often containing sand and broken shells. This minerogenic facies is succeeded by a 15 cm thick organic silt-clay with some detrital herbaceous material. A thin minerogenic intercalation less than 10 cm thick separates this organic silt-clay from the overlying 75 cm of *turfa*-rich silt-clay with *Spartina* remains.

Taking organic content as a crude index of water level, with organic sediments indicating higher marsh environments than more minerogenic mudflat deposits, the lithostratigraphy suggests two submergence events, the first prolonged and the second brief, coupled with two emergence events, the first brief and the second prolonged.

The variations in $SWLI_{(Pred)}$ produced by the transfer function are shown in Figure 6.1. The first, prolonged submergence event recorded in the lithostratigraphy is also apparent in the biostratigraphy (S1), although the transfer function reconstructs this as a two-stage event, with a second, abrupt decrease in $SWLI_{(Pred)}$ between 120 and 130 cm (S1b). This brief submergence relates to a pronounced peak in percentage frequencies of *Ammoscalaria runiana* and a small peak in percentage frequencies of test linings. This sample is without a modern analogue however, and consequently this second stage of submergence is equivocal. A decline in test linings succeeds this, associated with a reduction in *A. runiana* and an increase in *Miliammina fusca*, which is interpreted as an emergence event by the transfer function (E1). This reduction in water depth precedes the transition to organic-rich sedimentation.

The transfer function discerns a second submergence event centred between 85 and 100 cm that precedes the lithostratigraphic transition from organic to minerogenic sedimentation (S2). Between 60 and 85 cm the transfer function suggests a reduction in water depth but MAT indicates that the foraminiferal assemblages possess no modern analogues. Visual inspection of these assemblages suggests that they form the second part of the submergence event that began with S2 at *c.* 100 cm. Within this part of the sedimentary sequence a number of agglutinated taxa arise that are not encountered in the modern marsh fauna of Arne Peninsula. Most prominent among these taxa are *Reophax* species that reach a peak in the minerogenic intercalation and are associated with *Ammobaculites balkwilli* and *Eggerelloides scaber.* Murray (1991) indicates that such assemblages are typical of brackish lagoons supplied with organic detritus and this is consistent with the nature of the detrital organic facies between 85 cm and 100 cm. The lack of modern analogues means that it is unclear whether this faunal transition is accompanied by an increase in water depth although the reduction in organic material is consistent with this interpretation. Furthermore, low abundances of *E. scaber* immediately precede the S2 event and suggest that the E1 emergence may be equivocal.

Above the *Reophax* assemblage, the transfer function predicts a progressive reduction in water depth as *M. fusca* and test linings reduce in number whilst percentage frequencies of *Jadammina macrescens* and *Trochammina inflata* increase (E2). This faunal transition coincides with a change in lithostratigraphy to a *turfa*-rich silt-clay with *Spartina* remains. A final increase in water depth is intimated around 20 cm depth by a slight increase in the percentage frequencies of *M. fusca* and foraminiferal test linings (S3).

In summary, the fossil foraminiferal assemblages from ARN1-95-90 indicate:

1. An initial, possibly two-stage submergence (S1);

2. A possible emergence (E1) based largely upon a reduction in the abundance of foraminiferal test linings;

3. A submergence event (S2) associated with the inception of a **lagoonal phase** characterised by detrital organic sedimentation leading into a fine-grained minerogenic deposit characterised by a low abundance *Reophax* assemblage;

4. An emergence (E2) concomitant with the arrival of sediments containing *Spartina* macrofossils that is possibly interrupted by a brief submergence event (S3).

6.1.2 Testing the Sequence from ARN1-95-90

A single core in isolation provides information about changes in one part of the marsh system, but cannot be reliably used to interpret site scale variability unless the spatial consistency of the record can be demonstrated. The sequence of water level changes developed from ARN1-95-90 is therefore tested by comparing them with lithostratigraphic and biostratigraphic data from a number of locations across the marsh.

The summary lithostratigraphy is presented in Figure 4.3 and discussed in Section 4.1.4.1, but a number of salient points are reiterated here. The bedrock surface underlying the saltmarsh at Arne Peninsula exhibits a stepped profile, with an abrupt break in slope descending around 50 cm in height from approximately -0.6 m OD to around -1.1 m OD. Below -1.1 m OD, seaward of this break, organic-rich and humified basal peat deposits are widespread, whilst on the bedrock surface above this altitude, such deposits are absent. The minerogenic sediments overlying the basal peats commonly contain sand and suggest that these facies may be associated with higher energy, possibly erosive conditions. Indeed, in some areas the upper contacts of the basal peats are eroded. Another laterally persistent feature of the lithostratigraphy is a thin organic horizon recorded at around -0.4

m OD which is associated with the lagoonal phase recorded in ARN1-95-90. Finally, the upper portion of the sedimentary sequence consists of wet, *turfa*-rich silt-clay with *Spartina* remains, which Long *et al.* (in press) suggest is associated with the arrival of *Spartina anglica* documented around 1899 AD. The propensity of *S. anglica* to trap sediment resulted in greatly accelerated accretion rates and the rapid expansion of saltmarshes in Poole Harbour at this time (Section 4.1.2.1). The thickness of this '*Spartina* facies' increases to seaward, reflecting the invasion of previously unvegetated mudflats and the lateral migration of saltmarsh communities. In ARN1-95-90, a rise is *S. anglica* pollen is noted around 75 cm corresponding to an increase in the organic content of the sediment (Long *et al.*, in press).

Figure 6.2 shows the $SWLI_{(Pred)}$ derived from the transfer function plotted against altitude for cores ARN1-95-80 and ARN1-95-90. The initial submergence event recorded in ARN1-95-90 (S1) is also observed in ARN1-95-80, although the presence of preserved calcareous tests within the silt-clay implies that the submergence may have been larger than ARN1-95-90 suggests. Whilst the transfer function is incapable of distinguishing between calcareous dominated assemblages, test preservation strongly argues against deposition in an acidic low marsh setting.

The transfer function then indicates marsh emergence (E1) between *c.* -0.7 m OD and -0.4 m OD in ARN1-95-80. The emergence E1 is better represented here than in ARN1-95-90 although it is primarily generated on the basis of a reduction in the abundance of test linings. Once again this culminates in the deposition of a detrital, lagoonal facies.

Emergence event E1 is succeeded by an apparent increase in water depth (S2) within the lagoonal, detrital organic facies, although *E. scaber* is absent from these sediments. The transition into the *Reophax* assemblage is once again synchronous with a switch to minerogenic sedimentation. The presence of calcareous tests in association with the *Reophax* assemblage provides further support for the suggestion that the reduced organic content of the sediments resulted from a substantial increase in water depth at this time.

Once again, the termination of the *Reophax* phase is well defined, occurring around 0 m OD and followed by the progressive emergence event E2. It should also be noted that the final submergence event (S3) recorded at *c.* 0.5 m OD is also recorded in ARN1-95-80, and appears to be more than a localised phenomenon, or an artefact of the transfer function.

In Figure 6.3, the spatial scale is expanded to include five cores from more landward portions of the saltmarsh. ARN4-96-60 is located at the edge of the stepped bedrock profile and its basal sediments are dominantly sandy. The transfer function indicates that this is a deeper water deposit and its altitude clearly links it with the end of S1 and the start of emergent phase E1.

E1 appears to reach a climax just below -0.5 m OD, where the transfer function indicates an increase in water depth (S2). This coincides with the lagoonal phase which is well represented by *E. scaber* within the detrital unit, and succeeded by the arrival of the low abundance *Reophax* phase. Above *c.* 0 m OD, the E2 phase commences although there is no recognition of the S3 submergence in this area.

There is only a limited amount of data available from ARN4-95-50 but this also demonstrates the onset of S2 within the detrital organic unit, preceding the arrival of the *Reophax* phase.

ARN1-95-40 records the arrival of the *Reophax* phase and its abrupt termination associated with a reduction in water depth (E2). The two most landward cores indicate emergence around 0.6 m OD, and this is suggested to relate to E2 which is recorded at a higher altitude owing to the slower sedimentation rates experienced in the high marsh. Once again there is evidence for the S3 event in the topmost samples.

6.1.3 Changes in Mean Tide Level

Since SWLIs provide a direct measure of height above mean tide level, reconstructing palaeo-mean tide level is a simple matter of subtracting this elevation from the current altitude of the sample material. This procedure is applied to all fossil foraminiferal samples from Arne Peninsula to produce Figure 6.4, which shows the relationship between sample altitude and palaeo-mean tide level (PMTL). The relationship is broadly linear until a PMTL of -1.2 m OD where the gradient of the line begins to reduce. This indicates that apart from the region below -1.2 m OD, where the rate of mean tide level rise was greater than the rate of sedimentation, the two appear to have been broadly in equilibrium.

This implies that sedimentation rates adjust rapidly to variations in the rate of relative sea-level rise.

The *S. anglica* sediments above 0.4 m OD are known to be associated with greatly accelerated rates of sedimentation. Whilst there is some evidence of a small deflection to the left of the general trend, and an increase in the proportion of high marsh samples, this enhanced sedimentation rate is not clearly reflected by the transfer function. The foraminiferal assemblages associated with the *Spartina* sediments are of very low abundance, possibly reflecting the 'diluting' effect of rapid sedimentation. The lack of evidence for an acceleration in relative sedimentation rate derived from the transfer function may indicate that the arrival of *S. anglica* altered the marsh environment to such an extent that the behaviour of foraminiferal assemblages was affected. The fact that most of the fossil assemblages possess good modern analogues however, suggests that modern, post-*Spartina* vertical distributions are comparable to those of the pre-*Spartina* environment. An alternative explanation is that the graph indicates an acceleration in the rate of relative sea-level rise which, after causing an initial inundation, was matched by equally rapid *Spartina*-driven marsh accretion. It is entirely possible that the rapid spread of *S. anglica* was facilitated by an acceleration in the rate of relative sea-level rise since this would create an abundance of unvegetated inter-tidal environments and lead to increased environmental stress on the native flora which are known to be less tolerant of prolonged tidal submergence.

6.1.4 Vertical Movements of Relative Sea-Level

Six radiocarbon dates are available from Arne Peninsula, and these are presented in Figure 4.3 and Table 4.2. The transfer function is used to calibrate the altitude of PMTL for each dated sample and produce sea-level index points which are evaluated and screened in Appendix Three. On the basis of this analysis, three index points are screened out and the remaining three are presented in Figure 6.5, along with two biostratigraphic markers identified by Long *et al.* (in press), and outlined in Section 3.5.2. The rise in *Pinus* pollen, dated to 200 ± 50 Cal. BP reflects the documented expansion of coniferous plantations in the region, and the rise in *S. anglica* pollen, associated with the change in lithostratigraphy, is dated to *c.* 50 Cal. BP.

The removal of 50 % of the radiocarbon dated sea-level index points means that Figure 6.5 provides little detailed information on the timing of relative sea-level change at Arne Peninsula. All that can be deduced from this age-altitude information is a rise in PMTL during the last 3000 years, with an indication of an acceleration in the rate of relative sea-level rise after *c.* 200 Cal. BP.

6.1.5 Synthesis of Data from Arne Peninsula

The combined lithostratigraphy and biostratigraphy indicate that the saltmarsh sediments have accumulated under a range of water depths and, by implication, marsh accretion at Arne has not always been in equilibrium with sea-level movements. Phases of submergence or emergence record the lateral translation of depositional subenvironments in response to variations in the rate of sedimentation, fluctuations in the rate of relative sea-level rise, alterations to the hydrographic character of the marsh, and changes in wind/wave climate or coastal geometry (Section 2.5).

The sea-level index points presented in Section 6.1.4 fix the position of mean tide level vertically and temporally, and permit comparison of its vertical movements with the lateral translations identified above. Figure 6.4 is used to assign an approximate altitude of PMTL from the altitude of the fossil deposit under consideration. In this way, the marsh submergence and emergence events can be placed in a temporal framework by inspection of Figure 6.5.

Submergence event S1 is recorded between *c.* -1.1 m OD and *c.* -0.5 m OD in the fossil deposits. Inspection of Figure 6.4 suggests that this corresponded to a PMTL of between -1.7 m OD and -1.0 m OD. From Figure 6.5, it can be seen that this equates to the early phase of rising mean tide level and, from ARN#3 and ARN#5, this occurred sometime between *c.* 3300 Cal. BP and *c.* 1200 Cal. BP.

