Autocompaction of mineralogenic intertidal sediments

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To my Mum and Dad

Declaration

I confirm that no part of the material presented in this thesis has previously been submitted for a degree in this or any other university. In all cases the work of others, where relevant, has been fully acknowledged.

Matthew Brain, September 2006

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ABSTRACT

Autocompaction is a syn- and post-depositional diagenetic process that results in a volumetric reduction of sediments. An understanding of this process is required to inform conceptual and numerical models of relative elevation changes in intertidal environments. Autocompaction therefore introduces an error into studies of Holocene relative sea level changes obtained from intertidal sediments and, unless corrected for, results in an overestimation of the rate and magnitude of historical relative sea level variation.

Models of autocompaction previously applied to intertidal sediments, namely Terzaghi's compression law and consolidation equation, are based on conventional soil mechanics theories that focus on homogeneous, fully saturated, normally consolidated clays. The applicability of these models to intertidal sediments has hitherto been accepted despite a striking lack of empirical data describing the one-dimensional compression behaviour of these sediments.

This study presents results of field and laboratory investigations into the autocompaction behaviour of predominantly mineralogenic (i.e. > 50 % dry mass) intertidal sediments obtained from Greatham Creek, Teesside, UK. Analysis of the initial structure and compression behaviour of these materials using oedometer tests, demonstrates that low marsh and mudflat sediments are significantly overconsolidated. Preconsolidation stresses ranged between 20 – 27 kPa in the low marsh samples and 8 – 14 kPa in the mudflat samples. This overconsolidation is a result of loading by tidal waters, suction pressures resulting from groundwater downdraw and subaerial desiccation.

Temporal variations in effective stress at the depositional surface are calculated using hydrographic and groundwater data from a local tide gauge and piezometer in order to assess their impact on compression behaviour. During the 15-month study period, effective stresses throughout the stratigraphic column varied continuously. Tests using a back-pressured shear box simulate ground conditions in a realistic manner by superimposing the dynamic loading observed in the field on the longer-term burial stress. Results indicate that time-dependent creep processes are unlikely to act on the sediments because of the increased structural stability (overconsolidation) resulting from tidal and capillary surcharge loading.

Analysis of vadose zone stratigraphy and geochemical profiles suggest that diagenetic alteration of compression behaviour may be possible. However, statistical analysis of the effects of redox-controlled geochemical compounds suggests that voids ratios are primarily controlled by effective stress variations and diagenetic influences are unimportant. Oedometer tests undertaken on near-surface samples that display enrichment in diagenetic compounds reveal no significant influence of diagenetic point-contact cementation on compression behaviour.

These experimental data are used to develop empirically-based autocompaction models. The statistical nature of these models allows the full range of observed variations in structure and compression behaviour to be described. 'Conditional' regression models were employed to account for the overconsolidation in the sediments.

These models are applied to short stratigraphic sequences containing a twentieth century sea level record dated using radionuclides. Once decompacted, the rates of sea-level rise obtained from the reconstructions match that of a compaction-free tide gauge record of relative sea level variation. In contrast, decompaction using Terzaghi's compression law results in an overprediction of the historical rate of sea level rise by 1.1 mm yr^{-1}.

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CHAPTER 1: INTRODUCTION TO THE STUDY

1.1 RESEARCH CONTEXT AND MOTIVATION

Autocompaction is a fundamental syn- and post-depositional diagenetic process that causes a volumetric reduction of sediments (Allen, 1999). In low energy intertidal environments characterised by landforms such as mudflats and saltmarshes, autocompaction lowers the elevation of the land surface relative to the intertidal frame (Cahoon *et al.*, 1995). If this process occurs at a rate faster than that at which the landforms can vertically build upwards within the intertidal frame, the possibility exists that saltmarsh vegetation will no longer be able to function physiologically and will die (French, 1993; French and Spencer, 1993; Reed, 1990; 1995). This leads to the loss of socio-economically and intrinsically valuable intertidal wetlands (Spurgeon, 1999). These environments offer considerable benefits to society. In particular, the frictional drag created by saltmarsh vegetation reduces the energy of storm waves, offering a cost-effective, 'soft' engineering method of protecting against tidal wave damage (French, 2006; Möller *et al.*, 1999).

In order to protect intertidal landforms, considerable research has been undertaken into the processes that result in their creation and permit their survival in the face of rising sea levels (Davidson-Amott *et al.*, 2002, for example). Conceptual models have been developed to predict elevation change and hence inform management decisions about the best way to intervene to protect intertidal environments (e.g. Allen, 1990; French, 1993; Rybczyk *et al.*, 1998). However, the effectiveness of these models is critically dependent upon empirically informed, scientific inputs of each of the biogeomorphological processes operating in these environments, including autocompaction.

In addition, saltmarsh sediments also offer a useful tool to researchers working to reconstruct historical relative sea level changes. Variations in flooding frequency and duration result in an ecological and lithological zonation of sea level indicators within the intertidal frame that show a direct relationship with reference water levels such as mean sea level (Horton and Edwards, 2006). When such assemblages are found in Holocene intertidal stratigraphies, the former position of sea level can be fixed. When placed into a chronological framework by an appropriate dating method, temporal variations in mean sea level can be reconstructed (Van De Plassche, 1986). However, such reconstructions are only valid providing the 'sea level index points' are still at the altitudes at which they were deposited (Shennan and Horton, 2002). Unfortunately, autocompaction lowers sea level index points from their original, depositional altitudes (Allen, 2000a; Edwards, 2006). Unless corrected for, autocompaction results in an overestimation of the rate and magnitude of historical relative sea level variation in areas where relative sea level is rising (Shennan *et al.*, 2000).

Existing models of autocompaction that are required to predict elevation and altitude changes have been developed using an inflexible and phenomenologically inaccurate civil engineering framework. The theories are based on saturated clays and therefore do not consider the mineralogenic and organics silts and saltmarsh peats that are deposited and form in highly dynamic intertidal environments. The basic mechanisms of one-dimensional compression behaviour of such sediments are unknown (Pizzuto and Schwendt, 1997). Indeed, no empirically-informed framework of intertidal sediment autocompaction behaviour has been developed based on detailed field and laboratory investigations. This prevents the development of realistic conceptual and mathematical descriptions of autocompaction in the intertidal zone.

1.2 RESEARCH APPROACH AND FOCUS

This study addresses the autocompaction problem from a multidisciplinary perspective by combining palaeoenvironmental, geomorphological and soil mechanics theory and methods. Conventional soil mechanics techniques have been adapted in a novel manner to better reflect the dynamics of effective stress variation in the upper intertidal zone. Analysis of diagenetic alterations to recent, twentieth century stratigraphic profiles provides additional insight into the factors controlling volumetric and structural change in intertidal sediments.

The approach taken in this research is principally geotechnical, focusing on the compression behaviour of predominantly mineralogenic sediments. In contrast, the volumetric evolution of organogenic sediments is perhaps primarily controlled by biochemical decay of organic compounds and structures (Allen, 2000b). Such sediments are not considered in this study due to the considerable complications associated with their autocompaction behaviour.

1.3 AIM AND OBJECTIVES OF THE RESEARCH

The principal aim of this study is to develop a quantitative model of the autocompaction behaviour of contemporary and recently formed mineralogenic intertidal sediments based on field and laboratory data. The associated objectives are to:

- review current understanding of the autocompaction of intertidal sediments in order to identify existing knowledge gaps;
- investigate and quantify the causes of effective stress changes within intertidal environments;
- undertake a detailed geotechnical testing program that has been tailored to the specific effective stress conditions of the dynamic intertidal environment;
- identify the main settlement processes that cause volumetric variations in mineralogenic intertidal sediments;
- consider the importance of diagenetic processes in affecting the compression behaviour of these sediments;
- develop an empirically-informed predictive model of the autocompaction behaviour of mineralogenic intertidal sediments;
- apply the empirically-informed predictive model to a recent stratigraphic sequence and compare the accuracy of its predictions to those of previously employed models of autocompaction;
- consider the broader implications of the empirically-informed predictive model and speculate on its transferability;
- identify areas for future research.

1.4 ORGANISATION OF THE THESIS

The structure of this study reflects the sequential treatment of the thesis objectives and a progressive development of an understanding of autocompaction processes in mineralogenic intertidal sediments.

Chapter 2 expands on the rationale for developing an understanding of autocompaction by demonstrating its role as an active geomorphological process in low energy intertidal areas. It also reviews aspects of the sea level literature and illustrates how autocompaction confounds attempts to accurately quantify historical rates and magnitudes of relative sea level variations.

Chapter 3 outlines previous attempts to model autocompaction in intertidal environments. Existing approaches to modelling sub-sets of autocompaction processes developed in disciplines other than intertidal geomorphology are discussed. Models of autocompaction that have been applied to saltmarsh and mudflat sedimentary sequences are also outlined. The degree to which these models sufficiently account for the specific conditions experienced in the dynamic intertidal environment and the types of materials that form there are then examined. The chapter concludes by focusing the study to deal with recently deposited mineralogenic sediments, and the reasons for doing so. It suggests ways in which existing models of autocompaction can be improved by providing the research hypotheses that will be addressed in the experimental phase of the research.

Chapter 4 introduces the field site at Greatham Creek, Cleveland, UK. A justification is provided for the site selection. The field and laboratory methods that are employed in this study are introduced.

Chapter 5 examines the contemporary intertidal environment at Greatham Creek in terms of the types of sediments that form and are deposited there and the magnitude, frequency and periodicity of effective stresses applied to these materials. The sediment types for analysis in the thesis are selected. The idiosyncrasies of the intertidal environment and lithologies are compared to the assumptions of existing theories and methods. Accordingly, recommendations are made for the methods in order to better reproduce the effective stress variations in intertidal environments and hence improve theories and models of autocompaction behaviour.

Chapter 6 presents the material testing program that was designed on the basis of the conclusions reached in Chapter 5 regarding effective stress variations within the intertidal environment. The results of this detailed geotechnical testing program are then presented. The implications of these data for the development of accurate models of autocompaction in mineralogenic intertidal sediments are considered with reference to the research hypotheses outlined in Chapter 3.

Chapter 7 presents the results of an investigation into the operation of diagenetic processes in the overconsolidated vadose zone and their influence on compression behaviour. The relative importance of stratigraphic variations in effective stress, lithology and geochemistry on sediment structure are assessed. Further geotechnical test data are also presented to ascertain whether post-depositional processes significantly affect the compression behaviour of intertidal sediments.

Chapter 8 synthesises the main findings of the research in this study and develops models on the basis of these findings. Some practicalities of application of the empirically-based predictive models are addressed before the implications of these newly-developed autocompaction models are discussed. The models are then directly applied to a short stratigraphic sequence. By comparison with compaction-free records, an existing autocompaction model is compared with the newly-developed models to determine the accuracy of the predictive capacity of each model. The transferability of the model to different lithologies and different intertidal environments is also discussed.

Chapter 9 presents the conclusions of this study. The thesis objectives are readdressed and recommendations for future research are made.

CHAPTER 2: RESEARCH RATIONALE

2.1 INTRODUCTION

Rising sea level threatens coastal areas (Adger *et al.*, 2005; Pethick, 2001), potentially leading to a variety of problems such as permanent inundation and loss of coastal land (Dawson *et al.*, 2005; McInnes *et al.*, 2003; Nicholls, 2002; Walker *et al.*, 1987), erosion of coastal features, damage to structures (French, 2006; Li *et al.*, 2004) and saltwater intrusion into rivers, estuaries and coastal aquifers (Capaccioni *et al.*, 2005). Specific issues associated with rising sea level range from the potential for vulnerable, low-lying regions to become 'source areas' of 'eco-refugees', particularly in densely-populated areas (Adger *et al.*, 2005), to the protection of coastal landforms, such as saltmarshes, mangroves and coral reefs that provide significant benefits to society (DeLaune *et al.*, 2003; Spurgeon, 1999; Stokstad, 2005).