Emergence event E1 occurs between -0.5 m OD and -0.3 m OD which from Figure 6.4 equates to a PMTL of *c.* -1.0 m OD to -0.7 m OD. From ARN#3, this emergence began *c.* 1200 Cal. BP and was succeeded by the establishment of a brackish lagoon. Similar lagoonal deposits are recorded in saltmarshes from other parts of the Harbour (Long, *pers. comm.*; Section 6.2) suggesting that this event may have occurred throughout much of the system.

Faunal evidence indicates a second submergence event (S2) occurred during the lagoonal phase sometime after *c.* 1200 Cal. BP. This is consistent with the altitudinal offset between ARN#3 and ARN#1. It must be stressed however that the low $\delta^{13}C$ value of ARN#3, whilst consistent with deposition in a freshening environment, means that the age estimate of this date could be in error. Moreover, if this date does come from the onset of lagoonal conditions, it is possible that the $SWLI_{(Pred)}$ is too high, since the transfer function fails to predict accurately the other lagoonal samples. This means that the sea-level index point may also be plotting too low. This would serve to reduce the rise in mean tide level between *c.* 1200 Cal. BP and *c.* 300 Cal. BP. The *Reophax* assemblage, which is associated with a continued increase in water depth in the latter stages of S2, is delimited in time by the two floral chronohorizons, indicating it must have occurred between *c.* 200 Cal. BP and 50 Cal. BP.

The E2 emergence that succeeds the *Reophax* phase is linked to the arrival of *Spartina* on pollen and lithostratigraphic grounds, and has therefore occurred during the last century. This implies that mean tide level has risen between 25 cm and 45 cm in the last 100 years, corresponding to a rate of between 2.5 to 4.5 mm a^{-1}. This value is comparable to rates of sediment accretion recorded in other saltmarshes from the Solent region by Cundy & Croudace (1996) who dated the sediments using a variety of short-lived radionuclides (Section 4.1.3).

The final proposed submergence event (S3) is located at 0.5 m OD, equivalent to 30 cm below marsh surface in ARN1-95-90. Pollen data from this core indicates that above 25 cm, the abundance of *S. anglica* pollen decreases, perhaps in association with die-back (see Section 4.1.2). It is possible therefore, that S3 corresponds to a weakening in *S. anglica* growth and a concomitant reduction in sedimentation rate that resulted in the marsh being incapable of keeping up with relative sea-level rise during this period.

On the basis of these multiple lines of evidence, the following preliminary sequence of change is apparent at Arne Peninsula:

1. The early part of the record is characterised by rising relative sea-level inundating the basal peat deposits. Whilst impossible to infer accelerations in the rate of relative sea-level rise from the inundation of basal peat deposits, the evidence that water depth

increased markedly during the S1 submergence suggests that the rise may have been comparatively rapid and/or accompanied by erosion or non-deposition;

2. This is followed by the E1 emergence inferred from the biostratigraphy which occurred *c.* 1200 Cal. BP, perhaps reflecting a reduction in the rate of relative sea-level rise. The stepped nature of the bedrock surface and the associated sand layers above the lower peats adds some support for the idea of reduced relative sea-level rise, or perhaps even a relative sea-level fall at this time. Slowly changing sea level would serve to concentrate erosive forces within a limited vertical range, leading to destruction of basal deposits and incision into the Tertiary bedrock;

3. The E1 emergent phase culminates with a marked change in the environment at Arne Peninsula. The development of a brackish lagoon is indicated by the foraminiferal data, and supporting evidence suggests that this event may have occurred in other parts of the system. The decline in the rate of relative sea-level rise, or a fall in relative sea-level proposed in (2), would reduce the tidal prism and volume of marine water entering Poole Harbour. This would cause a relative increase in the contribution of freshwater inputs to the Harbour giving rise to shallow, brackish water conditions consistent with the foraminiferal evidence. Alternatively, the E1 event may reflect a reduction in marine influence caused by constriction of the Harbour entrance resulting from the extension of one or both of the spits flanking the Swash Channel (Section 4.1). Whilst a slow-down or fall in relative sea-level could instigate spit progradation, a change in the regime of relative sea-level rise is not essential and it is possible that the development of this lagoonal phase was solely the result of changes in coastal morphology;

4. A renewal in relative sea-level rise is suggested sometime after *c.* 1200 Cal. BP (S2) during the lagoonal phase and is associated with the introduction of abundant organic detritus that supported detritovores such as *E. scaber*. It is possible that relative sea-level rose by around 60 cm between *c.* 1200 BP and *c.* 300 Cal. BP at a long-term rate of around 0.7 mm a^{-1}. The S2 event could also reflect a widening of the Harbour mouth and the progressive re-introduction of marine influence into the system;

5. Between *c.* 200 Cal. BP and *c.* 50 Cal. BP, the organic phase of lagoonal sedimentation ends and minerogenic deposition associated with a low abundance agglutinated

Reophax assemblage occurs. This may reflect increased water depth resulting from an acceleration in the rate of relative sea-level rise during the latter stages of the S2 submergence. Alternatively, a more open communication with the sea may have increased the flushing of the Harbour causing the organic detritus to be removed more effectively. This latter explanation seems unlikely however given that cartographic evidence suggests the Harbour entrance has been stable for at least the last 400 years (Section 4.1.1);

6. Any increase in water depth associated with the end of S2 was short-lived and the arrival of *Spartina* at the beginning of the 20th Century resulted in a dramatic increase in sedimentation rate which outpaced relative sea-level rise, producing the emergent trend E2;

7. In the last few decades, there is evidence for a renewed submergence of the saltmarsh (S3), perhaps reflecting a further increase in the rate of relative sea-level rise, or alternatively responding to a decrease in sedimentation rate associated with *Spartina* die-back.

This provisional sequence of events, whilst internally consistent, cannot be considered as a reliable indicator of sea-level movement, since it may reflect processes particular to the marsh at Arne Peninsula. In order to investigate the spatial extent of such changes, the record from a second marsh within Poole Harbour must be examined to identify similarities and differences in the records.

6.2 NEWTON BAY

The marsh at Newton Bay is situated around 3 km from Arne Peninsula and is nearer to the mouth of Poole Harbour (Figure 4.1). This difference in location, coupled with a northerly facing aspect (as opposed to the easterly facing marsh at Arne) enables the record at Arne Peninsula to be evaluated for the significance of site specific marsh processes and the possible importance of hydrographic setting (such as enhanced tidal wave deformation or fetch).

6.2.1 Water Level Changes Inferred from NEB2-96-60

The core NEB2-96-60 is presented in Figure 4.7 as an example of the typical lithostratigraphy and biostratigraphy of the Newton Bay marsh. For this reason it is selected as a starting place from which to examine water level changes at this site.

The lithostratigraphy consists of a dark brown to black, humified basal peat in which are found a 4 cm thick clay-rich band at 175 cm, and a 1 cm thick sand layer at 162 cm. The upper contact of this peat bed is erosive, and is overlain unconformably at 158 cm by a grey sandy-silt with some clay, which in turn is succeeded by a grey sand extending between 98 cm and 70 cm. Between 70 cm and 28 cm, the sediments consist of a brown-grey silty-clay with some *turfa*, and the remainder of the sequence is similar in general composition but with an increased *turfa* content.

As in Section 6.1, by taking organic content as a crude indicator of water depth, the lithostratigraphy indicates that the marsh was inundated around 158 cm, but that the record of this submergence has been obscured by erosion and deposition of a high energy sand deposit. The two minerogenic layers within the top of the peat provide some evidence for short-term increases in water depth before the final inundation. Lower energy conditions return, above the erosive contact, although the sediments still possess a sand fraction, until around 98 cm when a second sand layer appears. The upper 70 cm of the marsh possess finer grained sediments with increasing organic content, indicating a progressive shallowing of water depth.

$SWLI_{(Pred)}$ for samples from NEB2-96-60 produced by the transfer function are presented in Figure 6.6. The foraminiferal content of the thin clay band at 175 cm is very low, but high numbers of test linings are present. The transfer function interprets this as evidence for a substantial submergence at this time, and accounts for the low $SWLI_{(Pred)}$ at the base of the core. The transition from peat into the overlying minerogenic sediment is accompanied by a progressive increase in percentage frequencies of *M. fusca* and *A. runiana*, coupled with an increase in the number of test linings and a decrease in the abundance of *J. macrescens*. The transfer function interprets this as indicating an increase in water depth, denoted in Figure 6.6 as S1.

The first emergence E1 occurs above 130 cm where a decrease in the abundance of test linings is associated with an abrupt increase in the number of tests recovered. There is little change in the agglutinated assemblage, and it is therefore difficult to assess how reliable the E1 event is.

The silt-clay facies is truncated by the emplacement of a large sand body at *c.* 100 cm, that is virtually devoid of foraminifera. The high number of test linings present in this unit is interpreted by the transfer function as indicating a submergence event. Whilst this is a possible interpretation, the same signal would arise from the incorporation of calcareous foraminifera in the sand, and could therefore equally have resulted from a storm without the need to infer marsh inundation. Since an implicit assumption in the interpretation of foraminiferal assemblages is that they are *in-situ*, I consider it unjustified to assign a $SWLI_{(Pred)}$ to the sand deposits.

When lower energy depositional conditions resume, the characteristic *Reophax* assemblage identified at Arne Peninsula is present, with associated peaks in *E. scaber* and *A. balkwilli*, perhaps indicative of post-lagoonal marsh submergence (S2). Above the *Reophax* assemblage there is a progressive reduction in water depth indicating marsh emergence at this time (E2).

6.2.2 Testing the Sequence from NEB2-96-60

The lithostratigraphic succession at Newton Bay typically possess two to three sand layers, suggestive of periods dominated by higher energy conditions than are observed today. In many cases, erosion has removed portions of the record and consequently there is considerable variability between cores from this site. This makes it less likely that similar sequences of events will be recognised in every core. Instead, the record of change will be fragmentary, requiring piecing together from a number of cores.

This spatial heterogeneity is reflected in the biostratigraphy, and the $SWLI_{(Pred)}$ show differing trends. Supporting evidence for S1 and E1 is lacking, largely because only NEB2-96-40 has samples from a comparable age/altitude (Figure 6.7). The counts below 135 cm in this core are very low and could account for the transfer function's inability to discern any changes in this area.

The large sand layer identified in NEB2-96-60 is well documented in cores NEB2-96-40, and NEB2-96-20, and similar layers are also apparent at comparable altitudes in the more landward cores. It is possible that these sand layers all correspond to the same period of formation and perhaps even a single high magnitude event. Examination of the lithostratigraphy (Appendix One) demonstrates a considerable degree of spatial variability in the occurrence and number of sand layers however, possibly indicating a more prolonged phase of periodic sand deposition. This could result from a rise in relative sea-level enhancing erosion of cliff material, an increase in the frequency or intensity of storm events, or from changes in the coastal geometry of the Harbour mouth caused by spit progradation and erosion.

The *Reophax* assemblage is consistently observed above the sand layer in cores NEB2-96-40, and NEB2-96-20. It is also present in NEB3-96-70 where it occurs above an organic intercalation characterised by *E. scaber* similar to that observed at Arne Peninsula. This could indicate that the early lagoonal phase at Newton Bay was terminated by an abrupt switch to high energy, erosive conditions that removed much of the evidence for its existence and emplaced the large sand layer.

The upper portions of cores NEB2-96-20, NEB2-96-40, and NEB2-96-60 all exhibit the emergent trend E2 associated with fine grained sedimentation indicating the return to relatively tranquil depositional conditions.

6.2.3 Changes in Mean Tide Level

The displacement of mean tide level is determined as in Section 6.1.3, and presented in Figure 6.8. As at Arne Peninsula, the points form an approximately linear distribution, although the scatter is greater and there is no evidence for accelerated mean tide level rise in the lower altitude samples. The submergence event related to the sand layer is evident between -0.5 m OD and -0.1 m OD, corresponding to an estimated palaeo-mean tide level of between *c.* -1.0 m OD and *c.* -0.6 m OD.

6.2.4 Vertical Movements of Relative Sea-level

Six radiocarbon dates are available from Newton Bay, and these are presented in Figure 4.6 and Table 4.3. As in Section 6.1.4, the transfer function is used to calibrate the altitude of palaeo-mean tide level for each dated sample, and the resulting sea-level index points, which document the changing altitude of mean tide level through time, are displayed in Figure 6.9. The screening procedure documented in Appendix Three removed one radiocarbon date, although it should be noted that NEB#2 may be unreliable owing to a very low organic content.