The ability to successfully manage the coastal zone in the face of rising sea level is of obvious importance (Leafe *et al.*, 1998) and is dependent upon a thorough scientific understanding of the behavioural response of coastal landforms to future changes in sea level, particularly in low-gradient coasts where the (bio-)geomorphological functioning of landforms is intimately linked to variations in sea level and wind and wave processes. Of equal importance is a detailed understanding of the forcing factors that control sea level on a variety of temporal and spatial scales. Attempts to achieve each of these goals is confounded by a lack of empirical data regarding specific processes and the timescales over which they operate (Davidson-Amott *et al.*, 2002).

There is a particular need to better understand the process of sediment autocompaction in upper intertidal environments (saltmarshes and mudflats) in order to improve our understanding of the morphodynamics of low energy intertidal systems and reduce altitudinal errors in studies of historical relative sea level (RSL) variations.

2.2 SALTMARSHES: INTRODUCTION AND SOCIO-ECONOMIC AND INTRINSIC VALUE

Saltmarshes are well-vegetated, saline flats which develop high in the intertidal frame, usually between mean high water neap (MHWNT) and mean high water spring (MHWST) tide levels (Frey and Basan, 1985). They form by the deposition of primarily fine-grained and organic sediments at a level that allows regular tidal flooding, although not on every tide (Allen and Pye, 1992b; Packham and Willis, 1997; Reed, 1990). Saltmarsh surfaces are vegetated by halophytic species which can tolerate regular saline flooding. The spatial distribution of halophytes is organized in characteristic 'zones' of (typically) low-, mid- and high- marsh vegetation communities (Figure 2.1). The spatial and elevational (i.e. in relation to the local tidal frame) distribution of these communities is determined by a complex interplay between physical, chemical and biotic factors affecting plant physiology that are related to variations in flooding frequency, wave/tidal energy, salinity and saturated/unsaturated flow through the substrate (Silvestri *et al.*, 2005). A network of channels, which narrow and bifurcate upstream, is commonly present (Allen, 2000). This carries the tidal flood onto the marsh and allows the ebb to drain away. Saltmarshes grade downwards into tidal mud- or sand-flats (Figure 2.1). These are unvegetated and typically have fewer and shallower channels. A sharp cliff or a more gentle topographic gradation may mark the boundary between the two environments (Figure 2.1).

The ecological and biogeomorphological character of saltmarshes varies through space and time. A distinction can be made, for example, between the predominantly mineralogenic (i.e. deriving the majority of their lithological make-up from tidally-advected mineral matter) saltmarshes of northwest Europe and their more organic-rich north American counterparts (Allen, 2000).

Saltmarshes and the surrounding wetlands provide considerable direct and indirect benefits to society in the form of products and services. In terms of direct economic value, saltmarshes have been exploited by humans for many centuries. Livestock grazing and haymaking began on Dutch saltmarshes at around 600 BC (Esselink, 2000). Beeftink *et al.* (1982) discuss how saltmarsh plants can be harvested for human consumption. Gordon (1988) summarises the use of saltmarshes for small-scale salt production using evaporation ponds. More significantly, saltmarshes effectively dissipate wind and wave energy due to the frictional effects of their topographically complex, vegetated surfaces (Möller *et al.*, 1999; Möller and Spencer, 2002). This reduction in wave energy provides a cost-effective, 'soft' engineering means of buffering high-energy storm tides and floodwaters (French, 2006; Knutson, 1988), reducing any damage to the terrestrial hinterland. Furthermore, the root-rhizome mat produced by marsh vegetation results in increased sediment shear strength, lowering the rate and amount of erosion in comparison to unvegetated shorelines (Crooks and Pye, 2000; Knutson, 1988). Saltmarshes also act as reservoirs for sediments, play an important role in carbon sequestration (Twilley *et al.*,

1992) and act as a sink for heavy metals and organic pollutants (Spencer *et al.*, 2003; Williams *et al.*, 1994).

Saltmarshes are important areas for a whole range of flora and fauna, but particularly for birds, both directly, offering feeding, roosting and nesting sites, and indirectly, *via* their position as primary producers in the estuarine food web, exporting organic carbon to adjacent habitats, particularly mudflat invertebrates (Hughes, 2004). The large numbers of migratory wildfowl and wading birds increase the socio-economic value of saltmarshes by providing recreational opportunities for birdwatchers.

2.3 DESTRUCTION AND LOSS OF INTERTIDAL WETLANDS

2.3.1 General threats to saltmarsh survival

Considerable areal losses of saltmarshes, mudflats and sandflats have occurred throughout the twentieth century (Orr *et al.*, 2003; Pethick, 2002). Such losses have resulted from land use change, ranging from the creation of embankments and sea walls for land claim (Allen, 2000; Allen and Pye, 1992a) and fixed sea defences; the use of recreational vehicles (Adams *et al.*, 2006); altered intertidal hydrodynamics and sediment supply caused, for example, by tidal barrages constructed for the purposes of flood defence and exploitation of tidal power (Carter, 1988); and dramatically increased inputs of pollutants to intertidal areas from industry and oil spills (Cundy *et al.*, 2003). However, the greatest pervasive threat to the long-term survival of intertidal landforms is arguably that of future accelerated sea level rise.

2.3.2 Relative sea level rise

It has long been recognised that global sea level is sensitive to long-term variations in climate as a result of variations in terrestrial ice volume (Church *et al.*, 2001; Milne *et al.*, 2005). An important impact of any anticipated anthropogenically-induced climate change will be an acceleration in the rate of 'eustatic' (absolute) sea level rise as a result of an increase in the volume of ocean water *via* thermosteric expansion of ocean waters; melting of sensitive alpine and glaciers; and enhanced melting of the Greenland and Antarctica ice sheets (Gornitz, 1995).

The Intergovernmental Panel on Climate Change (IPCC) concluded that the average rate of 'eustatic' sea level rise during the 20th Century was between 1 and 2 mm yr^{-1} (Church *et al.*, 2001; Church and White, 2006). Depending on the greenhouse gas emission scenario employed, estimated projections of 'eustatic' sea level rise for approximately the next century range from 9 cm to 88 cm; i.e. between 0.09 mm yr^{-1} and 8.8 mm yr^{-1} with a 'middle value' of 0.4 mm yr^{-1} (Church *et al.*, 2001).

'Eustacy' does not, however, describe the *relative* sea level changes (which includes both sea- and land-level changes, hence determining the position of the shoreline) that can potentially take place at regional and local scales (Cazenave and Nerem, 2004). For a given site, the change in RSL ($\Delta\xi_{rsl}$) at time τ and location φ can be expressed schematically as:

$$\Delta\xi_{rsl}(\tau, \varphi) = \Delta\xi_{eus}(\tau) + \Delta\xi_{iso}(\tau, \varphi) + \Delta\xi_{tect}(\tau, \varphi) + \Delta\xi_{local}(\tau, \varphi) \tag{1.1}$$

where:

$\Delta\xi_{eus}(\tau)$ is the time-dependent 'eustatic' function

$\Delta\xi_{iso}(\tau, \varphi)$ is the total isostatic effect of the glacial rebound process including both the ice (glacio-isostatic) and water (hydroisostatic) load contributions

$\Delta\xi_{tect}(\tau, \varphi)$ is any tectonic effect

$\Delta\xi_{local}(\tau, \varphi)$ is the combined effect of local scale processes at a site, such as sediment autocompaction and tidal range change (Shennan and Horton, 2002).

Near-field sites, undergoing crustal rebound following glacially-induced subsidence during the Late Devensian (such as Scotland) are currently undergoing RSL fall (Shennan and Horton, 2002). Such areas are less vulnerable to the effects of accelerated sea level rise than areas experiencing collapse of the glacial forebulge (such as Southern Britain) or significant autocompaction of the substrate, such as Chesapeake Bay, eastern USA (Kearney *et al.*, 1994).

Throughout the last 6,000 years, under modest rates of RSL rise (e.g. 1 mm to 4 mm yr^{-1}) (French, 1993), geomorphic factors such as sea level or sediment supply have resulted in a spatial reorganization and redistribution of individual, small-scale and larger-scale landforms, such as saltmarsh and estuarine systems respectively. Increased water-depths have led to wave and tidal energy being propagated further landward (Carter, 1988) and so saltmarshes, and indeed estuaries as a whole (though the adjustment time is longer), have migrated both landwards and seawards and grown vertically to maintain their equilibrium position within the intertidal energy gradient (Pethick, 2001). With sufficient space and time, and gradual increases in water depths, flooding frequency and salinity, landward migration of saltmarshes is possible through spatial and altitudinal relocation of marsh vegetation.

Low magnitude (bio-)geomorphic perturbations can be accommodated by a coastal system in equilibrium with forcing factors, providing critical thresholds are not exceeded. Saltmarshes have proven to be resilient to high magnitude, low frequency geomorphic disturbances such as storm surges, hurricanes and similar high energy erosive events. Palaeogeomorphic studies have suggested that it is such episodic events that provide 'elevational capital' that allows saltmarshes to maintain an equilibrium elevation during times of reduced sedimentation (Jennings *et al.*, 1995; Reed, 2002). Coastal landform resilience will be maintained as long as an equilibrium form is re-established in a time less than that of the disturbance event return interval (Knighton, 1998; Pethick and Crooks, 2000). If biogeomorphic disturbance were to persist and time for recovery between threshold events was insufficient, the system would not recover and would undergo progressive or sudden change to a new equilibrium form (Pethick and Crooks, 2000). Such a situation would arise in areas threatened by isostatic subsidence or autocompaction and where inland migration of saltmarshes is prevented by the existence by artificial structures and barriers (French, 2006). Consequently, concern is being raised regarding the ability of saltmarshes to accrete and maintain their elevation within the intertidal frame at a rate at least equal to that of local RSL rise (French, 1993; Orr *et al.*, 2003).

2.3.3 Ecological drowning of saltmarsh environments

If the rate of RSL increase exceeds the rate at which a saltmarsh can accrete vertically, key physical and biological processes that affect the character and persistence of saltmarshes will adversely change. Simultaneous and continued increases in marsh hydroperiod, salinity and soil waterlogging have been shown in both laboratory- and field-based studies to reduce net photosynthesis and stomatal conductance (Pezeshki *et al.*, 1987) leading to increased physiological stress and poor metabolism (Burdick *et al.*, 1989). This results in rapid tissue damage and causes significant reductions in above-ground biomass and stem density (McKee and Mendelssohn, 1989). Mineralogenic accretion may also be limited by sediment supply. This is particularly so where marsh edge erosion is low and sediment is not available for deposition on the marsh surface (Reed, 1988). A reduction in the quality and quantity of vegetation on the marsh prevents effective dissipation of wave energy. This allows erosion of the substrate, which itself has reduced shear strength due to the death and removal of soil-binding vegetation and a reduced sediment-trapping capability. A strong positive feedback cycle develops (Long *et al.*, 2006) where, as the marsh surface collapses (Delaune *et al.*, 1994) and is eroded, local RSL rises. This leads to a further increase in flood water depths, resulting in greater plant stresses, deeper inland penetration of tidal creeks, enhanced marsh drainage, surface subsidence and further marsh degradation. The process by which saltmarsh plant communities are irreversibly damaged, and that may lead to a shift to a new equilibrium landform such as a mudflat environment, has been termed 'ecological drowning' (French, 1993). It eventually results in a conversion to open water and an irreversible areal loss of saltmarsh (Delaune *et al.*, 1983; Delaune *et al.*, 1994; Reed, 1995).

2.3.4 Accretion and elevation deficits

Continuous stratigraphic horizons of saltmarsh sediments indicate that intertidal systems have continued to flourish under mid- to late-Holocene and contemporary, post-industrial revolution rates of RSL rise (Reed, 1990; 1995; 2002; Stevenson *et al.*, 1986). Similar evidence is provided by two types of data, the comparison of which has allowed an assessment of the vulnerability of coastal marshes to accelerated RSL rise:

1. tide gauge data (e.g. Woodworth *et al.*, 1999) which, where present and when the record is sufficiently long and continuous, provide a useful means of assessing local RSL change during the 'Anthropocene' (the most recent period in the Earth's history which began in the 18th century when anthropogenic activities first began to have a significant effect on global climate and ecosystems; Crutzen, 2002);
2. saltmarsh vertical accretion data.

Saltmarsh and mudflat sedimentation rates can be measured by a variety of methods. Analysis of short-term ($<10^1$ years) vertical accretion can be undertaken by monitoring the progressive burial of artificial markers (French, 1993; Harrison and Bloom, 1977; Stoddart *et al.*, 1989)) or by repeated levelling (Carr and Blackley, 1987; Dalby, 1970). Longer term ($10^1 - 10^3$ years) sedimentation rates are determined from dated horizons (Allen and Rae, 1988), x-ray radiographic laminae counts (Shi, 1993) or radionuclide profiles such as ^{210}Pb and ^{137}Cs (Craft *et al.*, 1993; Cundy and Croudace, 1996; Delaune *et al.*, 1983; Delaune *et al.*, 1978; Milan *et al.*, 1995).

Average saltmarsh sedimentation rates, both minerogenic and biogenic, vary through space (both within- and between-site) and time depending upon the relative contributions of RSL change, changes in tidal range and asymmetry, changes in storminess, saltmarsh creek channel variability, vegetation dieback, and variations in anthropogenic control of coastal morphodynamics (e.g. dredging and navigation, embanking and reclamation and mud digging) (Pye, 2000). Consequently, longer-term estimations of vertical accretion are more desirable for comparison with tide gauge data, since these eliminate short-term variations in sediment dynamics.

When the two long-term datasets are directly compared, an accretion deficit occurs when the rate of RSL rise is greater than that of vertical accretion (Baumann *et al.*, 1984; Stevenson *et al.*, 1986) and the potential for permanent submergence is increased. However, such an approach implicitly assumes that that a 1:1 relationship exists between vertical accretion and surface elevation change so that vertical accretion always raises the surface level of the marshes relative to a reference datum. If elevation gain is less than vertical accretion this critical assumption is incorrect and the accretion deficit will underestimate the potential for submergence.

2.4 SEDIMENT AUTOCOMPACTION AND SHALLOW SUBSIDENCE

Vertical accretion is likely to overestimate the rate of marsh surface elevation change (Kaye and Barghoorn, 1964). This is because it does not account for autocompaction processes operating throughout the stratigraphic column which reduce the volume of the sediments and lower the marsh surface relative to the tidal frame, independently of absolute changes in sea level. This contention was later suggested conceptually by Allen (1990) and French (1993) and was empirically proven by field experiments undertaken in Louisiana, Florida and North Carolina, southeastern USA, by Cahoon *et al.* (1995). Simultaneous measurements were made of vertical accretion, determined using feldspar marker horizons, and marsh elevation change relative to a subsurface datum. Elevation change was measured using a Sedimentation-Erosion Table (SET, Figure 2.2) – a levelling device connected to a benchmark pipe that is driven into the Holocene sediments until refusal, assumed to be bedrock or an incompressible Pleistocene surface that provides a stable reference datum. Nine 'pins' at the end of the horizontal arm are lowered to the marsh surface to measure elevation (precision of ± 2 mm).

Shallow subsidence, which refers to the difference between vertical accretion and surface elevation change, can be calculated by subtracting measures of surface elevation change from vertical accretion. Cahoon *et al.* (1995) found that surface elevation change was significantly lower than the observed vertical accretion at each study site throughout the two year study period (Table 2.1). Hence, despite organic and inorganic material accumulation on a marsh surface, this surface moves downward with respect to the intertidal frame because of sediment autocompaction (Figure 2.3). They concluded that surface elevation change is completely decoupled from vertical accretion in some instances and so vertical accretion is not a useful surrogate for elevation change due to the complex operation of subsurface autocompaction processes, which clearly plays a critical role in the geomorphology of intertidal landforms. It is therefore long-term elevation change, rather than vertical accretion, that should be compared with tide gauge data to assess the submergence potential of intertidal landforms.

Table 2.1 Two-year totals of vertical accretion, marsh surface elevation change and shallow subsidence. Source: Cahoon *et al.* (1995).

Site	Vertical Accretion (cm)[a]	Elevation Change (cm)[a]	Shallow Subsidence (cm)[c]	Depth of Benchmark Pipe (m)
Bayou Chitigue, Louisiana	5.19 ± 0.32	0.29 ± 0.15[b]	4.90	4
Old Oyster Bayou, Louisiana	2.07 ± 0.10	1.30 ± 0.09[b]	0.77	4
St. Marks NWR, Florida	0.89 ± 0.06	0.14 ± 0.04[b]	1.03	3
Cedar Island NWR, North Carolina	.77 ± 0.09	0.32 ± 0.11[b]	0.45	5

[a] Data are means ± 1 Standard Error
[b] Indicates that accretion and elevation means are significantly different (p = 0.001)
[c] Shallow subsidence = (vertical accretion) – (elevation change)

2.5 MODELLING SALTMARSH MORPHODYNAMICS

The presence of saltmarshes offers no guarantee that they will continue to survive the IPCC-predicted increased rates of 'eustatic' sea level rise, particularly where these are compounded by isostatic subsidence and autocompaction. The operation of intertidal geomorphic processes and subsequent behaviour of intertidal landforms under an accelerated upward tendency of sea level rise, particularly in these physically- and climatically-sensitive biogeomorphic systems, is largely unknown. Recent and current sedimentation and autocompaction rates cannot be forecast with any confidence. Radionuclide-derived sedimentation rates and SET-based measurements of autocompaction do not offer any insights into the processes behind the elevational adjustments, solely providing a glance at the system under the contemporary regime of RSL rise and offering limited predictive ability.

Stimulated by the threat of future sea level rise, a considerable amount of both conceptual and empirical research has been directed towards an understanding of the morphodynamic operation of saltmarshes and the surrounding (upper-) intertidal environments, specifically mudflats and sandflats. Saltmarshes operate over intermediate geological timescales (decades to millennia), thus prohibiting direct observation of their response to varying forcing factors. In order to partially overcome this problem, conceptual and numerical modelling of the intertidal system as a whole provides a powerful tool that allows the timeframe to be expanded and, by varying the relevant forcing agents, analysing the resulting effects on marsh behaviour.

2.5.1 Conceptual modelling of relative elevation change

Upper intertidal landforms are complex morpho-sedimentary systems that involve a variety of intricately linked components. Numerous conceptual models have been suggested to explain the development of saltmarshes from initiation, through 'youth' and to 'maturity' (Allen, 1990; Beeftink and Rozema, 1988; French, 1993; French and Stoddart, 1992; Pethick, 1969; Steel and Pye, 1997; Yapp *et al.*, 1917; a summary can be found in Allen, 2000). Despite minor differences, each of the proposed models includes common physiographic stages of development.

Based upon the work of several authors, principally Randerson (1979), Reed (1990) and French (1993), Allen (2000) summarised the important elements and linkages involved in the morphodynamic functioning of saltmarsh systems in a flow diagram (Figure 2.4). This conceptual model emphasises the role of four forcing factors.

Two of the forcing factors are external to the saltmarsh environment. Changes in RSL and tidal range either produce or deny 'accommodation space', within which marshes build upward. These factors critically control the rate of vertical sedimentation on a saltmarsh, assuming that the system is not limited by sediment availability. The second external forcing factor is the mineral sediment supply itself, as expressed by the characteristics (concentration, grain size, mineralogy) of the particles suspended in, and subsequently deposited by, tidal waters.

The remaining forcing factors controlling the geomorphic operation of saltmarshes are internal to the marsh environment. The productivity of the halophytic plants growing on the marsh, especially the below-ground rate, (Reed, 1995) is a further factor influencing surface sedimentation. The fourth forcing factor results from sediment autocompaction which, *via* a variety of processes, reduces the volume of the stratigraphic column and effectively lowers the marsh surface relative to the tidal frame. Therefore, like upward sea-level movement, autocompaction also provides accommodation space.

From initiation, saltmarshes grow vertically and prograde laterally due to both organic (*in situ* and detrital) and clastic sedimentation. Surface sedimentation is linked to the relative elevation of the saltmarsh surface and tidal inundation in a powerful feedback loop driven by the forcing factors (Figure 2.4). Over time, a saltmarsh grows to higher elevations in the intertidal frame. By doing so it is flooded by progressively fewer tides for progressively shorter periods by tides of a particular height and so the rate of vertical growth theoretically decreases with time. This was illustrated empirically by Pethick (1981). Another important loop involves surface sedimentation, marsh relative elevation, and autocompaction as a critical forcing factor.

Coupled with the rate of vertical growth is a change in the character of the tidal creeks. As a marsh matures, creeks become deeper and higher in density and so the overmarsh tidal prism decreases. Figure 2.5 illustrates the development of a marsh from 'youth' (i.e. early geomorphological functioning following marsh inception) to 'maturity' in terms of various hydraulic and geomorphological variables.

Based upon the forcing factors and feedback relationships outlined above, Allen (1990) and French (1993) developed conceptual models, later calibrated using empirical data from the Severn Estuary and North Norfolk respectively, to explain the development of saltmarshes. Allen's (1990) basic time-stepping, one-dimensional model for the vertical growth of a saltmarsh has the form:

$$\Delta E = \Delta S_{\min} + \Delta S_{org} - \Delta M - \Delta P \qquad (1.2\ a)$$

where:

ΔE is the change in the level of the sediment surface relative to the tidal frame

ΔS_{min} is the added thickness of tidally imported mineralogenic sediment

ΔS_{org} the thickness of organogenic, sedimentary material grown *in situ* (mainly below-ground root biomass)

ΔM the change in relative mean sea level

ΔP the amount by which autocompaction lowers the surface.

The third and fourth terms combine to give the total rate at which accommodation space is made available, whereas the first and second terms combine to describe the total rate at which sediment is introduced (mineralogenic element) and produced (organogenic element) to fill it.

When ΔE = 0, a saltmarsh is in stable dynamic equilibrium and has reached elevational maturity. When this occurs, the rate of sedimentation balances the rate at which accommodation space is created by the combined effects of RSL rise and autocompaction, i.e.:

$$(\Delta S_{\min} + \Delta S_{org}) = (\Delta M + \Delta P) \qquad (1.2\ b)$$

Similarly, if

$$(\Delta S_{\min} + \Delta S_{org}) > (\Delta M + \Delta P) \qquad (1.2\ c)$$

the rate of sedimentation is greater than the rate of provision of accommodation space. Such conditions have been shown to occur when organogenic sedimentation rates are high (e.g. Craft *et al.*, 1993), allowing the marsh surface to rise out of the intertidal frame (Allen, 1990).

Providing either of these conditions is met, saltmarshes can maintain their position within the intertidal frame, as has often occurred throughout the late Holocene.