From Figure 6.9, it is apparent that PMTL rose from *c.* -2.5 m OD around 4700 Cal. BP to *c.* -2.1 m OD by around 3700 Cal. BP. At some time after this, the rate of relative sea-level rise increased, rising to *c.* -1.0 m OD by *c.* 2400 Cal. BP. After 2400 Cal. BP there is an apparent reduction in the rate of relative sea-level rise, with PMTL remaining between -1.0 m OD and -0.5 m OD until *c.* 200 Cal. BP. In the last two centuries, the rate of relative sea-level rise must have increased again to elevate mean tide level to its current position.

6.2.5 Synthesis of Data from Newton Bay

Despite the complex lithostratigraphic record preserved at Newton Bay, the biostratigraphy indicates a consistent pattern of changing water depth:

1. An initial submergence (S1) is noted between -1.3 m OD and -0.9 m OD, corresponding to a PMTL of between -1.8 m OD and -1.5 m OD. From Figure 6.9, this equates to the period of rising relative sea-level between *c.* 4700 Cal. BP and *c.* 3000 Cal. BP;

2. Evidence for an emergence (E1) between -0.9 m OD and -0.6 m OD (PMTL -1.5 m OD to -1.3 m OD) is restricted to one core, but could indicate that the reduction in the rate of relative sea-level rise predates 2400 Cal. BP. The long term rate of relative sea-level rise was low between *c.* 2400 Cal. BP and 200 Cal. BP, and PMTL persisted between -0.5 m OD and -1.0 m OD throughout this period. As at Arne, there is a step in the bedrock at Newton Bay, this time found between -0.4 m OD and -1.2 m OD, corresponding closely to the position of PMTL during this slow phase of relative sea-level rise;

3. Deposition of the large sand layer or layers is reasonably well constrained in time by the radiocarbon dates. NEB#4 predates their deposition, indicating emplacement after *c.* 2400 Cal. BP. NEB#2 is taken from above a sand layer at the back of the marsh and suggests that this may have been deposited prior to *c.* 900 Cal. BP. The very low organic content of this date means that it may be unreliable, but assuming it is accurate, there is evidence for at least one period of increased energy conditions at Newton Bay between *c.* 2400 Cal. BP and *c.* 900 Cal. BP. This phase of sand deposition correlates

well with the reduction in the rate of relative sea-level rise and could be associated with the bedrock erosion postulated in (2);

4. The *Reophax* assemblage that occurs above the sand in a number of cores is found at the same altitude as the well-dated assemblage at Arne Peninsula and, as the uneroded core NEB3-96-70 indicates, in precisely the same stratigraphic context. On the basis of these facts and the unique and distinctive composition of the assemblage, it is reasonable to suggest that this provides further evidence for the widespread nature of this lagoonal phase, confirming it was not peculiar to the marsh at Arne Peninsula;

5. The age-altitude data indicates that there must have been an acceleration in the rate of relative sea-level rise during the last 200 years to elevate PMTL from its position *c.* 80 cm below current mean tide level. This equates to a rate of relative sea-level rise of *c.* 4 mm a^{-1} which is comparable to the results from Arne Peninsula (Section 6.1.5) and the radionuclide investigations of Cundy & Croudace (1996).

6.3 Relative Sea-Level Change in Poole Harbour

Lithostratigraphic, biostratigraphic, and chronostratigraphic data from two contrasting sites in Poole Harbour have been presented in Sections 6.2 and 6.3. The combination of these data permits the relative sea-level changes experienced within the Harbour as a whole to be investigated.

6.3.1 Vertical Movement of Mean Tide Level

The eight radiocarbon dated sea-level index points in association with the two floral chronohorizons are presented in Figure 6.10, normalised to account for the difference in mean tide level between Newton Bay and Arne Peninsula (mean tide level at Newton Bay is 13 cm below that at Arne Peninsula). NEB#2 and NEB#4 fill an important gap in the record from Arne Peninsula, but are consistent with the idea of low long-term rate of relative sea-level rise prior to *c.* 1200 Cal. BP. NEB#1 shows good agreement with ARN#1 and the *Pinus* rise index point, whilst NEB#5 is in close accord with ARN#5. Together they refine the view of relative sea-level change in Poole Harbour.

It is now clear that the period from *c.* 4700 Cal. BP to *c.* 2400 Cal. BP was characterised by rising relative sea-level, perhaps with a slight acceleration after *c.* 3500 Cal. BP. From 2400 Cal. BP to *c.* 1200 Cal. BP, PMTL appears to have remained relatively constant at around 1 m below modern mean tide level perhaps even falling slightly. Between *c.* 1200 Cal. BP and *c.* 900 Cal. BP, there is an apparent increase in the rate of relative sea-level rise which then slows after 900 Cal. BP until a final acceleration in the last couple of centuries.

6.3.2 Changes in Saltmarsh Sedimentation

Despite the differing stratigraphies of the two sites they possess some common features. Both show an initial submergence (S1) that is associated with the inundation of basal peat deposits generally older than 3000 Cal. BP and situated seaward of a distinct step in the underlying Tertiary bedrock. This appears to be an erosive feature, cut during a period of stable to falling relative sea-level and associated with an emergent trend (E1). In both marshes, this reduction in water depth culminates in the deposition of detrital organic lagoonal facies that is overlain conformably by deeper water, minerogenic sediments at Arne Peninsula (S2). A similar transition is apparent locally in Newton Bay although a large proportion of the organic lagoonal phase appears to have been eroded and replaced by a sand layer. The latter stages of the S2 submergence associated with the arrival of *Reophax* species occurred during a period of rising relative sea-level indicated by the age-altitude data (Figure 6.10) at time when the Harbour mouth was known to be stable (Section 4.1.1). A final emergence is recorded in the upper portions of the stratigraphy at both sites (E2).

6.3.3 Synthesis of Change in Poole Harbour

Consideration of the age-altitude information provided by the sea-level index points, coupled with the information on changing depositional conditions on the saltmarsh gleaned from combined lithostratigraphy and biostratigraphy produces a coherent picture of late Holocene relative sea-level change in Poole Harbour.

Phase I - Early Relative Sea-Level Rise.

Between *c.* 4700 Cal. BP and *c.* 2400 Cal. BP, relative sea-level was rising in Poole Harbour, resulting in the inundation and preservation of a sequence of basal peat deposits pre-dating this time. The age-altitude data may indicate that the rate of rise increased after *c.* 3600 Cal. BP.

Phase II - Stable to Falling Relative Sea-Level

At some point *c.* 2400 Cal. BP, the rate of sea-level rise decreased significantly, and PMTL remained at or below -1.0 m OD until at least 1200 Cal. BP. Hubbard & Stebbings (1968) report a similar succession at Keysworth Marsh (western Poole Harbour) where basal peat below *c.* -1.10 m OD is replaced by a sand layer between *c.* -1.10 m OD and -0.80 m OD. It is also instructive to note that the Tertiary bedrock surface indicated by Hubbard & Stebbings resides at *c.* -1.0 m OD, as it does beneath Arne Bay to the north of the study marsh of Arne Peninsula (Gilbertson, 1967). This evidence indicates a phase of erosion between *c.* 2400 Cal. BP and *c.* 1200 Cal. BP during which there was widespread incision into the Tertiary bedrock underlying the Harbour. It is probable therefore that a number of breaks in the sedimentary record exist removing some evidence of depositional conditions during this period. Jarvis (1992) inferred a fall in relative sea-level of *c.* 0.7 m between *c.* 3000 Cal. BP and *c.* 1700 Cal. BP on the basis of archaeological evidence. Whilst the data from Arne Peninsula and Newton Bay provide no conclusive evidence for a relative sea-level fall, both records are entirely consistent with such an event which would explain the inferred breaks in sedimentation.

Phase III - The Lagoonal Period

After *c.* 1200 Cal. BP, the age-altitude data indicate a rise in relative sea-level within the Harbour which appears to have been short-lived, giving way to another period of stability after *c.* 900 Cal. BP. During this time lithostratigraphic and biostratigraphic evidence suggest the formation of a widespread, brackish water lagoon. The rise in relative sea-level would have flooded many marginal areas of the Harbour producing these shallow water, lagoonal conditions. These appear to have started to infill with organic detrital sediments possibly giving rise the emergent phase E1 which may also be related to the *c.* 900 Cal. BP deceleration. It is possible that the former lagoon of Holton Mere, now infilled beneath the *Spartina* sediments of Keysworth Marsh (Hubbard & Stebbings, 1968), was a recent analogue for this type of environment. It must be stressed however that this emergence is largely inferred on the basis of test lining abundance, and the lack of modern analogues for the lagoonal foraminiferal assemblages makes it difficult to unequivocally determine water depth changes.

The early section of the lagoonal deposits at Newton Bay are extensively eroded and replaced by sand layers, deposition of which appears to have ceased *c.* 900 Cal. BP during the end of the phase of rising relative sea-level. The principal source of sand-sized material for modern beaches in Poole Harbour is the erosion of cliffs incised into the Tertiary bedrock of the area (Section 4.1.2). Sand deposition appears to have started during *Phase-II* probably as a result of enhanced erosion, and it is likely that the renewed rise in relative sea-level recorded in *Phase-III* led to deeper water conditions and enhanced wave activity within the Harbour which rapidly mobilised much of the material eroded during the *Phase-II* and deposited it in exposed areas of the Harbour. Sand deposits are less evident in the stratigraphy of Arne Peninsula where the full sequence of lagoonal development is usually present. Arne Peninsula is protected behind Long and Round Islands and is therefore sheltered from wind\wave action that could mobilise this coarse grained material. Additionally, it situated much further away from the entrance to Poole Harbour than Newton Bay in its lower energy, finer grained western reaches.

Changes in the coastal geometry of the Harbour mouth resulting from spit progradation or erosion have been proposed as possible causes for some of the observed depositional changes in Sections 6.1.4 & 6.2.2. These may have arisen as a consequence of changes in relative sea-level but are unlikely to have occurred independently of them. For example, whilst closure or constriction of the Harbour mouth could have created the lagoonal facies this would have been accompanied by a pronounced reduction in marine influence. There is certainly no evidence to suggest that the Harbour was entirely isolated from the sea and the presence of foraminifera indicate a continued marine influence throughout the lagoonal phase. Furthermore, the age-altitude data indicate that relative sea-level was rising during this period and it is difficult to see how this effect could have been produced by a reduction in inlet size.

It is entirely conceivable that the inlet narrowed during the period of lower relative sea-level (*Phase-II*) when the reduced tidal prism would result in diminished current velocities within the Swash Channel. Likewise, during the rise in relative sea-level inferred in *Phase-III* the inlet may have widened under the influence of enhance current velocities which could have caused a switch from ebb to flood dominated conditions, resulting in the injection of large quantities of sand into the Harbour.

After *c.* 900 Cal. BP PMTL appears to have remained largely stable until *c.* 300 Cal. BP. There is very little evidence for sediment accumulation during this period and it is possible that another hiatus is present within the record or that a slight fall in relative sea-level occurred at this time.

Phase IV - Recent Renewed Relative Sea-Level Rise

Between *c.* 300 Cal. BP and 50 Cal. BP an apparent rise in relative sea-level led to the decline of organic lagoonal sedimentation and marsh submergence (S2). This submergence started prior to the *Pinus* rise (200 Cal. BP) within the organic lagoonal sediments and culminated in deposition of the minerogenic silt-clay containing the *Reophax* assemblage which is firmly dated as occurring between the *Pinus* rise and the arrival of *Spartina anglica* (200 Cal. BP - 50 Cal. BP). The mouth of the Harbour is known to have been stable during this period and so cannot be implicated in this change.

The effects of this rise in relative sea-level were quickly moderated by the arrival of *Spartina anglica* and its propensity to enhance sedimentation rates which resulted in marsh emergence (E2). The rise in relative sea-level may have actually assisted the dramatic spread of *S. anglica* since it can tolerate more prolonged periods of tidal inundation than other marsh plants (Section 4.1.3). Ranwell *et al.* (1964) and Hubbard & Stebbings (1968) both note the low altitude of *Spartina* remains in fossil cores from Poole Harbour, and suggest that this could be indicative of a rise in relative sea-level since *Spartina* colonisation. These authors also note that the low altitudes could be explained by compaction of the wet, rapidly accreting *Spartina* sediments. Whilst compaction could have lowered the altitude of the two floral chronohorizons, ARN#1 is a basal sample and NEB#1 overlies a sequence dominated by sand and so is unlikely to have experienced significant compaction.

6.4 BURY FARM

To determine whether the observed changes described in Section 6.3.3 are peculiar to Poole Harbour or of wider significance, the record from Bury Farm, located in the neighbouring system of Southampton Water, is now examined.

6.4.1 Water Level Changes Inferred from BF-96-11

The core BF-96-11 is presented in Figure 4.12 as an example of the typical lithostratigraphy and biostratigraphy of the marsh at Bury Farm. For this reason it is selected as a starting place from which to examine water level changes at this site.

The lithostratigraphy consists of a humified basal peat overlain by 140 cm of grey silt-clay. This minerogenic sediment is interrupted by a thin peat band between 85 cm and 80 cm, before the grey silt-clay sediments return. The upper 35 cm of the sequence is typified by a brown, organic silt-clay with *turfa*.

As in Sections 6.1.1 and 6.2.1, using organic content as a crude indicator of water depth, the lithostratigraphy indicates marsh submergence at 225 cm, followed by a prolonged period of marine sedimentation. This is briefly interrupted by marsh emergence at 85 cm, before a second inundation at 80 cm. The increasing organic content in the upper 35 cm of the sequence suggests a progressive reduction in water depth.

The variations in $SWLI_{(Pred)}$ produced by the transfer function are shown in Figure 6.11. The sample at 84 cm is devoid of foraminifera and this, coupled with its high organic content and the general foraminiferal trend exhibited by adjacent samples, is strongly suggestive of deposition above the upper limit of marine influence. This sample is therefore assigned a SWLI of 228, which equates to the present level at which foraminifera become absent.

The transition from the lowermost freshwater peat into the marine silt-clays obviously reflects a submergence event, termed S1. The presence of calcareous tests implies that the magnitude of S1 is underestimated by the transfer function in a similar way to the events in Poole Harbour. The decline in calcareous tests and *M. fusca*, coupled with an increase in *J. macrescens* above 0.2 m OD indicates an emergence event designated E1. This appears to continue until it culminates in the complete emergence of the saltmarsh between 1.0 m OD and 1.1 m OD.

The remaining record is one of submergence (S2) above the transgressive contact, after which conditions appear to remain stable for sometime until a second submergence event (S3) at around 1.6 m OD. The magnitude of the S3 event is exaggerated due to the transition between deshrinking methods used and it is therefore unclear exactly how significant this change is. The return of *M. fusca* in these samples however, does suggest that there was at least some increase in water depth at this time.

A pollen profile produced from this core is presented in Long & Scaife (1996), and provides some additional information regarding the depositional environments of the various units. The freshwater origin of the basal peat inferred from the foraminiferal data, is consistent with its pollen signature, which possesses high percentage frequencies of *Quercus*, *Alnus*, and *Corylus*, indicating a dry oak-alder woodland with hazel. There is also a rise in *Alnus* between 100 cm and 75 cm, suggesting an expansion of coastal woodland, that corresponds closely to the thin peat band observed between 85 cm and 80 cm. Finally, there is also evidence for a rise in *Pinus* frequencies above 20 cm which may be used as an additional chronohorizon, as at Arne Peninsula.

6.4.2 Vertical Movements of Mean Tide Level

Six radiocarbon dates are available from the Bury Farm marsh and these are presented in Figure 4.11 and Table 4.4. As in Sections 6.1.4 and 6.2.4, the transfer function is used to calibrate the altitude of palaeo-mean tide level for each dated sample. Three dates were screened out during this validation process (Appendix Three), and the remaining three age estimates in association with the *Pinus* rise index point are displayed in Figure 6.12.

On the basis of this evaluation it is suggested that PMTL at Bury Farm was rising until at least *c.* 3100 Cal. BP. After this time there is little evidence for significant movements in mean tide level, which appears to have resided between *c.* -1.2 m OD and -0.5 m OD until at least 1500 Cal. BP. After this point, there was a second, renewed rise in relative sea-level to reach the conditions observed today.

6.4.3 Synthesis of Data from Bury Farm

From the combined lithostratigraphic, biostratigraphic and chronostratigraphic data described above, it is evident that the early period of the record at Bury Farm is characterised by inundation of the site (S1) during a phase of rising relative sea-level. The foraminiferal data, coupled with the erosive nature of many of the lower peat contacts at this site, suggest that the S1 event may have been quite abrupt.

The emergent episode E1 appears to have occurred during the period after PMTL had stabilised *c.* 3000 Cal. BP. The culmination of this emergent phase is the peat bed between 85 cm and 80 cm which, from BF#2 and BF#3, formed between *c.* 3000 Cal. BP and 1500 Cal. BP. The fact that this peat bed persisted for up to 1500 years but is very thin suggests that sedimentation rates were extremely low or that some of the peat has been removed by erosion. The expansion of coastal woodland at this time, coupled with the well developed nature of the deposit means that it may indicate a fall in relative sea-level between 3000 Cal. BP and 1500 Cal. BP which gave rise to such low sedimentation rates.

The terrestrial conditions associated with peat formation were terminated by the arrival of marine conditions after 1500 Cal. BP. After the initial inundation there is little change in water depth, suggesting that any acceleration in the rate of relative sea-level rise was short-lived and quickly succeeded by a deceleration.

The S3 event is dated to *c.* 200 Cal. BP on the basis of its proximity to the *Pinus* rise although its magnitude may be overestimated owing to the transition between deshrinking methods used to produce the $SWLI_{(Pred)}$ values.

6.4.4 Comparison with the Record from Poole Harbour

Figure 6.13 shows the sea-level index points from Bury Farm plotted with those from Poole Harbour, having all been normalised to account for the altitudinal differences between mean tide level at each site.

BF#3 and BF#4 fill a gap on the rising limb of the mean tide level curve prior to 3000 Cal. BP and are consistent with index points from Arne Peninsula and Newton Bay providing further support for the *Phase-1* rise in relative sea-level. These dates seem to support the suggestion of an acceleration in the rate of relative sea-level rise between 3500 Cal. BP and

2700 Cal. BP, although this is equivocal owing to the errors associated with age-altitude analysis. Such an acceleration cannot be determined by the existing biostratigraphy since the S1 event is associated with the inundation of basal peat deposits (see Section 2.3.3). The fact that S1 appears to have been characterised by a significant deepening is consistent with, if not proof of, an acceleration at this time.

The new data from Bury farm places the inflexion of the mean tide level curve at *c.* 2800 to 3000 Cal. BP, slightly earlier but still consistent with the data from Poole Harbour. This supports the suggestion that *Phase-II* was characterised by a reduction in the rate of relative sea-level rise or a relative sea-level fall. The fact that a peat layer associated with an expansion in coastal woodland developed at Bury Farm during the course of the E1 emergence, gives added support to the notion of a relative sea-level fall at this time.

There is no evidence of a lagoonal facies similar to the *Phase-III* deposit of Poole Harbour at Bury Farm but the pattern of relative sea-level change indicated is comparable. The data from Poole Harbour suggest that *Phase-III* was accompanied by a rapid, short-lived rise in relative sea-level that caused flooding and lagoon inception. The inundation (S2) of the peat bed at Bury Farm dated to *c.* 1500 Cal. BP appears comparable in terms of age and altitude to the S2 event *c.* 1200 Cal. BP identified at Poole Harbour. Furthermore, the change at Bury Farm also appears to be characterised by a brief acceleration in the rate of relative sea-level rise followed by a return to a period of stability. This strongly suggests that the changes observed in Poole Harbour are principally caused by variations in relative sea-level and are not significantly distorted by modifications of the Harbour entrance.

Phase-IV consists of a submergence event associated with the *Reophax* assemblage in Poole Harbour that has been dated *via* the *Pinus* rise and increase in *Spartina* to between 250 Cal. BP and 50 Cal. BP. The S3 event at Bury Farm also occurs around the *Pinus* rise, and it is therefore possible that the two events are indicative of a near synchronous acceleration in the rate of relative sea-level rise during the last 250 years. It should be noted that there is a vertical offset between the inferred PMTL of the two *Pinus* rise deposits of *c.* 20 cm to 70 cm. This may indicate inaccuracies in the tide data used to compile the $SWLI_{(Pred)}$ at each site, or errors in the assignation of the $SWLI_{(Pred)}$ values. Alternatively it may reflect greater sediment compaction at Arne Peninsula than at Bury Farm.

In summary, the data from Bury Farm provides support for all four phases of relative sea-level change inferred from Poole Harbour and serves to more tightly constrain the timing of these periods. Together these data provide compelling evidence that late Holocene sea-level change in this section of the Solent region was not characterised by a continuous, slow paced rise to its current level, but that up to three accelerations in the rate of relative sea-level rise may have occurred during the last 3500 years.

6.4.5 Comparison with Existing Sea-Level Data from the Solent Region

Existing sea-level data from the Solent region is sparse and has been comprehensively reviewed by Long & Scaife (1996). Only five studies have employed modern sea-level reconstruction techniques combining lithostratigraphic, biostratigraphic, and chronostratigraphic data to produce reliable sea-level index points: Hodson & West (1972) at Fawley in Southampton Water; Long & Tooley (1995) at Stansore Point where Southampton Water joins the Solent; Long & Scaife (1996) at various locations within the body of Southampton Water itself; Long *et al.* (in press) in Poole Harbour; and Long & Scaife (in press) along the northern coast of the Isle of Wight. From these studies, only four radiocarbon dates fall within the last 3500 Cal. years and the resulting sea-level index points along with the other, older dates, are presented in Figure 6.14 with data collected in this thesis.

Where an age overlap exists, the pre-existing sea-level index points show good agreement with the data collected in this thesis. Of particular note is the sea-level index point at *c.* 1300 Cal. BP that comes from a transgressive contact at Hythe, Southampton Water (Long & Scaife, 1996). This shows excellent agreement with BF#2 and ARN#3, adding further support to the proposed renewal of relative sea-level rise prior to *c.* 1500 Cal. BP to 1300 Cal. BP within Solent region. In additions a radiocarbon date collected from the lower peat at Bury Farm (Long & Scaife, 1996) plots in excellent agreement with ARN#5 which comes from a comparable basal peat in neighbouring Poole Harbour.

The general form of the age-altitude data in Figure 6.14 indicate that the rate of relative sea-level rise in the Solent has progressively declined until *c.* 3000 Cal. BP when a period of minimal change is apparent. This view is in good accord with the pattern of events inferred from the data collected in this thesis.

There are four intervals in the age-altitude data from the Solent region where sea-level index points are particularly sparse or entirely absent. These occur between *c.* 6000 Cal. BP and 5200 Cal. BP, 4200 Cal. BP and 3500 Cal. BP, 3000 Cal. BP to 1600 Cal. BP, and 1100 Cal. BP to 500 Cal. BP. Whilst the altitudinal uncertainties of the data inhibits detailed interpretation of this distribution, it is tempting to speculate that these periods equate to phases of reduced relative sea-level rise during which time erosion and reworking of deposits has destroyed part of the sea-level record.

Two periods of possible stable to falling relative sea-level have been inferred from the records in Poole Harbour and at Bury Farm. The first was suggested to have occurred during *Phase-II* of the relative sea-level history, equating to the period between *c.* 2400 Cal. BP and *c.* 1500 Cal. BP to 1200 Cal. BP. Whilst no precise sea-level data exists for this period, there is some archaeological evidence that supports the idea of lowered relative sea-level during the late Holocene period. Jarvis (1992) suggests that PMTL lay between -3.2 and -2.8 m OD around 1800 Cal. BP to 1700 Cal. BP in Poole Harbour. At a similar time, evidence from a Roman salt-kiln on the Isle of Wight suggests that PMTL lay around -2.3 m OD (Long & Scaife, in press). These clearly place relative sea-level well below the PMTL of -1.5 m OD to -0.5 m OD produced by a linear interpolation between the sea-level index points of this time. The low altitudinal precision of such archaeological evidence warns against their quantitative application to studies of relative sea-level change (Long & Scaife, in press). Nevertheless, these data provide support to the qualitative inference of a fall in relative sea-level during *Phase-II* that was recognised in a number of localities within the Solent. It is interesting to note that the early work of Godwin referred to in Section 2.5.1.1 suggested a fall in relative sea-level after *c.* 2000 Cal. BP that was replaced by a renewed sea-level rise *c.* 1200 Cal. BP.

The second period of possible relative sea-level fall is tentatively inferred after *c.* 900 Cal. BP but having ceased prior to *c.* 300 Cal. BP. Archaeological support for this is more equivocal but some data are available from the Wootton-Quarr study on the Isle of Wight (Tomalin *et al.*, in press). A late Norman occupation site dated to *c.* 750 Cal. BP suggests a PMTL below that indicated by NEB#2 (*c.* 900 Cal. BP) from Poole Harbour. Long & Scaife (in press) speculate that one interpretation of these archaeological data is that there have been two significant oscillations in relative sea-level since the Roman period. The data collected from Poole Harbour and Southampton Water provide some support for this conjecture.