2.5.2 Empirical modelling of saltmarsh biogeomorphic processes

The conceptual models outlined above identify the basic controls on saltmarsh and mudflat operation and the relationships and feedbacks between these controls. Any numerical model capable of predicting the response of the relative elevation of intertidal landforms to forcing factors requires that each of the process terms in Allen's (1990) equation (Equation 1.2) are quantified and parameterised with empirical data from field investigations and laboratory experiments. The ultimate goal is then to incorporate the relative contributions of each process sub-model into the general relative elevation model (Rybczyk *et al.*, 1998).

A detailed understanding and quantification of processes is critical. The use and extrapolation of imperfectly or poorly understood processes beyond the limits of simple monitoring and observation results in large uncertainties in the final model (French and Spencer, 1993). Any such inaccuracies associated with individual processes will ultimately be amplified when implemented (De Vriend, 1987; GESAMP, 1991) (Figure 2.6).

The degree to which each of the terms in Allen's (1990) equation has been accurately quantified varies greatly. The mineralogenic term is perhaps the best understood due to previous development of general sedimentological theories in closely related disciples for environments that are morphologically and hydraulically analogous to saltmarshes and mudflats, such as river floodplains (Allen, 1990). Both Allen (1990) and French (1993) employed relatively well-constrained mineralogenic sedimentation sub-models to quantify the amount of mineral matter deposited on the saltmarsh or mudflat surface in relation to relative elevation and flooding frequency, expressed as a vertical accretion increment. These sub-models consider a variety of spatially and temporally dependent variables, such as the concentration of suspended sediment in tidal waters, the characteristic terminal settling velocities for the suspended mineral particles, and functions to represent the probability of particle resuspension/erosion.

The organogenic sedimentation term is less well quantified. Because organogenic sedimentation varies spatially and temporally in a complex manner in response to halophyte succession and changes in marsh nutrient status in response to tidal exchanges (French, 1993; Long and Mason, 1983; Silvestri *et al.*, 2005), and because the relative contributions of above- and below-ground components and their controls are highly variable, it could not be simply modelled and so was accordingly treated as a constant by both Allen (1990) and French (1993). However, given the mineralogenic nature of the systems under consideration, a detailed organic sedimentation sub-model was not a modelling priority.

Similarly, the autocompaction term was set to zero in both models, as it was not considered an active process. Prolonged exposure due to seasonal and neap-tide desiccation and dewatering in these upper intertidal deposits suggested to Allen (1990) and French (1993) that they were of a sufficient density and compressive strength to resist the pressures created by overburden sedimentation. As a result, the units were entered into the model in a 'precompacted' state.

The uncertainties in understanding individual processes limited the conceptual models discussed here to an exploratory, rather than predictive, capacity (a point acknowledged by Allen, 1990). Such models are only useful if predictions are made within the range of data used to establish the model (Burt, 1994) and if the data themselves are robust and firmly grounded in empirical, process-based observations. Without further parameterisation of processes, the models cannot be reliably employed in the management of intertidal areas, particularly in highly organic marshes where plant growth is the principle way in which accommodation space is filled and where autocompaction of the substrate is more significant (Cahoon *et al.*, 2003; Cahoon and Lynch, 1997; Cahoon *et al.*, 2000; Rybczyk *et al.*, 1998).

In addition to the lack of empirically-based process sub-models, models of upper intertidal geomorphology developed in the early 1990s failed to fully encapsulate the complexity of these landforms. Firstly, they only give a one-dimensional view of marsh response to forcing factors. A number of studies have illustrated the spatial and temporal variability of accretion over a marsh surface. French and Spencer (1993) determined the rate and pattern of vertical accretion within a large backbarrier marsh in Norfolk, UK by recording the progressive burial of artificial markers over a 5-year period. Spatial variations in vertical accretion varied from 8 mm yr^{-1} adjacent to larger channels to less than 1 mm yr^{-1} on the highest marsh surfaces, remote from the creek network. Reed (1988) presented data from the Dengie Peninsula, Essex, UK, which illustrate a landward, altitudinal decrease in sedimentation rate that is accompanied by a decline in accretion rate away from the larger creeks. Similar spatial trends were found by Kastler and Wiberg (1996) across a Virginia saltmarsh. Such data suggest that, in addition to the elevational control on the general pattern of accretion described in the basic Allen/French model, sedimentation patterns are also related to marsh surface flooding pathways and hence proximity to creek networks. Field measurements of flow velocities/vectors and suspended sediment concentrations have begun to provide explanations of the observed sedimentation rates and patterns in terms of saltmarsh and tidal creek hydrodynamics and tidal controls, such as tidal range (e.g. Davidson-Arnott *et al.*, 2002). Improvements can be made to the mineralogenic sedimentation sub-model so that they sufficiently incorporate spatial and temporal variability (Luternauer *et al.*, 1995) when these shorter term, process-based studies are combined with (a) longer-term radionuclide-derived sedimentation rates and (b) estimates of lateral translation of the areal extent of saltmarshes as determined from historical documents and long-term remote sensing datasets.

Reed (2002) reinforces the need to investigate the organic contribution to saltmarsh functioning. She provides a conceptual model of saltmarsh biogeomorphology that synthesises the organic and inorganic components and how they interact over a variety of timescales, particularly decadal, to control the long-term sustainability of intertidal wetlands – something that previous simulation models have failed to do. The Reed (2002) model is based upon the millennial-scale cyclic growth of the Mississippi Delta Plain and illustrates how the combined effect of spatially and temporally variable organic and inorganic accumulation processes (and the feedbacks between them and other components of the system, such as by varying the rates of autocompaction) adequately sustain coastal marshes during RSL rise. Furthermore, by considering the historical movement of the Mississippi River, Reed (2002) suggested how changing marsh accretionary processes were manifested spatially and in relation to coastal configuration and RSL change, increasing the dimensionality of the models.

Improvements to our understanding of the factors affecting saltmarsh functioning have clearly been ongoing and have allowed more detailed multi-parameter, multi-timescales models to be developed. Although dramatically improved, these models remain largely conceptual (e.g. Davidson-Arnott *et al.*, 2002, Figure 2.7 cf. Figure 2.4). Attempts to empirically and holistically model wetland behaviour still only focus on the simulation of sub-sets of processes that affect saltmarsh elevation change, and essentially ignore others (Rybczyk *et al.*, 1998). Whilst empirical models of some processes (such as mineralogenic deposition) increase in their sophistication and the accuracy of their predictive capacity, others still lag significantly behind.

Fagherazzi (2005) acknowledged many of the developments that have been made in integrating ecological and geomorphological processes. He identified the following key remaining research areas:

- the quantification of below-ground production of organic material;
- the relative role of vegetation and marsh morphology on tidal hydrodynamics and resultant deposition and erosion processes;
- the feedbacks between marsh morphology and vegetation biomass and zonation;
- the influence of biogeochemical processes on marsh ecosystems and morphology.

As with the majority of existing models of saltmarsh morphodynamics, the role of processes acting to provide accommodation space is not considered by Fagherazzi (2005), perhaps due to their long-term operation, the inaccessibility of the material of interest (i.e. sub-surface, often at considerable depth, and subtidal) and the relative difficulty of making direct observations and measurement. Nonetheless, focusing on saltmarsh accretion, rather than the mechanisms of autocompaction and future sea level trajectories, only provides a partial assessment of the long-term sustainability of these landforms and ecosystems.

The empirical findings of Cahoon *et al.* (1995) have yet to be developed into predictive models of change. An in-depth understanding of autocompaction is integral to a complete explanation of intertidal geomorphological functioning. Without developing predictive models of autocompaction, holistic models of saltmarsh behaviour are potentially meaningless. An understanding of autocompaction processes is now critical for the successful formulation of saltmarsh management strategies.

The simple extrapolation of current sea level trends into the future is inadequate because the processes governing sea level change, such as climate, will themselves vary. Furthermore, the IPCC-estimates of 'eustatic' sea level are of little use to those interested in the coastal zone as they do not account for other factors such as regional isostacy and local scale processes (Equation 1.1). The threat of future sea level rise to saltmarshes is site-specific and, in many cases, insufficiently quantified. Predictive models of future sea level change over a variety of spatial and temporal scales are required; the degree to which these have been produced is discussed in the following sections.

2.6 RELATIVE SEA LEVEL STUDIES

2.6.1 Scope of relative sea level studies

By using a range of geomorphological, archaeological, and bio-, chemo- and litho-stratigraphic sea level indicators in combination with dating techniques, it is possible to reconstruct historical vertical fluctuations in sea level and hence extend the contemporary instrumental (tide gauge) record back through time (Gehrels *et al.*, 2002).

A strategic selection of RSL histories from locations close to and at increasing distance from the centres of Late Quaternary ice sheet loading of the Earth's surface (i.e. 'near-' to 'intermediate-' and 'far-field' sites) presents the possibility of developing predictive models of RSL change. For example, constraints on the 'eustatic' function are provided by far-field sites (Clark *et al.*, 1978; Milne *et al.*, 2005; Peltier, 2002; Yokoyama *et al.*, 2000). These sea level records can then be compared with the high-resolution climate records obtained from ice sheets, ocean cores and terrestrial deposits. This helps to develop a quantitative understanding of the nature of climate-sea level relationships and shed light on any time lags, threshold effects and spatial discontinuities that may occur (Slaymaker and Spencer, 1998). It also provides baseline data against which the existence and potency of any anthropogenic impact on sea level change can be assessed. For instance, Donnelly *et al.* (2004) reconstructed relative sea level for the past 700 years in eastern Connecticut, USA, and detect a nearly three-fold increase in regional RSL rise in the late 19^{th} Century. This increase in rate coincides with the increased global climate warming observed in instrumental and proxy records and also with increased greenhouse gas emissions, tentatively suggesting an anthropogenic influence in sea level change. In a similar study, Gehrels *et al.* (2005) produced a precise ($\pm$ 0.055 m), high-resolution record of sea level change for the past 1000 years from foraminiferal and chronological (^{14}C, ^{206}Pb, ^{207}Pb, ^{210}Pb, ^{137}Cs and ^{241}Am) analyses of a 2 m thick high marsh peat sequence at Chezzetcook, Nova Scotia, Canada. Although radiocarbon dating errors and calibration uncertainties prevent rates of sea level change between 1500-1550 and 1700-1800 to be accurately resolved, Gehrels *et al.* (2005) reconstructed a long-term, pre-industrial revolution rate of 1.7 mm yr^{-1}. During the 19^{th} century, a similar mean rate of 1.6 mm yr^{-1} was recorded in the sediments. However, an acceleration to the contemporary mean rate of 3.2 mm yr^{-1} occurred between 1900 and 1920; once again, this strongly suggests an anthropogenic influence on climate and sea level change.

By comparing the 'eustatic' signal with the RSL signal at near-field locations, insights can be gained into the isostatic function of Equation 1.1 at a particular location. Shennan (1989) calculated late Holocene rate of isostatic uplift or subsidence by subtracting a model of 'eustatic' sea level from individual sea level index points and then calculating the best-fit linear trend to these modified sea level data. In a similar study, Shennan and Horton (2002) updated the 'isobase' map produced by Shennan (1989) by including new sea level index points and considering errors associated with tidal range change (Shennan *et al.*, 2000b) and autocompaction. However, rather than subtracting a 'eustatic' model from sea level index points, Shennan and Horton (2002) only considered those which are younger than 4000 cal. yr BP, since the 'eustatic' model is essentially zero over this period in Shennan's (1989) original analysis. This enabled the production of a UK map of late Holocene relative land and sea level changes (Figure 2.8). A spatial consistency is revealed; the highest relative uplift/sea level fall is centred around the area of Scotland with the maximum ice load at the time of the Last Glacial Maximum (Ballantyne *et al.*, 1998). In contrast, maximum relative sea level rise (land subsidence) has been occurring in southern and eastern England (i.e. in the peripheral forebulge collapse zone of the British and Fennoscandinavian ice sheets). By comparison with twentieth century tide gauge records (Woodworth *et al.*, 1999), Shennan and Horton (2002) illustrate an acceleration in the rate of mean sea level change in the order of 1 mm yr^{-1} as a result of steric effects and loss of terrestrial ice volume, potentially as a result of the anthropogenically-enhanced 'greenhouse effect'.