6.5 LLANRHIDIAN MARSH

The discussion so far has been restricted to marshes from neighbouring systems on the southern coast of Britain. Both these areas have many features in common, such as similar rates of crustal subsidence, and the fact that they both are southerly facing and influenced by conditions in the English Channel. Llanrhidian Marsh is in stark contrast to both these systems, in that it is westerly facing, and affected by processes operating in Carmarthan Bay and the Bristol Channel. Additionally, it has a much greater tidal range and the saltmarshes are correspondingly larger in extent. Investigation of the record of relative sea-level change recorded at this site permits the spatial scale of the changes observed on the south coast to be investigated. It will also provide information on the suitability of these contrasting environments for the generation of high resolution records of sea-level change.

6.5.1 Changes in Depositional Environment

A detailed investigation of the lithostratigraphy at Llanrhidian is presented in Appendix One and summarised in Figure 4.16. The abundance of organic material and intercalated minerogenic deposits offers the potential for establishing a number of well dated sea-level index points. Analysis of these sediments however, revealed that they are generally devoid of foraminifera, and frequently possess no microfossils in the >63μm size range. Where present, foraminifera species are restricted to those taxa characteristic of the highest marsh environments (Section 4.3.4.2). The low abundance nature of the deposits at Llanrhidian, coupled with the poor vertical zonation exhibited by the modern marsh foraminiferal assemblages (Section 5.5), means that it is not possible to perform a combined lithostratigraphic and biostratigraphic investigation of changing marsh depositional environments as has been achieved for the sites on the south coast. Instead, a very basic interpretation of depositional environment must be made, relying heavily on the lithostratigraphy supplemented where possible by the sparse biostratigraphic data.

The lithostratigraphy indicates that below 4.1m OD, the sequence at Llanrhidian is dominated by well developed peat deposits. These typically rest upon a grey, frequently iron-stained, consolidated clay with large sand clasts, found between 2 m OD and 4 m OD. The peat deposits are largely unbroken in the eastern transects, but minerogenic intercalations increase toward the west. The two transects from Stavel Hagar Mill contain two principal episodes of minerogenic sedimentation. The first occurs between 2.2 m OD and 3.2 m OD, whilst the second, more widespread deposit resides between 3.5 m OD and 3.8 m OD. Above 3.8 m OD, there is commonly a thin, well developed peat which terminates, often erosively, at around 4.1m OD. The remaining sediment is a grey silt-clay, which is often iron-stained.

The lower peat bed contains abundant detrital wood fragments, and these are often found in the lower, intercalated brown-grey organic-rich silt clays in the landward cores. The biostratigraphy provides no evidence to suggest that these silt-clays are marine in origin and, from the abundance of detrital organic material contained within them and the adjacent peat units, it appears that the environment was predominantly freshwater, perhaps with periods of standing water. In the most seaward core at Stavel Hagar (STH2-97-80), the lower minerogenic intercalation possesses lumps of detrital peat within it, and the presence of low numbers of foraminifera suggests a marginal marine influence at this locality. It is possible therefore, that whilst the more landward minerogenic sequences are not marine themselves, they may be related to an elevated water table induced by an increase in marine influence. It seems significant that the minerogenic intercalations are more pronounced in the vicinity of Stavel Hagar Mill, where there is an abundant supply of freshwater draining from the hinterland (Section 4.3.4). These springs could not only supply a moderating freshwater influence at this location, but also the mechanism by which terrestrial sediments are introduced onto the marsh.

At Stavel Hagar Mill, the upper peat is very similar in composition to the lower peat. In the eastern transects there is no minerogenic intercalation and the peat bed extends up to between 4.0 m OD and 4.2 m OD, where it is replaced by a heavily iron-stained, crumbly clay-silt with a trace of *turfa*, very similar in composition to the sediments found in the modern day highest marsh areas. At Stavel Hagar Mill, a second minerogenic intercalation interrupts the peat accumulation at around 3.6m OD. This silt-clay, which is frequently rich in organic matter and contains *Phragmites* remains, is also largely devoid of foraminifera, although does contain some very low abundance, high marsh assemblages. This unit, whilst very variable in composition, becomes increasingly organic to seaward which once again suggests that its minerogenic component is not of marine origin.

A thin but laterally extensive humified peat layer is found at Stavel Hagar Mill, centred around 4 m OD. This frequently possesses an eroded upper contact and appears to be related to the termination of the main peat bed observed in the more eastern transects. Indeed, the overlying iron-stained silt-clay is comparable to that observed above the main peat in these areas.

An 'inverse stratigraphy' is apparent in STH2-96-10 centred around 80 cm. Here, a thin organic layer intrudes between the less organic silt-clay of the surrounding unit. The foraminiferal counts shown in Figure 6.15 indicate that they reach a peak within this organic unit, and decline on either sides in association with the silt-clay. In this way, a traditional lithostratigraphic regressive contact is actually a biostratigraphic transgressive contact.

6.5.2 Vertical Movements of Mean Tide Level

The five AMS radiocarbon dates taken from the marshes at Llanrhidian are presented in Figure 4.16 and Table 4.5. The transfer function cannot be used to calibrate the altitude of palaeo-mean tide level for samples from Llanrhidian Marsh, and so the intention was to assign $SWLI_{(Pred)}$ visually as outlined in Section 5.8.3. The fossil foraminiferal assemblages recovered however, were dominated by *J. macrescens* almost to the exclusion of all other taxa. This rendered it impossible to assign a $SWLI_{(Pred)}$ on the basis of foraminiferal assemblage composition. Instead, foraminifera are used to confirm the presence of a marine influence in the sample material which is assigned a classical indicative meaning as described by Shennan (1986). The five sea-level index points are evaluated and screened in Appendix Three, and on this basis one index point is rejected. The remaining four sea-level index points are presented in Figure 6.16.

The principal error associated with the sea-level index points at Llanrhidian is their altitude. The index points are subject to vertical uncertainties since precise $SWLI_{(Pred)}$ values could not be assigned visually. Additionally, there is the possibility that the organic-rich sequences may have undergone considerable compaction. The sea-level index points indicate a progressive rise in relative sea-level until *c.* 1900 Cal. BP. At some point after this relative sea-level rise must have increased, apparently elevating PMTL above its present altitude by *c.* 600 to 700 Cal. BP. No age-altitude data are available for the last 700 years of change at Llanrhidian Marsh, but PMTL must have remained relatively stable to falling to result in the conditions observed at the marsh today.

6.5.3 Synthesis of Data from Llanrhidian Marsh

From an evaluation of the various lines of evidence from Llanrhidian Marsh, the following sequence of events is suggested. Between *c.* 4200 Cal. BP and *c.* 2500 Cal. BP, relative sea-level was rising, inundating the seaward cores at Stavel Hagar Mill. In the back marsh environment at this time, the elevated water table coupled with the input of freshwater by springs in the marsh hinterland, led to a period of localised brackish to fresh standing water in this area.

After *c.* 2500 Cal. BP, the growth of a well developed peat in the seaward cores suggests a removal of marine conditions at this time, and the return of organic rich sedimentation across the marsh may also indicate a relative fall in water table. Whilst there is no evidence to support a reduction in the rate of relative sea-level rise at this time, the age-altitude data are too sparsely distributed to preclude this suggestion.

After *c.* 1900 Cal. BP there is evidence of another increase in marine influence in the seaward cores at Stavel Hagar, possibly relating to the proposed acceleration in the rate of relative sea-level rise after this time. Once again, the back marsh environment appears to have responded to a rising water table by a freshening of the environment, and the lithostratigraphy suggests an extensive *Phragmites* reed bed developed during this period.

There is a progressive increase in organic content of the seaward cores and the limited biostratigraphic evidence suggests that the thin peat bed centred around 4.2 m OD and dated to *c.* 600 to 700 Cal. BP, is of freshwater origin. This apparent removal of marine conditions correlates well with the inferred stillstand or slight fall in relative sea-level which must have occurred to bring mean tide level to its current altitude. In many places, this upper peat contact is eroded, which is consistent with the focusing of erosive forces in a restricted vertical range.

6.5.4 Comparison with Existing Sea-Level Data from the Bristol Channel

As noted is Section 4.3.3, there are no reliable sea-level index points from the area around the Gower Peninsula during the last 3000 years and consequently there are no data against which to evaluate the record compiled during this study. Figure 6.17 shows the sea-level index points from the Loughor Estuary plotted with available data from the Bristol Channel area. The data appear to be broadly consistent with the general pattern of change inferred from the Bristol channel, but there are insufficient data to infer changes of comparable resolution to those derived from the Solent Region.

6.6 COMPARISON WITH THE RECORD FROM THE SOUTH COAST

The record from Llanrhidian Marsh is obviously of much lower resolution than those derived from the marshes on the south coast. A comparison between the two regions is still useful however, and the more detailed information from the Solent region could provide additional information on relative sea-level change in the Gower.

Before the relative sea-level records from these two regions can be compared it is necessary to take account of the different crustal movements experienced by them during the late Holocene period. The most recent survey of crustal residuals in the UK is the work of Shennan (1989). The study areas in this thesis however, are situated in regions where crustal movements are poorly resolved due to a lack of available sea-level data. Using the same methods employed by Shennan (1989), the sea-level index points produced during the course of this study are used in association with the existing sea-level data catalogued in Appendix Three, to calculate crustal residuals for both study regions. This is achieved by subtracting the smoothed, regional eustatic curve of Mörner (1984) from the relative sea-level data. The results are displayed in Figures 6.18 and 6.19. The results indicate that the Solent region has experienced linear subsidence of around 0.5 mm a^{-1} during the late Holocene period whilst the Loughor Estuary has subsided at around 0.4 mm a^{-1}. These values are consistent with the pattern of crustal residuals produced by Shennan (1989), but it should be noted that only four index points are available for the Loughor Estuary estimate.

On the basis of these crustal data it appears that both regions have experienced similar regimes of subsidence during the study period. Figure 6.20 shows the screened sea-level index points from all sites reduced to a common mean tide level with the data from the Loughor Estuary adjusted for the 0.1 mm a^{-1} difference in subsidence. It should be noted that the vertical error terms have been increased by ± 0.15 m to take account of the altitudinal errors associated with the Ordnance Survey benchmark network as discussed in Section 3.3.

The sequence of events from the south coast commences with a period of marsh submergence under rising relative sea-level, continuing until *c.* 3000 Cal. BP after which there was a slowing of sea-level rise and evidence for emergence until between *c.* 1500 Cal. BP. This is broadly consistent with the evidence from the Gower which shows a rise in relative sea-level between *c.* 4200 Cal. BP and *c.* 2600 Cal. BP, after which there was an inferred reduction in the rate of relative sea-level rise until *c.* 1900 Cal. BP. Whilst the sequence of events is the same, the timings appear to differ by around 400 years. This discrepancy is accountable for in terms of the age errors which are typically ± 200 years for deposits of this age, but there is insufficient evidence to confirm that these events are synchronous.

The south coast record shows evidence of a submergence after *c.* 1500 Cal. BP to 1200 Cal. BP corresponding to a possible short lived acceleration in the rate of relative sea-level rise until at least *c.* 900 Cal. BP, after which there was evidence for a general still stand or low rate of relative sea-level rise. This sequence is also consistent with the record from the Gower which shows evidence for a submergence after *c.* 1900 Cal. BP that was followed by an emergence event culminating around 600 Cal. BP. Once again, it must be stressed that without more precise age-altitude control, this evidence is fairly circumstantial.

The uppermost sediments at Llanrhidian Marsh provide no indication of changing water depths after *c.* 600 Cal. BP, and there certainly is no evidence for an acceleration in the rate of relative sea-level rise within the last 200 centuries. This may simply reflect the lower sensitivity or resolution of the record at Gower, or may indicate different processes operating between the two systems. There is a possibility however, that the inferred acceleration in the rate of relative sea-level rise recorded in Poole Harbour and Southampton Water is erroneous, and caused by a combination of underestimates of water depth during the lagoonal phase, and the compaction of *Spartina* sediments causing an apparent lowering of mean tide level.

6.7 SUMMARY

- A combination of lithostratigraphic, biostratigraphic, and chronostratigraphic data from three sites in the Solent Region indicates five major phases of relative sea-level change during the late Holocene.

- *Phase-I*: A rise in relative sea-level between *c.* 4800 Cal. BP and *c.* 2800 Cal. BP resulting in the inundation and preservation of basal peat deposits;

- *Phase-II*: A deceleration in the rate of relative sea-level rise probably resulting in a still-stand or fall in relative sea-level between *c.* 2800 Cal. BP and *c.* 1500 Cal. BP. During this time emergence appears to have occurred although the period is characterised by extensive erosion of coastal deposits removing significant portions of the record;

- *Phase-III*: A renewed rise in relative sea-level commencing prior to *c.* 1500 Cal. BP to 1200 Cal. BP and terminating around 900 Cal. BP;

- *Phase-IV*: A period of comparatively stable relative sea-level persisting until *c.* 300 Cal. BP;

- *Phase-V*: A recent acceleration in the rate of relative sea-level rise during the last 300 years.

- The record from the Loughor Estuary is of much lower resolution that that available for the Solent region but exhibits variations consistent with those expected to arise from the five phases of relative sea-level change outlined above.

- A comparison of crustal residuals from the two areas implies that they have both experienced long-term subsidence during the late Holocene amounting to between 0.4 mm a^{-1} and 0.5 mm a^{-1}.

Late Holocene Sea-Level Change and Climate in Southern Britain

Evidence for climate change during the late Holocene period has been presented in Chapter Two where it was used to produce a qualitative prediction of the expected response of sea level (Section 2.7 and Figure 2.36). In Chapter Six the evidence for late Holocene relative sea-level change in southern Britain derived from four study sites was presented. In this chapter these two records are compared to test the hypothesis that there is a discernible climate signal in the relative sea-level data from southern Britain. This chapter comprises the following:

- A comparison of the timing of relative sea-level change recorded in southern Britain with the sea-level variations proposed by the climate-based prediction;
- An assessment of the magnitude of relative sea-level change involved and the degree to which this could be explained by climate-related processes;
- An examination of evidence for a North Atlantic teleconnection between the climate and sea-level records from the west and east coasts.

7.1 EVIDENCE FOR A CLIMATE SIGNAL IN THE SEA-LEVEL RECORD FROM SOUTHERN BRITAIN

7.1.1 Record Synchroneity and Correlation

Synchroneity is an assumption inherent within the climate-based prediction of late Holocene relative sea-level change presented in Section 2.7. It is founded upon the idea that sea level will respond rapidly, and in a predictable manner, to changes in climate and any associated alterations in oceanic circulation.

The climate record presented in Chapter Two indicates that the timing, direction and magnitude of climate fluctuations vary with location. In addition, many climate proxies such as the high resolution tree ring data compiled by Briffa *et al.* (1995), show rapid oscillations at timescales of years to decades. The simplified record of climate change presented in Figure 2.15, which was used as the basis of the relative sea-level prediction in Section 2.7, attempts to overcome this irregularity by combining records from a number of different sources and locations, and by focusing on generalised changes that persist for several decades or longer. Nevertheless, portrayal of climate as a single variable affecting sea level is an oversimplification, and consequently the age and duration of an individual climate change event will actually be an average value derived from a number of different sources. This is similar to the 'suck-in' effect that arises when relating a wide range of age-estimates to a single dated phenomena (Baillie, 1991). Consequently, it is important to remember that the apparently abrupt transitions in Figure 2.15 may have associated age uncertainties of several decades to centuries.

The timing of a 'eustatic' signal's recognition in the relative sea-level records from southern Britain will be influenced by crustal movements. Information regarding crustal movements in this region is limited, but from Section 6.6 it is evident that both the Solent and Gower Peninsula have experienced subsidence during the late Holocene period. This means that accelerations in the rate of sea-level rise will be recognised more rapidly in their relative sea-level records than decelerations, and relative sea-level still-stands or falls will have greater significance. Furthermore, differences in the rates of crustal subsidence between the two study areas may result in the same eustatic event being identified at different times in the relative sea-level records. Intuitively however, these differences in timing will be minimal, assuming that the differential subsidence rate of 0.1 mm a^{-1} indicated in Section 6.6 is reliable.

The factors outlined above mean that synchronous changes in climate and sea level may appear offset due to limitations in defining the age of an event and lags in its recognition in the relative sea-level record. Furthermore, there may be time lags involved between a change in climate and its effect on sea level, resulting in increased diachroneity. These limitations mean that synchroneity is not the definitive measure of a climate-sea level relationship. However, in the absence of more a reliable alternative, such visual comparison is a common method of investigation (e.g. van de Plassche, 1991; Fletcher *et al.*, 1993).

7.1.2 Comparing the Records

Figure 7.1 shows the climate-based sea-level prediction (Figure 2.36) plotted against the record of relative sea-level change from the Solent region and the lower resolution record from the Loughor Estuary. It is immediately apparent that these data exhibit a large degree of synchroneity within the limitations of error described above.

The climate-based scenario of relative sea-level change proposed in Section 2.7 consists of five phases that will be considered in turn.

7.1.2.1 RSLC-I: A reduction in the rate of relative sea-level rise **c.** *3000 Cal. BP.*

The data from Poole Harbour describe the transition from a period of rapidly rising relative sea-level (*Phase I*) to a period of relative sea-level stability (*Phase-II*) at between *c.* 2800 Cal. BP and 3000 Cal. BP. This is in excellent agreement with the predicted deceleration in relative sea-level rise (*RSLC-I*) caused by an extended period of cooling identified in ice core records from Greenland as starting between 3500 Cal. BP and 3000 Cal. BP (Dansgaard *et al.*, 1984). The recognition of a still-stand or fall in relative sea-level on a subsiding coast indicates a fall in **eustatic** sea level. The expansion of glaciers around this time would have resulted in a glacio-eustatic lowering of sea level and therefore provides a possible mechanism by which this change may have been affected. In the Swiss Alps, Patzelt (1974) suggests a phase of glacier advance from *c.* 3000 Cal. BP to 2300 Cal. BP (Göschener I advance) and proposes that the glaciers remained in their advanced positions until *c.* 1900 Cal. BP. This correlates closely with the prolonged still-stand or fall in relative sea-level identified in *Phase-II* of the Solent record. This is also consistent with the record from Greenland glaciers which indicate advance commenced between *c.* 3500 Cal. BP to 3000 Cal. BP and climaxed around *c.* 2000 Cal. BP (Kelly, 1980; Weidick *et al.*, 1990).

The post-3000 Cal. BP climate deterioration is also associated with an increase in wetness across parts of northwest Europe and is broadly correlated with the Sub-Boreal/Sub-Atlantic transition. This shift to cool and wet conditions is analogous to the situation encountered when the NAO-index is high, deepwater formation shifts to the Labrador Sea, and the Sargasso Sea becomes relatively warm and quiescent (Sections 2.2.2 & 2.2.3;

Figure 2.16). The Sargasso Sea record of Keigwin (1996) which shows a warming phase around 3000 Cal. BP supports this interpretation. This implied reduction in the strength of the Gulf Stream would serve to reduce the volume of water transported northward and also lead to a reduction in SST of the waters around the British Isles (Section 2.2.1; van Geel *et al.*, 1996). This suggests that steric effects, possibly augmented by dynamic changes in ocean currents, may also have contributed to the eustatic fall in sea level inferred at this time.

The record from the Loughor Estuary is less precise than the Solent record, and the timing of observed changes is more strongly influenced by the distribution of a small number of sea-level index points. Nevertheless, the pattern of change is broadly comparable and a deceleration in the rate of relative sea-level rise is inferred around 2500 Cal. BP.

7.1.2.2 RSLC-II: An increase in the rate of relative sea-level rise after c. *2000 Cal. BP.*

The Solent record indicates an acceleration in the rate of relative sea-level rise *c.* 1500 Cal. BP to *c.* 1200 Cal. BP that resulted in the inundation of well-developed peat deposits in Southampton Water, and the inception of lagoonal sedimentation in Poole Harbour. It is quite possible that the rise in relative sea-level commenced prior to this and took some time before it was recorded in the sediments. The sequence from Bury Farm indicates that the peat inundated *c.* 1500 Cal. BP was freshwater in origin and could therefore display an initial lagged response to a rise in relative sea-level. Furthermore, archaeological data from Poole Harbour suggests that PMTL began to rise from its *Phase-II* low after *c.* 1700 Cal. BP (Section 4.1.3).

A variety of records provide evidence for a warming in climate at this time (Section 2.4). The ice core record of Dansgaard *et al.* (1984) suggests that milder conditions commenced around 2000 Cal. BP and the accumulation data of Meese *et al.* (1994) indicate that warmth was more prevalent after *c.* 1300 Cal. BP. There is supporting evidence from Greenland where glaciers interrupted their advance between *c.* 1500 Cal. BP and 1000 Cal. BP. Tree ring data seems to suggest climate change of a compensatory nature, with continental and southern Europe showing a cooling from 1100 to 800 Cal. BP whilst the Fennoscandian record shows a warming phase at the same time. Mörner (1995) suggests that this type of signal results from a change in Gulf Stream circulation when an intensification of the northerly flowing branch (the North Atlantic Drift) causes warmer water to be introduced to the high latitudes, and reduces the supply of equatorial water to southern Europe. A change in ocean circulation at this time is consistent with marine records which indicate warm and stable conditions in the Nansen Fjord between *c.* 1200 Cal, BP and 800 Cal. BP (Jennings & Weiner, 1996) associated with a strong Irminger Current (and therefore northerly Gulf Stream branch). At the same time, Keigwin (1996) records a cooling phase from *c.* 2000 Cal. BP to *c.* 1500 Cal. BP in the Sargasso Sea.

The record from the Gower is also consistent with a proposed acceleration after *c.* 2000 Cal. BP. An acceleration in the rate of relative sea-level rise is inferred around 1900 Cal. BP on the strength of an apparent increase in marine influence in the seaward cores and a raised water table creating a *Phragmites* reed bed at the back of the marsh.

7.1.2.3 RSLC-III: A reduction in the rate of relative sea-level rise around 1000 Cal. BP to 800 Cal. BP.

The record from Poole Harbour shows a transition at *c.* 900 Cal. BP from a regime of rising relative sea-level that characterised the initial part of *Phase-III*, to a period of much more slowly rising, possibly stationary, relative sea-level. This phase of minimal vertical movements in relative sea-level persisted until *c.* 300 Cal. BP. The record from Southampton Water provides little new evidence about this period, but the lack of change in water depth after the initial submergence of the peat deposit around 1500 Cal. BP is consistent with a phase of reduced relative sea-level change, especially when set against the trend of long-term subsidence identified in Section 6.6.

Ice core records show an abrupt cooling of climate around 800 Cal. BP (Section 2.4.1.1) and there is widespread evidence for glacial advances at this time (Section 2.4.1.2). This reduction in temperature is less pronounced in tree-ring records which seem to suggest temperatures oscillated around those of the present day (Section 2.4.1.3). The sensitive ocean record from Nansen Fjord indicates that polar water expanded southward around 800 Cal. BP and may have contributed to the cooling recorded in the high latitudes. This was accompanied by increased reports of sea ice around Iceland which may ultimately have led to the demise of Norse colonies (Lamb, 1977).

The record from the Loughor Estuary is particularly sparse during this period with no sea-level index points over a 1000 year period. The only available date *c.* 600 Cal. BP indicates that after this point, relative sea-level rise must have been minimal or slightly negative to produce the conditions observed within the estuary today.

7.1.2.4 RSLC-IV: An increase in the rate of relative sea-level rise around 500 to 300 Cal.BP.

The acceleration proposed around 500 Cal. BP to 300 Cal. BP is based upon a combination of data from terrestrial and oceanic sources. Whilst the last 600 to 800 years have been typified by cold conditions, evidence for a warmer phase around *c.* 400 Cal. BP to 300 Cal. BP is apparent in ice core, tree ring, and peat bog data. Furthermore, the Sargasso Sea record of Keigwin (1996) shows evidence for a brief reduction in Gulf Stream strength *c.* 600 Cal. BP to 500 Cal. BP.

The Solent data show no evidence for a change in the rate of relative sea-level rise directly correlating with this period. The acceleration of *Phase-V* however, immediately follows it and may therefore be a lagged response to these changes. An alternative explanation is that the perturbation in climate at this time was too brief to produce a recognisable signature in the relative sea-level record from southern Britain.

No information is available from the Loughor Estuary record for the last 600 years.

7.1.2.5 RSLC-V: A possible acceleration in the rate of relative sea-level rise during the last 150 years.

This final phase of predicted relative sea-level rise is the most tentative of all the proposed periods of climate-induced change. Whilst the temperature records clearly indicate a warming during this period, no discernible signal is present in the tide gauge data (Section 2.3.2). The Solent record does show an acceleration in the most recent period of the record, but the biostratigraphic data suggests that this was already under way before the *Pinus* rise dated to *c.* 200 Cal. BP. The arrival of *Reophax* species and the switch to minerogenic sedimentation occurs during this period. The lack of modern analogues however, means that it is impossible to determine whether this change is a continuation of the submergence that began prior to 200 BP or evidence of a more rapid deepening, perhaps indicative of a further acceleration in the rate of relative sea-level rise.