By considering the temporal and spatial variations in glacio- and hydro-isostatic adjustment throughout the Holocene and, in some cases, the Late Devensian/ Weichselian, a number of hypotheses regarding the geophysical properties of the Earth, such as mantle viscosity and rheology, can be tested. Data collected in this way is of particular use to researchers interested in modelling glacial isostatic adjustment (e.g. Lambeck, 1995; Peltier, 1995; Peltier *et al.*, 2002) and determining the dynamics of the mass balance of glaciers and ice sheets since the Last Glacial Maximum (LGM). When used in conjunction, geological records of former sea levels and geophysical models are highly effective tools in understanding the driving mechanisms of global environmental change (e.g. Clark *et al.*, 2002). Such synergy is dependent upon accurate and precise reconstructions of sea level, since geophysical models heavily rely on these reconstructions for calibration, refinement and validation of critical parameters.

Aside from ascertaining the nature of climate-sea level relationships, studies of sea levels and coastal evolution are also used in a variety of other applications (Pirazzoli, 1996). For example, through analysis of coastal sedimentary sequences on tectonically active coasts, an understanding of long-term seismic hazards can be developed. Atwater *et al.* (1991) dated the abrupt submergence of coastal trees into the intertidal zone to test the timing and magnitude of Holocene earthquakes in southern Washington, USA. Long and Shennan (1994) further tested aspects of the 'earthquake deformation cycle'; rather than focusing on the coseismic element, they used biostratigraphic techniques to investigate the interseismic strain accumulation aspect. From here stemmed higher resolution investigations into sediments that record the 'earthquake deformation cycle' which identified changes in the rate, magnitude and tendency of RSL change prior to major earthquakes (Hamilton and Shennan, 2005; Long and Shennan, 1998; Shennan *et al.*, 1999; Zong *et al.*, 2003), presenting the potential for predicting earthquake hazards in terms of timing, recurrence interval and magnitude.

Reconstructed Late Holocene RSL changes have also been used in conjunction with coastal archaeological deposits to help to understand historical human exploitation of coastal areas and, conversely, how evolution of the coastal landscape exerted a control on coastal communities. Behre (2004), for example, discussed how the habitation patterns of the Clay District of Lower Saxony, northern Germany were dependent upon fluctuations in RSL. Variations in the altitude of settlements, the style of buildings within them, and the timing of occupation and abandonment of different sites reflected various phases of RSL tendency. Roep and van Regteren Altena (1988) hypothesised that a RSL rise of approximately 0.75 m between 3275 BP and 2620 BP caused the desertion of the Bronze Age settlements of Bovenkarspel, northwestern Netherlands.

Whatever the final application of sea level data, it is imperative that sea level reconstructions accurately and precisely describe former rates and magnitudes of sea level change, and that individual, local records are of sufficient quality to allow regional correlation and any subsequent global 'up-scaling'. The following sections discuss some of the developments that have taken place to allow meaningful extraction of sea level records from coastal stratigraphies, and the remaining imperfections of existing sea level methodologies.

2.6.2 Sea level reconstruction: age-altitude analysis

Sea level indicators require an indicative meaning, which refers to the relationship of the local environment in which a sea level indicator formed/accumulated to a contemporaneous reference tide level (e.g. mean high water spring tide, mean tide level etc), with an associated error range – the indicative range, which refers to the vertical range in which the coastal sample can occur (Van De Plassche, 1986) (Figure 2.9).

When a sea level indicator can be dated and levelled to a reference datum, it can be employed as a sea level index point (SLIP). A SLIP is composed of the following components:

- a geographical location;
- an indicative meaning;
- an altitude, referenced to a stable geodetic datum (cf. elevation);
- an age;
- a tendency - a description of whether a sample displays an increase or decrease in marine influence (Morrison, 1976).

The creation of SLIPs allows the production of Holocene sea-level curves – graphs of altitude (y-axis) plotted against age (x-axis). In the United Kingdom, for example, altitudes are referenced relative to Ordnance Datum (OD), which approximates to mean sea level since it was determined from six years of continuous tide gauge records at Newlyn, Cornwall (May 1915 – April 1921).

Some sea level indicators are limited in their value in reconstructing sea level, having imprecise indicative meanings and large indicative ranges. For example, some corals, such as *Acropora palmata*, have large indicative ranges (c. ± 5 m), since the lower water depth limit of their growth is poorly constrained (Blanchon and Shaw, 1995). Such an indicative range may be equal to or larger than the RSL change under investigation and hence limits the value of the ensuing, poor-precision sea level reconstruction. Rock-clinging oyster beds and other such fixed biological indicators are often isolated in temporal and spatial occurrence, meaning that continuous records of sea level change are difficult to obtain. Morphological features such as palaeo-shoreline notches, palaeo-reef flats and beach deposits are sporadically distributed, are often difficult to accurately and precisely date, and may have large indicative ranges.

Some of these problems can be overcome through palaeoenvironmental investigations into Holocene sediments deposited in estuarine and other low energy coastal environments. Such locations may preserve long, continuous (providing erosion has been absent or minimal) sedimentary sequences of microfossil-rich clays, silts and peats (dateable by radiocarbon assay) since the time they accumulated. Such studies have yielded a great deal of information regarding the vertical accretion of biogenic and clastic sediments that can provide important Holocene analogues of coastal evolution and behaviour for future coastal changes (e.g. Long *et al.*, 2006; Shennan, 1995). When placed into a chronological framework and sensibly interpreted in the light of saltmarsh morphodynamics and regionally validated, intertidal sedimentary deposits can provide accurate and largely unbroken records of past vertical movements of RSL.

2.7 Errors in sea level reconstruction

Following International Geological Correlation Programme Project 61 (IGCP 61) on 'Holocene Sea Level Changes', there was a growing realisation of the potential for errors in sea level research. Analysis of the results in the final report of the IGCP 61 UK Working Group (Tooley, 1976; 1978) showed that sea level data were uneven both geographically and temporally. Also, the inadequacies displayed by some of the data were so numerous that great care had to be taken during subsequent analysis. The report stimulated research to overcome the geographical bias of sea-level index points and, more importantly, resulted in the recognition that 'the failure to employ a unified methodology...makes correlation of marine events at best elusive and at worst erroneous and misleading' (Tooley, 1985, page 113). Similarly, van de Plassche (1986) discussed the need for a more prolific, more detailed global distribution of sea-level data and graphs of better quality derived from organised, systematic and 'agreed-upon' methodological approaches.

In subsequent IGCP programmes (Projects 200, 274, 367 and 437 e.g. Shennan and Horton, 2002; Projects 200, 274, 367 and 437 e.g. Van De Plassche, 1986) a general homogenisation of methodology emerged. The nomenclature used in sea level studies was standardised with the introduction of defined terms, such as transgressive/regressive overlaps and tendencies of sea level (Morrison, 1976; Shennan, 1982; 1983), enabling more meaningful correlations between datasets and the development of non-deterministic multiple working hypotheses. From such logistical and organisational improvements stemmed a great deal of research that aimed to improve the quantification of historical sea level changes by developing techniques to reduce the errors associated with the estimation of the age and altitude of SLIPs. However, through technological and methodological advances, the effect of each source of error has been estimated and quantified and, in some circumstances, significantly reduced. The major sources of error, and examples of the ways in which these errors can be estimated or reduced, are discussed below.

2.7.1 Levelling the altitude of stratigraphic boundaries

(Shennan, 1980; 1982; 1986) identified three general sources of altitudinal error associated with levelling the altitude of stratigraphic boundaries:

1. measurement of depth within a borehole;
2. levelling of the study site to an Ordnance Survey (OS) benchmark;
3. the accuracy of the benchmark to Ordnance Datum Newlyn.

A summary of the individual errors which constitute these general error types and their approximate ranges is provided in Table 2.2. Although many of them are largely unavoidable, improvements to levelling equipment have reduced the error terms associated with levelling to benchmarks. Similarly, development in differential global positioning system (DGPS) technology has increased the accuracy and precision of estimates of longitudinal, latitudinal and altitudinal location (± 1 cm).

2.7.2 Problems in quantifying the indicative meaning and indicative range

Established methods of reconstructing sea level from saltmarsh deposits were developed by Shennan (1982) using contemporary data from the Fenland. This 'stratigraphic overlap' approach exploits the vertical zonation of vascular plants within the intertidal zone and assumes that intercalated sedimentary facies found in the stratigraphy formed in environments that exist in lateral juxtaposition. By relating elevation/flooding frequency-controlled lateral variations in sedimentary facies to a reference tide level, Shennan (1982) estimated the indicative meaning of commonly dated materials from intertidal areas. By quantifying the range of elevations across which the transition between one sub-environment and another occurs, an indicative range was assigned to a lithostratigraphic or biostratigraphic contact (Table 2.3).

This approach, although of considerable value and widely applied, was limited. Following experimental modelling of saltmarsh accretion, Allen (1995), for instance, suggested that a lag may exist between a change in RSL and its manifestation in a sedimentary sequence. The established Shennan (1982) method also relied upon the presence of organic saltmarsh deposits in the stratigraphic record for radiocarbon dating and so ignored the typically more abundant mineralogenic portion. Additionally, the stratigraphic overlap method is restricted to a small number of saltmarsh sub-environments. Although microfossil data could be used to identify the onset or removal of marine conditions and complement SLIPs developed from stratigraphic overlaps, such data could not be assigned an indicative meaning, and so a quantitative understanding of RSL changes between overlaps was not possible.

The development and implementation of 'transfer functions' has helped to overcome these problems. Transfer functions are multivariate regression models that use changes in floral and faunal assemblages (microfossils such as foraminifera, diatoms and, to a lesser extent, saltmarsh pollen) to generate quantitative palaeosea-level reconstructions based on the statistical relationships between contemporary ('training set') and fossil micro-flora and –fauna (Birks, 1995; Gehrels *et al.*, 2001; Horton and Edwards, 2006). Transfer functions are advantageous in the production of SLIPs due to the speed of response of microfossils to changes in flooding frequency, the increased range of environments that can be used, a removal of the over-reliance on organogenic stratigraphic overlaps, and a decrease in the indicative range (increased precision) (Horton, 1997; Horton and Edwards, 2006). Foraminiferal-based transfer functions, for example, have been used to reconstruct high-frequency (decadal to millennial), low amplitude (< 10m) changes of Holocene sea level to ± 5-10 cm (e.g. Horton, 1999; Horton *et al.*, 1999; Scott and Medioli, 1978; 1980; Varekamp *et al.*, 1992).

However, the transfer function approach is not without its assumptions and problems. For example, it is assumed that elevation remains the dominant control on foraminiferal distributions and that other environmental variables, such as salinity or pH, do not exert a strong or changeable influence through time (de Rijk, 1995; Horton, 1997; Horton, 1999). A second important assumption is that the composition of the contemporary foraminiferal ('training set') assemblages are representative of those found in the sub-surface sediments. Taphonomic and other post-depositional processes such as infaunal foraminiferal activity, destruction or dissolution of foraminifera tests, or transport and reworking of sediments may violate this assumption and introduce errors into reconstructions (Horton and Edwards, 2006).

2.7.3 Changes in tidal amplitude

Shennan *et al.* (2000b) pointed out that the majority of samples used in Holocene sea level reconstructions formed close to present day MHWST rather than mean sea level (MSL). However, as sea levels changed, sediments were deposited and the configuration of the coastline and offshore bathymetry also changed, meaning that significant changes in tidal range may occur. If this is the case, sea level records will differ from the 'true' mean sea level curve. Gehrels *et al.* (1995) stated that few studies of Holocene RSL change include numerical reconstructions of tidal range changes through time. Tidal range models that did exist (e.g. Austin, 1991; Gehrels *et al.*, 1995; Hinton, 1992) were limited in their application due to the relatively low number of tidal constituents used in analysis, unrealistic and low spatial resolution palaeogeographic reconstructions within major estuaries and coastal lowlands and poorly constrained spatial and temporal resolution of RSL changes (Shennan and Horton, 2002; Shennan *et al.*, 2000b).

The data gathered in the UK Land-Ocean Interaction Study (LOIS) (Shennan and Andrews, 2000) provided greatly improved, higher resolution information on the palaeobathymetric, palaeogeographic, sedimentatation and RSL histories of the Southern North Sea. When combined and analysed with greater computing power than previously possible, the number of tidal harmonics was increased from 6 to 26 to give more complete and accurate estimates of historical tide levels.

Model predictions of tidal range changes in the western North Sea during the Holocene predict an increase in tidal range, with major changes occurring before 6 ka BP (Shennan *et al.*, 2000b). Shennan and Horton (2002) discuss the implications of tidal range changes, stating that the historical rate of land subsidence would be overestimated without quantification through modelling.

2.7.4 Dating of sea level index points

The age of a SLIP, plotted on the x-axis of age-altitude graphs, has conventionally been calculated using radiocarbon dating. Because of this, estimated SLI ages have been prone to a number of unavoidable errors associated with the technique, such as variations in the atmospheric production of ^{14}C (e.g. De Vries, 1958; Suess, 1965; 1980) isotopic fractionation due to preferential absorbance by plants of ^{12}C relative to ^{14}C (Van De Plassche, 1986) and the marine reservoir effect (Reimer *et al.*, 2004). Improvements in field sampling, laboratory techniques and calibration procedures have allowed these problems to be overcome to some extent. Radiocarbon dates are still plotted within two standard deviations to account for precision of measurement of the radioactivity within a sample and the likelihood that this level of radioactivity represents a particular age.

Further technological developments have decreased the dependence on saltmarsh peats and plant macrofossils for radiocarbon dating. Accelerator Mass Spectrometer (AMS) techniques has allowed smaller samples to be radiocarbon dated (e.g. Tornqvist *et al.*, 1998). As a result, mineralogenic deposits containing calcareous foraminifera can be used to create both indicative meanings and supply material for AMS radiocarbon dating. Also, there has been research to assess the potential for luminescence dating of water-lain mineralogenic deposits (e.g. Clarke and Rendell, 2000; Plater and Poolton, 1992).

2.7.5 Sediment autocompaction

The majority of estuarine sediments employed as SLIPs have been subjected to post-depositional displacement in altitude due to the autocompaction of the underlying peats, clays, silts and sands. Unless corrected for with accurate 'decompaction' techniques, this lowering of SLIPs from their original elevations will result in an overestimate of the rate and magnitude of RSL rise (Shennan *et al.*, 2000a) (Figure 2.10).

The problem of sediment autocompaction was recognised in studies by Bloom (1964) and Kaye and Barghoorn (1964). Similarly, IGCP 61 formally drew attention to the problem in the late 1970s (Tooley, 1982; 1985), though the sea level research community continued to simply dismiss it as an inconvenient limitation. With the advent of high-resolution microfossil transfer functions (Gehrels, 2000; Gehrels *et al.*, 2001; Horton, 1997; Horton and Edwards, 2006), it became very important to recognise and clearly state the magnitude of autocompaction errors as they may be equal to or exceed trends in RSL.

Table 2.2 Errors affecting the measured altitude of stratigraphic boundaries based on data from the Fenland, UK. Source: Shennan (1982).

Identification of boundary	± 0.01 m
Measurement of depth – hand coring	± 0.01 m
Measurement of depth – commercial U4	± 0.05 m
Measurement of depth – commercial (disturbed)	± 0.25 m
Compaction and extrusion of piston cores	up to 0.06 m
Duits gouge sampler (not for ^{14}C samples)	up to – 0.20 m
Angle of borehole	up to 0.04 m
Levelling to benchmark	up to ± 0.02 m
Accuracy of benchmark to OD	± 0.15 m
Sampling density - 1 borehole per 2 m^2	± 0.06 m
Sampling density – 1 borehole per 5,400 m^2	± 0.14 m
95% limits =	± 0.30 m

Table 2.3 Indicative range and reference water level for commonly dated materials. Source: Shennan (Shennan, 1982).

Material	Indicative range	Reference water level
Phragmites or monocot peat:		
directly above saltmarsh deposit	20 cm	((MHWST + HAT)/2) – 20 cm
directly below saltmarsh deposit	20 cm	MHWST – 20 cm
directly above fen wood deposit	20 cm	MHWST – 10 cm
directly below fen wood deposit	20 cm	((MHWST + HAT)/2) – 10 cm
middle of layer	70 cm	infer from stratigraphy
Fen wood peat:		
directly above *Phragmites* or saltmarsh deposit	20 cm	(MHWST + HAT)/2
directly below *Phragmites* deposit	20 cm	MHWST
Basis peat	?80 cm	MTL to MHWST

Indeed, one of the aims of IGCP 367: 'Late Quaternary Coastal Records of Rapid Change' was to investigate changes in sedimentation rate and response, including the issue of sediment autocompaction (Haslett *et al.*, 1998). However, the response to this aspect of the project was wholly underwhelming and so, in the absence of a suitable technique for quantitative assessment, sea level researchers generally either ignore the problem or acknowledge it as an unquantified error term (Shennan and Horton, 2002).

2.7.6 Estimation of combined errors

As a result of the errors outlined above, it is clearly impossible to regard any sea level datum as a single point on a sea level curve. Even carefully analysed data can, at best, be plotted within error bars or as an error 'box' (Kidson, 1986). For age (x-axis) errors, Shennan and Horton (Shennan and Horton, 2002), for example, indicate the maximum and minimum ages, as defined by a radiocarbon date calibration package (e.g. CALIB 4.3;Stuiver *et al.*, 1998a; 1998b) using 95% (2 σ) confidence limits, with the vertical bar representing the median age. The altitudinal error (y-axis) is calculated *via* a root squared error of all the quantified or estimated height errors (e_1, $e_2 \ldots e_n$), including those associated with field levelling, benchmark uncertainty, present tide heights and interpretation of the indicative meaning (Shennan and Horton, 2002):

$$\text{Total vertical root squared error} = \sqrt{\left(e_1^2 + e_2^2 + \ldots e_n^2\right)} \quad (2.3)$$

2.8 AUTOCOMPACTION: THE PERVASIVE ERROR

Unlike other potential sources of error, the problem of sediment autocompaction within sea level research has not been fully resolved and remains a pervasive, prohibiting the successful interpretation of sea level data derived from intertidal sedimentary deposits.

Shennan and Horton (2002) stated that autocompaction of deposits with a high sand fraction is very low whilst autocompaction of peat may be as high as 90% by volume. Such figures, however, solely describe maximum autocompaction potential – extreme states unlikely to be attained in coastal situations over a geologically short period of several thousand years (Greensmith and Tucker, 1986) and hence offer no way of decompacting Holocene RSL curves.

The effects of autocompaction-induced volumetric reduction are also clearly illustrated by the distortion and displacement of coastal stratigraphies consisting of intercalated beds of clays, silts, sands, and peats. Figure 2.11 illustrates the contemporary stratigraphic architecture at Woltzeten in the Ems marshes, northwest Germany (Streif, 1971, cited in Allen, 2000). On the local spatial scale (10^1 – 10^2 m), the upper peat-silt/clay transgressive contact can be seen to be generally isochronous and so locally, diachroneity cannot completely explain the extent to which beds now vary in altitude. At this scale, the beds (other than palaeochannels) can therefore be assumed to have been effectively horizontal at the time of deposition. The present stratigraphic geometry, attained after significant autocompaction, is a distortion of the depositional form. Other than in areas where a basal stratum directly overlies the stable, compressible basement, all overlying material has been displaced downward from its approximately planar depositional position.

Some studies use basal peat deposits to quantify the magnitude of sediment autocompaction. Basal peats overly incompressible bedrock or Pleistocene boulder clay surfaces, and so cannot be lowered from their depositional altitudes. By directly comparing the altitudes of compacted and hence lowered intercalated peat strata with isochronous basal peats from the same stratigraphic sequences, the magnitude of autocompaction can be calculated. Haslett *et al.* (1998), for example, document the variable altitude of a peat-clay contact within the Somerset levels, southwest Britain. The maximum altitude of the contact is where the peat overlies a bedrock high; this is taken to be the minimum pre-compaction altitude (4.64 m OD). The lowest recorded altitude of the transgressive overlap was 2.42 m OD and so the maximum observed magnitude of autocompaction was taken to be 2.22 m.

Long *et al.* (2006) employed detailed bio-, litho- and chrono-stratigraphic evidence to quantify the degree to which autocompaction lowered a peat surface during the late Holocene in Romney Marsh, southeast England, UK. Microfossil evidence documents a gradual inundation of coastal wetland after 1263 – 1085 cal. yr BP and the establishment of a saltmarsh. Subsequently, a rapid increase in water depths is recorded in the litho- and bio-stratigraphy. This is followed by the rapid deposition of 4.8 metres of laminated intertidal mudflat and tidal channel sediments in a few hundred years. On the basis of existing knowledge of late Holocene 'eustatic' trends and isostatic adjustments (Waller and Long, 2002), such a change in water depths cannot be attributed to absolute sea level rise. Therefore, Long *et al.* (2006) attribute these rapid and catastrophic changes in palaeoenvironment to the autocompaction of the underlying peat. According to radiocarbon chronology and historical records, the accommodation space provided by the autocompaction of the peat was infilled within a few hundred years, indicating the staggering rate of catastrophic coastal change.

Based upon the data on the magnitude of autocompaction provided in these studies, Figure 2.12 illustrates how the process is the largest potential source of error in sea level reconstructions, greatly exceeding each of the other sources outlined in Section 2.7.

2.8.1 Basal peats

The issue of autocompaction clearly requires prompt attention. It can be partially dealt with by selecting sub-sets of SLIPs that are less prone to the autocompaction error, such as basal peats (Jelgersma, 1961; Shennan and Horton, 2002). Following analysis and interpretation of the indicative meaning, dating of these autocompaction-resistant basal deposits provides a compaction-free chronology of sea level change, since they are unable to move downwards through the stratigraphic profile. This can be seen on an age-altitude graph; SLIPs from intercalated strata plot below same-age basal index points (Shennan and Horton, 2002).

Törnqvist *et al.* (2004), for example, successfully used basal peats obtained from an extensive sampling program in the Mississippi Delta, Louisiana, USA to solve a long-running debate regarding the nature of Holocene relative sea level rise in the Gulf Coast. Prior to their study, mutually conflicting RSL curves existed for the region, with some suggesting a monotonic rise, others a 'stepped' pattern of alternating stillstands and rapid metre-scale rises and even a Holocene high stand of 2 m above present mean sea level. The basal peat reconstructions of Törnqvist *et al.* (2004) presented compaction-free evidence that a relatively smooth rising trend had occurred between 8000 – 3000 cal. yr BP. They concluded that the area has been responding glacio-isostatically to the forebulge collapse associated with the melting of the Laurentide Ice Sheet.