7.1.3 Magnitude of Change

In Section 2.5.1.1 the importance of being able to account for the magnitude of sea-level change was stressed by citing the example of Ters (1987) who proposed an 8.5 m rise in relative sea-level during a period when climate was cooling markedly. In order for the relative sea-level variations described above to be linked with climate, it is necessary to ensure that fluctuations in temperature are of great enough magnitude and duration to account for the addition/removal or expansion/contraction of sufficient water to realise the change. This question of magnitude is rarely considered because the relationship between climate and sea level is poorly quantified. For example, despite the known rise in temperature during the last century, limitations inherent in the measurement of secular sea level mean that it is difficult to accurately assign an associated rate of eustatic sea level rise (Section 2.3.2). Considerable effort has been directed to the prediction of future sea level rise based upon climate models, and at present these data provide the best indication of magnitudes of sea-level change resulting from climate variations (Houghton *et al.*, 1990, 1992, 1996).

The IPCC report embodies the most recent consensus on climate change and associated sea level rise (Houghton *et al.*, 1996). Figure 7.2 shows the predicted rise in sea level until 2100 AD calculated for a variety of forcing scenarios. Whilst the precise values are clearly open to debate, the expected magnitude of change remains less than one metre for temperature variations of between 1 °C and 5 °C. It is apparent that the changes inferred from the relative sea-level records of southern Britain are of comparable magnitude to the climate-based rise in sea level predicted for the next century. On this basis, the climate variations documented during the late Holocene period are of sufficient magnitude to account for the relative sea-level changes proposed in this study.

7.1.4 Conclusion

The relative sea-level records from southern Britain clearly display a strong correlation with the phases of sea-level change predicted on the basis of documented variations in late Holocene climate. Whilst such correlation does not prove causation, the record of relative sea-level change from southern Britain is broadly consistent in timing and magnitude with that expected to arise as a consequence of climate change. From these data there is no justification for the rejection of the hypothesis proposed in Section 1.2.1 which stated that 'late Holocene relative sea-level change in southern Britain reflects documented climate changes during this period'.

Whilst limitations in determining the timing of changes in climate or sea-level preclude precise determination of lead and lag times (Section 7.1.1) some general remarks can be made regarding data distribution. The two phases of stable or falling relative sea-level inferred from the Solent region correlate well with the predicted decelerations of *RSLC-I* and *RSLC-III*. The two proposed accelerations in relative sea-level rise however, appear to lag behind those predicted in *RSLC-II* and *RSLC-IV*. The lack of tide gauge evidence for an acceleration in the rate of relative sea-level rise during the last 150 years, despite a well quantified warming, is consistent with the existence of such a lag.

Changes in the rate of relative sea-level rise in southern Britain appear to be closely associated with the Sargasso Sea record produced by Keigwin (1996) and may therefore indicate variations in Gulf Stream strength are invlolved (Section 2.4.4). There is some evidence from Figure 7.1 for the existence of a *c.* 300 to 400 year lag between shifts in ocean circulation inferred from this record and changes in the rate of relative sea-level rise recorded in southern Britain. It must be stressed that the temporal limitations in both records outlined above render such relationships equivocal.

7.2 EVIDENCE FOR NORTH ATLANTIC TELECONNECTIONS

In Section 2.3.4, analysis of the North American sea-level data revealed three periods when similar changes in the rate of relative sea-level rise were observed at a number of marshes: a general deceleration in the rate of relative sea-level rise between 4000 and 3000 Cal. BP; an acceleration around 1800 Cal. BP; and an acceleration or series of accelerations during the last 1000 years, most notably around 800 Cal. and also possibly between *c.* 400 and 300 Cal. BP. Once again, the Sargasso Sea record of Keigwin (1996) correlates extremely well with the timing of variations in the rate of relative sea-level rise identified in the data from North America. Whilst the same constraints on determining the precise timing of change applies, these data appear to indicate that the North American records do not exhibit the same lag as those from southern Britain. Consequently, changes in relative sea-level recorded from North America appear to lead similar variations noted in southern Britain by a few centuries.

If this pattern is correct, it may indicate changing ocean circulation is the principal agent of relative sea-level change in the late Holocene period. Following a change in Gulf Stream strength, variations are first recognised in the western North Atlantic before propagating across the ocean in a similar way to the temperature anomalies described in Section 2.2.3. This would explain the spatial and temporal variability of climate periods such as the Little Ice Age since warming or cooling would be realised at different times in the northern hemisphere, reflecting the redistribution of heat and mass by ocean currents. This idea supports the contention of Mörner (1983, 1995) that climate and sea level changes in the last 5000 to 6000 Cal. years are the result of changes in the strength of the Gulf Stream.

7.3 CONCLUSIONS

The precision of current dating techniques and the inevitable wide age ranges resulting from the combination of a variety of records means that at present, precise correlation of an observed change in relative sea-level with a climate event is impossible. Within the limitations of these techniques however, the relative sea-level records produced from southern Britain are consistent with documented variations in climate. These data appear to indicate changes in oceanic circulation are principally responsible for the observed variations in relative sea-level, whether this be directly through ocean temperature and dynamic changes, or indirectly by forcing atmospheric perturbations. Furthermore, these oceanic changes appear to be registered first along the western North Atlantic coast suggesting that North American sea-level data will lead records from Europe by a number of centuries.

If reliable, an important implication of these data is that the most recent acceleration in the rate of relative sea-level rise observed in the Solent record does not reflect the rise in temperature that has occurred over the last 150 years. Instead, this rise appears to be a response to changes in oceanic circulation that clearly predate any human disturbance of the system. This is consistent with the conclusions of Nydick *et al.* (1995) who state that the most recent acceleration in their record cannot be related to an enhanced greenhouse effect.

It must be stressed however, that diachronous events may be endowed with artificial synchroneity by miscorrelation associated with the large age uncertainties inherent in the climate and sea-level records, or by invoking an apparent lead/lag effect. This means that it is impossible to reject the null hypothesis that there is no climate signature within the relative sea-level records from southern Britain. The only certain conclusion is that existing data distribution, accuracy and precision are of insufficient quality to unequivocally determine a climate signal within the relative sea-level records of southern Britain.

7.4 SUMMARY

- The precise timing of variations in climate and sea-level are subject to age uncertainties resulting from the combination of a variety of records, and limitations in dating associated with their generation.

- Consequently, synchronous changes in climate and sea level may appear offset due to limitations in defining the age of an event or lags in its recognition in the relative sea-level record. Alternatively, time lags involved between a change in climate and its effect on sea level will resulting in increased diachroneity.

- These limitations mean that synchroneity is not the definitive measure of a climate-sea level relationship, but in the absence of more a reliable alternative, such visual comparison is the only viable method of investigation.

- The scenario of relative sea-level change presented in Section 2.7 correlates well with the record produced from southern Britain presented in Section 6.7.

- *RSLC-I* corresponds to the transition between *Phase-I* and *Phase-II*;

- *RSLC-II* appears to predate the acceleration in the rate of relative sea-level rise noted in *Phase-III*.

- *RSLC-III* correlates closely with the inferred reduction in the rate of relative sea-level rise characterising *Phase-IV*.

- *RSLC-IV* may correspond with the most recent acceleration noted in *Phase-V*.

- *RSLC-V* is suggested to have no expression in the relative sea-level record from southern Britain, indicating that the effects of human activities on climate have yet to be recognised. The possibility remains however, that the shift to minerogenic sedimentation noted at Arne Peninsula is indicative of this most recent acceleration.

- Whilst correlation does not prove causation, the record of relative sea-level change from southern Britain is entirely consistent in timing and magnitude with that expected to arise as a consequence of climate change. From these data there is no justification for the rejection of the hypothesis that 'late Holocene relative sea-level change in southern Britain reflects documented climate changes during this period'.

- The limitations inherent in the climate and sea-level records considered in this thesis also means that the null hypothesis proposing no climate signal in the relative sea-level records from southern Britain cannot be entirely refuted. This must serve as a cautionary note when considering the suggestions presented below.

- The climate data appear to indicate that changes in oceanic circulation are principally responsible for the observed variations in relative sea-level, whether this be directly through ocean temperature and dynamic changes, or indirectly by forcing atmospheric perturbations.

- Oceanic changes appear to be registered first along the western North Atlantic coast suggesting that North American sea-level data will lead records from Europe by a number of centuries.

- There appears to be no indication of an acceleration in the rate of relative sea-level rise associated with human-induced global warming. Whilst an acceleration in the rate of relative sea-level rise is noted in the most recent part of the record, biostratigraphic data suggests this had already begun prior to 200 Cal. BP. This implies that any effects of human-induced global warming have yet to be registered as relative sea-level rise in southern Britain.

CHAPTER EIGHT

Conclusions

This thesis represents the first attempt to construct a high resolution record of late Holocene relative sea-level change from the saltmarshes of southern Britain with the aim of evaluating it for evidence of a climate signal. It combines the age-altitude approach commonly employed in many sea-level investigations with the foraminiferal-based 'marsh palaeoenvironmental curve' technique pioneered in the saltmarshes of Atlantic North America. It also expands the work of Horton (1997) by employing a foraminiferal-based transfer function to reconstruct changes in water depth from fossil saltmarsh sequences.

In Chapter One, the following three research aims were identified:

1. The generation of a high resolution record of late Holocene relative sea-level change for southern Britain;

2. Investigation of the hypothesis that late Holocene relative sea-level change in southern Britain reflects documented variations in climate during this period;

3. The evaluation of the applicability and efficacy of existing research techniques to the study of late Holocene relative sea-level change.

This chapter evaluates the extent to which these research aims have been realised and identifies possible avenues for further work.

8.1 METHODOLOGICAL EVALUATION

Before discussing the late Holocene relative sea-level record presented in Chapter Six, it is first desirable to evaluate the success of the methodology employed to produce it. In this way limitations associated with the results will be fully appreciated.

8.1.1 The Foraminiferal-Based Transfer Function

In Chapter Five a transfer function was developed with the aim of producing quantitative reconstructions of water depth changes from fossil foraminiferal assemblages. This was then used in Chapter Six to infer periods of marsh submergence and emergence. Application of the transfer function to the fossil foraminiferal assemblages exposed a number of weaknesses in its ability to accurately reconstruct water depth changes and consequently the results were only used qualitatively to infer submergence or emergence.

8.1.1.1 Limitations

The transfer function's primary downfall is its inability to assign accurate $SWLI_{(Pred)}$ to samples at the extremes of the environmental gradient. This causes the elevation of highest marsh samples to be underestimated and reduces its ability to resolve change in the low marsh to mudflat environment. Consequently, sea-level index points derived from the highest marsh environments have to be visually assigned a SWLI and indicative range, whilst the magnitude of submergence events tend to be understated. This situation arises partly as a consequence of fundamental limitations inherent to the weighted-average technique used to produce the transfer function (Section 5.7.3.1). More significant in this instance however, are the behaviour of saltmarsh foraminifera and the quality of the modern training set.

The foraminiferal assemblages most sensitive to relative sea-level change are those occupying the highest marsh environment, at the interface between marine and terrestrial conditions (Scott & Medioli, 1980a). Unfortunately these assemblages are predominantly of very low abundance, particularly in the UK where fossil foraminiferal concentrations are lower than those recorded in North American marshes. This renders them poorly suited to statistical analyses, necessitating most of the sensitive assemblages to be screened out on the basis of low counts during development of the transfer function. A further complication is the post-depositional dissolution of calcareous tests which serves to further lower sample counts and reduce the diversity of taxa available. This required the artificial 'dissolution' of modern saltmarsh samples and the utilisation of foraminiferal test linings as a proxy for calcareous taxa. This is undesirable since test linings are very light and easily lost during sample preparation. Furthermore, changes in the calcareous assemblage can no longer be used to refine the water depth reconstruction, resulting in $SWLI_{(Pred)}$ values 'saturating' toward the lower elevations and becoming incapable of discerning further increases in water depth.

This situation is compounded by the limited vertical range over which modern foraminiferal samples have been collected. The contemporary database consists of saltmarsh to mudflat assemblages but logistical limitations rendered it impractical to collect subtidal samples which may serve to increase the predictive range of the transfer function. An improved knowledge of sub-tidal foraminiferal distributions would help to distinguish between those fossil samples that possess no modern analogue as a consequence of a change in marsh environment, and those that merely represent a deeper water, unsurveyed assemblage.