Nevertheless, basal peat derived sea level reconstructions are not without criticism. Interpretations of the indicative meanings of samples from the bases of basal peats are subject to debate. Jelgersma (1961) considered the process of basal peat formation to be a result of distance to shoreline, tidal range and permeability of the underlying subsoil. To correctly interpret the indicative meaning of a basal peat sample, it is necessary to know whether sea level or local groundwater level changes initiated their formation. When analyzing sea-level data collected by Jelgersma (1961) from the SW Netherlands, Kiden (1995), for example, identified that the basal data showed an anomalously high age/altitude position relative to sea-level curves for other areas in the Netherlands. His conclusion was that this was due to early peat growth above contemporaneous MSL resulting from groundwater-gradient effects on the gently inclined Pleistocene subsoil (Kiden, 1995). Van de Plassche (1991) had earlier considered this problem and concluded that basal peat samples should only be employed in sea-level reconstructions following a detailed study of the relief of underlying Pleistocene deposits. Ideal conditions for basal peat sampling therefore occur where there is sufficient slope in the underlying Pleistocene deposits to altogether avoid the groundwater-gradient effect.

Further issues exist regarding the use of basal peats in sea level reconstructions. Although potentially independent of any autocompaction-induced altitudinal inaccuracies, basal peats only allow the estimation of a long-term trend in sea level change, and cannot provide insights into higher resolution, short-term fluctuations superimposed on the trend. Conversely, the transfer function approach that uses both basal and intercalated peats produces a high resolution record but is highly susceptible to autocompaction height errors. Furthermore, basal peats cannot identify falls in RSL. Most importantly, only the extreme base of a basal peat is autocompaction free and hence basal peats thicker than a few centimetres will undoubtedly have experienced significant autocompaction.

Gehrels (1999) produced a RSL history for Machiasport, Maine, USA spanning the past 6,000 years through the ^{14}C dating of basal peats. This provided the long-term trend of sea-level rise. Through detailed biostratigraphic (foraminiferal) analysis of the saltmarsh stratigraphy, it became possible to reconstruct changes in the height of the marsh surface relative to sea level. By a direct comparison with basal peats from the same stratigraphic sequences, the intercalated SLIPs were corrected for autocompaction to their original altitudes.

This technique should only be regarded as approximate. The sedimentary sequences involved here are simple in comparison to the Holocene saltmarsh sequences of northwest Europe, particularly at mid-Holocene levels (*c.* 6.5-2.5 Ka BP), where peat and silt are often vertically stacked in distinctive sediment couplets (Allen, 1997). Furthermore, such geometric correction techniques are also limited by the availability of borehole and chronological data. Van de Plassche (1991) concluded that diachroneity as well as sediment autocompaction determines the altitudes of stratigraphic boundaries. Hence, despite attempts to rule out diachroneity as a factor creating an altitudinally-variable transgressive contact, the borehole density may not be sufficient to disregard it as an important factor. Also, decompaction to a basal datum is assumed to be linear in rate and magnitude and provides no information on the differential autocompaction of the underlying strata.

If sea-level research were limited to using autocompaction-free basal peats with sufficiently sloping underlying strata, the presently available database would be severely limited and, by ignoring 30 years of data collection and improvements in sea level research methodology, would lack sufficient accuracy, precision and resolution to detect decimetre-scale fluctuations in RSL.

2.9 Conclusions

This chapter has demonstrated the importance of autocompaction as an active geomorphological process in low energy upper intertidal environments. Conceptual models of the functioning of these landforms, and the numerical experiments which have been undertaken using these models (e.g. Allen, 1990; French, 1993), have illustrated the integral role of autocompaction in intertidal geomorphological evolution. Indeed, empirical contemporary (e.g. Cahoon *et al.*, 1995) and stratigraphic (Long *et al.*, 2006) data have displayed the importance of autocompaction in driving relative elevation change in intertidal environments. Two main areas of research exist that would benefit enormously from an understanding of autocompaction processes:

1. Saltmarshes provide significant direct and indirect benefits to society, particularly as inexpensive 'soft' engineering solutions to high energy storm wave attack, as sinks for pollutants and as important wildlife areas. Rising sea level threatens to permanently submerge these environments if they cannot accrete at a rate equal to or greater than that at which relative sea level is rising. In order to inform policy and management decisions, numerical modelling offers a means of gaining insight into the future behaviour of these systems under predicted scenarios of sea level rise. However, such models will lose significant predictive capacity if autocompaction of the substrate continues to be ignored. Estimations of autocompaction based on field data offer no long-term predictive ability and so attempts to forecast rates and magnitudes are entirely speculative.

2. Autocompaction confounds attempts to quantify past sea level changes obtained from intertidal deposits. Autocompaction lowers sea level index points from the altitudes at which they were deposited. This results in an overestimation in the rate and magnitude of inferred relative sea level change. Hence, any subsequent practical application of sea level data, such as the analysis of causal relationships between climate and sea level, the detection of anthropogenic impacts on the rate of sea level rise and calibration and refinement of geophysical models, is potentially error prone and highly misleading. Although analysis of stratigraphic sections allows autocompaction to be crudely quantified in terms of surface lowering, this commonly only suggests the minimum amount of volumetric reduction that could have occurred. Furthermore, it offers no insight into rates of autocompaction between the observed boundary conditions.

In order to address these issues, pre- and retro-dictive models of autocompaction are required to either feed into numerical models of saltmarsh elevation change or to correct sea level index points to their depositional altitudes. A pre-requisite for such models is a detailed understanding of the mechanisms by which intertidal sediments compact and deform. Such data are, however, minimal and consequently empirically-informed process models are similarly sparse. Existing models, their benefits and problems, are discussed in Chapter 3.

CHAPTER 3: MODELLING AUTOCOMPACTION IN INTERTIDAL ENVIRONMENTS

3.1 AUTOCOMPACTION: GENERAL DEFINITION

The volumetric deformation of earth materials, and the associated surface subsidence, is of relevance to a number of disciplines. For example, civil engineers are interested in autocompaction behaviour in the context of foundation design and the minimisation of settlement and associated damage to structures. Autocompaction is also of interest to structural and petroleum geologists to determine the evolution of porosity and permeability during burial and the development of growth faults and normal faults, which in turn may assist in the analysis of hydrocarbon migration and location.

The varied application of autocompaction analysis has resulted in a confusing nomenclature, where sub-sets of autocompaction processes are in some cases assigned different yet synonymous names, and in others are given the same name with contrasting meaning. For example, in civil engineering, 'consolidation' refers to the time-dependent expulsion of water from sediment interstices and the consequent reduction in volume as a result of, for instance, loading by superincumbent material (e.g. Lambe and Whitman, 1979). Some geologists use the term 'gravitational compaction' in the sense that engineers use 'consolidation' (Gill and Lang, 1977; Skempton, 1970). Conversely, for engineers, 'compaction' means mechanically reducing void spaces by using machinery when attempting to control subsidence prior to construction (Powrie, 2004). The definition of Skempton (1970) is lithology specific, defining 'consolidation' as the combined result of all processes, including particle bonding, desiccation, cementation and expulsion of pore water (i.e. 'consolidation' using the engineering definition), that transform an argillaceous sediment to a soft mud and through to a mudstone or shale.

Allen (2000b) defines autocompaction as the 'group of interlinked processes whereby the sediment within a growing stratigraphic column diminishes in volume, on account of burial and self-weight, leading to a rearrangement of the mineral skeleton, and in the case of vegetable matter, a loss of mass as the result of biological and chemical decay' (Allen, 2000b, page 1186). Allen (2000b) therefore views 'autocompaction' as an holistic, 'catch-all' term describing the full range of (bio-)mechanical and (bio-)chemical processes that lead to a volumetric reduction of sediments. The definition is applicable to all lithologies, whether clastic or organic.

Allen's (2000b) definition is favoured in this thesis to describe both the range of processes that act to reduce the volume of (i.e. to compress) a sedimentary column. However, individual autocompaction processes are discussed throughout this thesis and these are also defined in this chapter.

3.2 MODELLING AUTOCOMPACTION IN INTERTIDAL ENVIRONMENTS: THE BASIC PROBLEM

Before the relevance and predictive capacity of existing models of autocompaction can be assessed, it is pertinent to consider the system that requires modelling (Figure 3.1, modified from Audet, 1995).

A layer of autocompacting sediments overlies a basement of incompressible rocks or sediments. The top of the sediment layer is the depositional surface. A height co-ordinate, z, can be defined where $z = 0$ is the depositional surface and $z = h$ is the basement, where h is the total thickness of the sediment column (which may consist of several lithological units, each of thickness T, which may vary depending on the lithological characteristics and geotechnical properties of each layer – Tovey and Paul, 2002). Over time, sedimentation occurs, where ΔS_{min} is the added thickness per unit time of largely mineralogenic tidally imported detrital sediment to the growing column of sediment, and ΔS_{org} is the added thickness per unit time provided by *in situ* growth of organic material. Gravitational forces, g, act in the z direction. Acting within the sediment column is P, the full range of autocompaction processes that will reduce the volume of the accumulating sediment. At different times and depths within the sediment, and depending on lithology, different combinations of autocompaction processes will act on the sediments to reduce their volume. ΔP describes the volumetric reduction that results from the operation of P.

The main goal of a (de)compaction model is to realistically describe the volumetric behaviour observed in the field and in laboratory experiments in response to controlling autocompaction processes, such as the rate of biological or chemical decay or application of overburden stress. A stratigraphic sequence is typically divided into several layers on the basis of homogenous lithology (e.g. Allen, 1999) or, for example, geotechnical properties (e.g. Tovey and Paul, 2002). A new layer of sediment is added to the sediment column at the depositional surface over time. This layer exerts a compressive stress on the underlying material. During the time taken for sedimentation of this layer to occur, time-dependent diagenetic processes such as chemical alteration and biological decomposition also take place to a degree determined by a number of factors such as the sedimentation rate, the nature of the sedimentary/diagenetic environment and the range of lithologies present in the stratigraphic column (Gutierrez and Wangen, 2005). These processes reduce the thickness of a given layer (Figure 3.2). The resultant cumulative volumetric loss of each layer reduces the initial total thickness of a sedimentary column, h_1, to a new, model-generated total thickness, h_2 (Figure 3.2). This procedure is required in a saltmarsh management context where there is a need to predict surface elevation change in relation to rising relative sea levels.

In contrast, sea level index points that form at the depositional surface will be lowered by autocompaction from their depositional altitudes, affecting calculated rates and magnitudes of relative sea level change inferred from sea level curves. Here, the autocompacted thickness of the sedimentary column, h_2, is known and the decompacted thickness, h_1, at the time of deposition, t_1, is the required output of a decompaction model (Figure 3.3). This is done by considering the *in situ,* compacted thicknesses of lithologically and geotechnically uniform layers individually. The uppermost (most recently deposited) layer is removed ('backstripped') and the previous autocompaction conditions (e.g. stress state, degree of biological decay etc.) that existed in each layer prior to the deposition of the uppermost layer are calculated. Using an autocompaction model, individual layer reductions that took place can be ascertained. By summing these individual layer height corrections and adding them to the *in situ* thickness of the sediment column, the total decompacted height, h_1, at the time of deposition, t_1, can be calculated (e.g. Paul and Barras, 1998).

In both cases, without an empirically-derived conceptual understanding and mathematical description of autocompaction processes, the rate and magnitude of the autocompaction trajectory are unknown and so quantitative assessments are speculative at best.