8.1.1.2 Possible Solutions

Recent research by Charman *et al.* (1998) indicates that freshwater testate amoeba (thecamoebians) are vertically zoned with respect to tide level and occur in the high marsh to upland environment. Incorporation of thecamoebians with saltmarsh foraminifera would allow precise delimitation of the transition from marine to freshwater environments. Furthermore, since the upper limit of marine influence would reside in the middle of the environmental gradient rather than at the upper extreme, the problems associated with underestimation described above would be removed.

Horton (*pers. comm.*) is currently working on improving his transfer function's predictive precision by employing the statistical technique of bootstrapping advocated by Birks (1995). This is capable of producing sample specific errors and estimates and consequently $SWLI_{(Pred)}$ generated from taxa with small tolerances can be given smaller vertical ranges than less tolerant species. Any refinement that this approach may bring however, will be unimportant until sufficiently diverse modern training sets accompanied by thecamoebian data are compiled. Unless the accuracy of a $SWLI_{(Pred)}$ value can be assured there is little merit in attempting to improve its precision.

8.1.2 Dating Limitations

The second major problem encountered in the course of this research was the collection of suitable organic material for dating. The general decline in the abundance of coastal organic deposits during the last 3000 Cal. years has been discussed in Chapter Two and will not be reiterated here. This paucity of material necessitated sampling of largely minerogenic deposits and the use of AMS radiocarbon dating. No identifiable plant macrofossils were present in these sediments and it was therefore necessary to date the entire sample. Many of the samples returned anomalously old age estimates strongly suggestive of contamination by old carbon. The virtual absence of high marsh samples required lower elevation samples to be collected. These environments are clearly more susceptible to the introduction of allocthonous material owing to their more regular inundation and greater current velocities.

8.1.3 Site Sensitivity

In Chapter Three protected sites with low tidal ranges, and correspondingly small saltmarshes, were suggested to offer the best chance of producing high resolution records of relative sea-level change. This proposition was tested by investigating small (*c.* 150m wide) marshes in micro-tidal Poole Harbour and large (up to 1500 m wide) marshes in the macro-tidal Loughor Estuary. The modern distribution of saltmarsh foraminifera in Poole Harbour displayed a distinct vertical zonation that facilitated production of the transfer function described in Chapter Five. The situation in the Loughor Estuary however, was much more complex. Here no vertical zonation of foraminifera was apparent, indicating that salinity does not vary uniformly with altitude across the large and topographically heterogeneous marshes at Llanrhidian. These findings are consistent with those of De Rijk (1995a) and De Rijk and Troelstra (1997) who observed a similar absence of vertical zonation at the irregular and expansive Great Marshes in Barnstable, Massachusetts.

The fossil foraminiferal assemblages contained within saltmarsh deposits in the Loughor Estuary were also of much lower diversity and abundance than their counterparts in Poole

Harbour. This may be a consequence of the reduced periods of inundation experienced in the highest marsh environments at Llanrhidian, where conditions approximate more closely to a terrestrial environment. The highest marsh is inundated during very high tides but the development of **physically ripe** soils at Llanrhidian indicates that they experience prolonged periods of subaerial exposure. Furthermore, the influx of freshwater onto the marsh serves to limit the influence of saline waters (Bridges, 1977). Together, these conditions appear to be poorly suited to the maintenance of a rich foraminiferal population or its preservation in sub-surface deposits. This evidence supports the initial conjecture that sites with low tidal ranges will be better suited to the construction of high resolution records of relative sea-level change.

8.1.4 Implications for the UK record of relative sea-level change

The foraminiferal techniques applied in this thesis have been developed in the marshes of North America. The foraminiferal abundances in these marshes are substantially higher than those observed in the UK and it is therefore quicker and easier to obtain statistically significant counts from these environments. Furthermore, the North American marshes (which are largely micro-tidal) comprise predominantly organic sediments, possessing well-preserved plant macrofossils that can be individually dated by the AMS radiometric technique. This permits variations in water depth determined from fossil foraminifera to be precisely dated, and research is in progress to produce a high resolution record of late Holocene relative sea-level change employing forty to fifty AMS radiocarbon dates from a single, two metre core (van de Plassche, *pers. comm.*).

This type of resolution is not obtainable from the minerogenic deposits that typify the late Holocene period in southern Britain, where erosion and non-deposition has removed portions of the sea-level record. Furthermore, the distribution of radiocarbon dates is dictated by the availability of suitable sediment resulting in long periods with no age control. This research demonstrates that it is possible to obtain some information on late Holocene relative sea-level change from the UK but that careful selection of sites and extensive reconnaissance coring is essential. Until a method of dating minerogenic sequences throughout the time period covered by the late Holocene is developed, the record of relative sea-level change from southern Britain will have to be pieced together from a series of 'snap-shots'. The need for almost continuous 'monitoring' of change is particularly important when investigating the link with climate since changing rates of relative sea-level rise are of paramount significance. Widely spaced sea-level index points may mask a number of accelerations or decelerations and give the impression of apparent stability. Furthermore, it is rarely possible to delimit the beginning and end of a phase of relative sea-level change precisely and consequently quantitative measurements of changing rates of rise cannot be made.

8.2 Late Holocene Relative Sea-Level in Southern Britain

Phase	Description
I	A rise in relative sea-level between *c.* 4800 Cal. BP and *c.* 2800 Cal. BP resulting in the inundation and preservation of peat deposits
II	A deceleration in the rate of relative sea-level rise probably resulting in a still-stand or fall in relative sea-level between *c.* 2800 Cal. BP and *c.* 1500 Cal. BP. During this time emergence appears to have occurred although the period is characterised by extensive erosion of coastal deposits removing significant portions of the record
III	A renewed rise in relative sea-level commencing prior to *c.* 1500 Cal. BP and terminating around 900 Cal. BP
IV	A period of comparatively stable relative sea-level persisting until *c.* 300 Cal. BP
V	A recent acceleration in the rate of relative sea-level rise during the last 300 years

Table 8.1 *Phases of relative sea-level change inferred from the study marshes in the Solent*

A combination of lithostratigraphic, biostratigraphic, and chronostratigraphic data from three sites in the Solent Region indicates five major phases of relative sea-level change during the late Holocene. These are summarised in Table 8.1.

The record from the Loughor Estuary is of much lower resolution that that available for the Solent region but exhibits variations consistent with those expected to arise from the five phases of relative sea-level change outlined above.

8.3 EVIDENCE FOR A CLIMATE SIGNAL IN THE LATE HOLOCENE RELATIVE SEA-LEVEL RECORD FROM SOUTHERN BRITAIN

In Chapter Two a scenario of predicted relative sea-level change was generated on the basis of documented changes in climate during the late Holocene period. This consisted of the following five phases of change:

RSLC-I: A reduction in the rate of relative sea-level rise *c.* 3000 Cal. BP;

RSLC-II: An increase in the rate or relative sea-level rise after 2000 Cal. BP;

RSLC-III: A reduction in the rate of relative sea-level rise around 1000 Cal. BP to 800 Cal. BP;

RSLC-IV: An increase in the rate of relative sea-level rise around 500 to 300 Cal. BP;

RSLC-V: A possible acceleration in the rate of relative sea-level rise during the last 150 years.

This scenario of relative sea-level change correlates well with the record from southern Britain presented in Table 8.1, and this relationship is summarised in Table 8.2.

RSLC	Description
I	This corresponds to the transition between *Phase-I* and *Phase-II*
II	This appears to predate the acceleration in the rate of relative sea-level rise noted in *Phase-III* although it is suggested that this had begun prior to 1500 Cal. BP
III	This correlates closely with the inferred reduction in the rate of relative sea-level rise characterising *Phase-IV*
IV	This has no synchronous correlative with the record from southern Britain. Noting the timelags involved in the recognition of the previous acceleration in the rate of relative sea-level rise however, it is suggested that *RSLC-IV* corresponds to the most recent acceleration noted in *Phase-V*
V	This is suggested to have no expression in the relative sea-level record from southern Britain, indicating that the effects of human activities on climate have yet to be recognised. This is consistent with the tide-gauge data that show no acceleration in the rate of relative sea-level rise during this period

Table 8.2 *A comparison between the scenario of relative sea-level change suggested from the climate-based prediction and the phases of relative sea-level change recorded in the study marshes*

I have approached the subject of a climate signal in late Holocene records of relative sea-level change from a critical, even sceptical, perspective. The frailties of existing climate and sea-level records and inadequacies in the techniques and methodologies used to produce them have been highlighted. Nevertheless, the results presented above, whilst subject to these limitations, are consistent in terms of the sequence, timing, and magnitude of events that were predicted to arise on the basis of the climate data outlined in Chapter Two. Whilst in isolation the apparent synchroneity of two poorly delimited events is equivocal, the fact that the four correlative episodes outlined above are constrained stratigraphically and occur in the correct sequence strongly argues against the idea of a purely coincidental occurrence. On balance therefore, whilst the precise relationship remains uncertain, there are strong grounds for suggesting a possible climate signal in the relative sea-level record from southern Britain, and these results recommend further investigation.

8.4 IMPLICATIONS FOR FUTURE SEA-LEVEL RISE PREDICTIONS

The comparison of a variety of climate data with the relative sea-level record from southern Britain suggests that variations in oceanic circulation may be the principal agents by which changes in climate influence sea level (Chapter Seven). The spatial variability apparent in terrestrial records of temperature change suggests that averaged values of atmospheric temperature derived from a wide geographical area may not reveal climate change when this is related to redistributions of heat or changes in circulation. The apparently close relationship of relative sea-level change in southern Britain with variation in oceanic circulation suggests that future rises in sea level will vary spatially. At present IPCC predictions of future sea level rise are based upon suggested increases in the volume of ocean water that are then converted to an equivalent global rise in ocean level. Spatial variability however, means that predictions of future local relative sea-level rise required by planners and decision makers cannot simply be produced by combining a global estimate of eustatic sea level rise with the known crustal movements of the area. The use of coupled ocean-atmosphere-land surface models (e.g. Manabe & Stouffer, 1997) appears to offer a promising method by which such local to regional variability may be predicted in the future.

Whilst there are a number of limitations associated with the dating of the climate and sea-level records produced in this thesis (Section 7.1.1) there is some evidence to suggest a time lag in the region of two to four hundred years exists between the recognition of a warming in climate from oceanic records and its manifestation as an acceleration in the rate of relative sea-level rise. If this assertion is correct, it would suggest that human-induced global warming may become apparent in the tide-gauge records from southern

Britain within the next century. This assumes that climate change 'driven' by human alteration of atmospheric composition proceeds in the same way as previous variations.

8.5 RECOMMENDATIONS FOR FUTURE WORK

In this thesis, I have demonstrated that there is some evidence for a climate signal within the relative sea-level records of southern Britain. These records of late Holocene relative sea-level change are only fragmentary however, and there remain long periods (e.g. *Phase-II*) where data are absent. It is therefore important that more data from the late Holocene period be collected from around the Solent region to confirm or refute the pattern of change suggested in Chapter Six. In particular, data from the last 2000 Cal. years is required to more precisely constrain the timing of suggested variations in the rate of relative sea-level rise during this period. Furthermore, the techniques employed in this thesis can be applied to other localities around the UK to examine whether a similar pattern of change is observed elsewhere. Data from uplifting areas of the UK may be able to constrain the rate of relative sea-level rise if suitable sedimentary sequences can be located. As the body of late Holocene data increases it will become possible to apply statistical methods such as time series analysis, to determine whether the climate and relative sea-level records are systematically linked. At present however, the science remains in its infancy and the emphasis must be placed on collection of high quality data from the critical time periods outlined above.

The use of transfer functions to produce quantitative reconstructions of changing water levels offers the potential for reconstructing more detailed pictures of relative sea-level change. Before these can be employed in the UK however, an improved modern training set consisting of foraminifera and thecamoebians is required.

Accurate age estimation remains the principal obstacle to the construction of high resolution records of relative sea-level change in the UK. The British Isles possesses a rich archaeological heritage and there is a growing appreciation of the wealth of material present in the modern coastal zone (e.g. Fulford *et al.*, 1997; Tomalin *et al.*, in press). Multi-disciplinary studies combining archaeological, palaeoenvironmental, and sea-level investigations may offer the potential for elucidating the timing and magnitude of change during periods where suitable material for radiocarbon dating is absent.

The importance of changing oceanic circulation has been highlighted during the course of this research. There is a clear need for more detailed information regarding late Holocene variations in Gulf Stream strength, and changes in key areas such as the Greenland, Iceland, Norwegian, Labrador, and Sargasso Seas. It is perhaps unsurprising that the secret of sea level response to climate change may rest within the oceans themselves.

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