3.3 Geotechnical modelling of mechanical autocompaction processes

Sub-sets of autocompaction processes have been wholly or partly addressed. Of particular relevance is the consideration of the mechanical behaviour of sediments following loading by overburden materials provided by soil mechanics.

Clastic and biogenic sedimentation taking place at the depositional surface is of bulk density, ρ_b. When multiplied by the gravitational constant (g = 9.81 m/s^2) to calculate the unit weight of the sediment, denoted by γ and measured in kN/m^3, vertical total stresses in the ground can be calculated. A sediment element at depth, z, below the ground surface is subjected to a vertical total stress (σ, measured in kPa) created by the overlying column of sediment:

$$\sigma = \gamma z \tag{3.1}$$

Following the application of a vertical total stress, sediments respond with a volumetric strain. However, this strain does not take place immediately due to the presence of water in the voids between the soil particles. This water also exerts a pore water pressure, denoted by u. When a saturated soil is subjected to an increase in total stress, the entire load is initially carried by the additional induced pore water pressure, the excess pore water pressure (denoted by u_e). At this initial stage, the pore water pressure is equal to the applied total stress. This results in a non-equilibrium pore water pressure gradient and so, if the clay/silt soil is over- and/or under-lain by permeable strata, the excess pore water pressure will result in an outflow of water from the clay/silt soil into the surrounding layers (Powrie, 2004; Head, 1988). As the water drains, the total stress is gradually transferred to the soil particles and the pore water pressure correspondingly falls.

The difference between the total stress and the pore water pressure is called the 'effective stress' (denoted by σ') and is equal to the stress carried by the soil skeleton (Terzaghi, 1936):

$$\sigma' = \sigma - u \tag{3.2}$$

This time-dependent process whereby soil particles are packed increasingly closely together in response to the application of continued pressure is called primary consolidation (Figure 3.4) (Powrie, 2004). In high permeability soils such as sands and gravels, consolidation is rapid and even immediate. However, in clay- and silt-rich soils, volumetric reduction is much slower due to the low permeability of the soil fabric and the associated inhibiting effect on the dissipation of excess pore water pressures. Gibson (1958) suggested that for most natural sediments, the rate of deposition operates at a sufficiently slow rate to allow dissipation of excess pore water pressures during an annual increment of sedimentation. Therefore, sediments may be almost fully consolidated at any stage throughout the deposition process. It is therefore often assumed that compression and loading effectively operate synchronously and so some authors do not consider the time dependence of consolidation in natural sediments (e.g. Massey *et al.*, 2006; Paul and

Barras, 1998; Tovey and Paul, 2002). Instead, they use simple rheological models to describe stress-strain behaviour.

The volumetric state of a soil is often described in soil mechanics by its voids ratio, e, which is defined as the ratio of the volume of voids (V_v) to the volume of solids (V_s), $e = V_v/V_s$ (Powrie, 2004). When plotted against vertical effective stress, the one-dimensional compression behaviour of a soil can be ascertained. If an accurate regression line can be fitted to these data, the volumetric state of sediments can be predicted for any vertical effective stress. Such rheological models can be used to estimate the magnitude of consolidation-induced settlement that will occur in field-based problems (Das, 1998; Head, 1988; Powrie, 2004).

Various mathematical functions have been used to describe autocompaction behaviour (Boudreau and Bennett, 1999). Allen (1999; 2000a) described the effects of autocompaction with the following asymptotic, exponential decay function:

$$T = (T_0 - T_{\min}) - \exp(kH) + T_{\min} \tag{3.3}$$

where:

T = the final *in situ* thickness of a sediment layer
T_0 = the thickness at the time of deposition
T_{min} = the limiting thickness (theoretical 'zero' porosity)
k (m^{-1}) = an empirical coefficient describing the compressibility of the layer
H = the overburden thickness (assumed in the particular environment to fill the accommodation space to the height determined by sea level at the time).

Equation 3.3 suggests that the effects of autocompaction will be critically dependent upon:

1. the thickness of the compacted sequence above the 'basement', which is assumed to be stable and incompressible;
2. the range of lithologies present (from gravel/sand, silt to peat, in order of increasing compressibility);
3. the vertical order in which the lithologies were deposited.

Allen (1999; 2000a) applied this model in a simplified, semi-empirical way in order to develop a wider understanding of how autocompaction lowers depositional surfaces/marker horizons from their original, depositional elevations over time. From a number of numerical experiments, Allen (1999; 2000a) presented the results as vertical lithological sequences built up using Equation 3.3 and empirical estimates of k obtained from field studies undertaken by Haslett *et al.* (1998), DeLaune *et al.* (1983), Bell (1995) and Hawkins *et al.* (1989). All of Allen's (1999; 2000a) simulations demonstrate that the synthetic beds experience a progressive stratigraphic displacement and a loss of depositional age-altitude relationships. Encouragingly, the model is able to replicate Holocene stratigraphic cross-sections found in northwest Europe. For example, the burial of a bedrock 'hill' by silts and peats results in differential autocompaction which mirrors the stratigraphic geometry observed by Haslett *et al.* (1998) in the Axe Valley, Somerset Levels, southwest England.

Allen's (1999; 2000a) results are correct in general form but not in detail because the model is only partially calibrated with empirical data. Despite its semi-empirical nature, it clearly illustrates the potential that autocompaction modelling has for predicting landscape change and altitudinal reconstructions. It will only be possible to improve upon the exploratory models of Allen (1999; 2000a) if empirically derived, accurate and precise models of autocompaction behaviour can be established on the basis of field observation and laboratory experimentation.

3.3.1 Empirical observation of mechanical autocompaction behaviour

The one-dimensional vertical strain (compression) behaviour of a soil can be investigated in the laboratory using an oedometer (Figure 3.5). The oedometer test is the classic laboratory experiment used for the determination of the consolidation characteristics of soils, conventionally clay soils of low permeability. It provides important information regarding the compressibility of soils, which is a measure of the amount by which the soil will compress when loaded and allowed to consolidate (Head, 1988). The test is carried out by applying a sequence of increasing vertical loads to a laterally confined specimen having a height approximately one quarter of its diameter (typically 20 mm high with a diameter of 75 mm) (Figure 3.5). The vertical compression under each load is observed over a period of time, which is usually constant for each load increment (e.g. 24 hours) but can be varied depending on the soil and the nature of the test (Head, 1988).

When the observed settlement is a result of primary consolidation only, and when the soil is being compressed for the first time, the data points lie on a unique straight line – the one-dimensional normal (virgin) compression line - when voids ratio, e, is plotted against the common logarithm of vertical effective stress, $\log_{10}\sigma'$ (Bjerrum, 1967; Powrie, 2004).

The slope of this $e\log_{10}\sigma'$ is known as the compression index, C_c, and is equal to the change in voids ratio for one $\log_{10}$ cycle of pressure (Powrie, 2004). Values of C_c are material specific. A soil which is on the normal compression line has never before been subjected to a vertical effective stress greater than the current value and is described as normally consolidated (Figure 3.6). The one-dimensional normal (virgin) compression line is represented by the equation:

$$e = e_1 - C_c \log_{10} \sigma' \tag{3.4}$$

where:

e = the voids ratio at a given vertical effective stress, σ'
e_1 = the voids ratio at an effective stress of 1 kPa.
C_c = the compression index
σ' = vertical effective stress.

This is termed Terzaghi's compression law (Massey *et al.*, 2006; Paul and Barras, 1998). The compression index can be calculated from the following equation (Selby, 1993):

$$C_c = \frac{e_1 - e_2}{\log_{10}(\sigma'_1) - \log_{10}(\sigma'_2)} = \frac{\Delta e}{\Delta(\log_{10} \sigma')} \tag{3.5}$$

3.3.2 Overconsolidation

If a soil has been subjected to a vertical effective stress greater than the present overburden it is described as being overconsolidated. This can occur following the erosion of deposits that overlie a particular soil stratum, or following the removal of glaciers during deglaciation (Head, 1988). The maximum previous effective stress to which the soil has been subjected is known as the preconsolidation stress, σ'_c. The ratio of the preconsolidation stress to the existing effective overburden stress, σ'_o, is known as the overconsolidation ratio (OCR):

$$\text{OCR} = \frac{\sigma'_c}{\sigma'_o} \tag{3.6}$$

The OCR is a simple but important indicator of the existing stress state in a soil in relation to its previous loading history (Powrie, 2004). Overconsolidated soils are much stiffer (resistant to the effects of an effective stress increase) than normally consolidated deposits. When the effective stress acting on normally deposited soil is decreased, the soil will not regain the full volumetric loss that was experienced during the corresponding, identical previous effective stress increment. Instead, the soil will follow a different stress path of a reduced gradient (Figure 3.7). Unless the preconsolidation stress is reached, when plastic, virgin consolidation results again, the voids ratio (and hence volume/layer thickness) will follow an elastic unload-reload hysteresis loop (Figure 3.7).

Overconsolidation also leads to higher densities (lower voids ratios) and higher shear strengths than theoretical values calculated for the existing overburden (Greensmith and Tucker, 1971a). Overconsolidation can also result from desiccation during prolonged exposure; physico-chemical changes also then occur within the sediment (Greensmith and Tucker, 1986; Greensmith and Tucker, 1971a).

3.3.3 Sedimentation compression curves

Skempton (1970) provided data on the *in situ* consolidation of natural clays through geological time. By taking undisturbed sediments samples from known depths at a variety of locations and calculating γ, σ' and e, Skempton (1970) was able to create 'sedimentation compression curves' by plotting e against $\log_{10}\sigma'$. These sedimentation compression curves illustrate the progressive changes in e from recently deposited muds located approximately 10 cm below the sea bed to Quaternary clays at the depths of several metres to hard clays and mudstones of Pliocene and late Pleistocene age at depths of 3 kilometres. Skempton's (1970) findings are displayed in simplified form in Figure 3.8. At any given value of σ', the voids ratio of a particular argillaceous material depends on the nature and amount of the clay minerals present, as reflected in the liquid limit of the sample. This refers to the moisture content above which the clay will behave as a liquid. The higher the liquid limit, the higher the voids ratio. A further and rather striking trend is the converging pattern formed by the compression curves (Burland, 1990). Also, as observed in oedometer tests, increases in $\log_{10}\sigma'$ result in an essentially linear decrease in e – i.e. natural cohesive clastic sediments conform to Terzaghi's compression law. The linear nature of the sedimentation compression curves suggests that the sediments are normally consolidated, and this was confirmed by oedometer laboratory testing by Skempton (1970). It is therefore reasonable to assume that normally consolidated $e\log_{10}\sigma'$ curves developed in the laboratory are good approximations of the volumetric evolution of natural sediments with respect to geological effective overburden stresses (Paul and Barras, 1998) providing that the co-ordinates of the $e\log_{10}\sigma'$ curve represent the end of primary ('EOP') consolidation phase of each load increment only (Mesri and Ali, 1999).

3.4 The use of Terzaghi's compression law in models of autocompaction in intertidal and marine environments

The empirical laboratory and field trends outlined in Section 3.3 above suggests that the exponential decay function employed by Allen (1999; 2000a) may be wrong. Other models of intertidal autocompaction have instead used a seemingly more realistic logarithmic decay function (i.e. Terzaghi's compression law, cf. Burland, 1990; Skempton, 1970). These models are relatively easy to implement, since they require only minimal data on the physical and material properties of the sediment, namely a value for the compression index and a reference voids ratio. This is usually the intercept value of Equation 3.4, e_1 – the voids ratio at 1 kPa σ' (Tovey and Paul, 2002).

In order to apply corrections for altitudinal displacement in the Nar Valley in the Fenlands (Norfolk, UK), Smith (1985) estimated amounts of compression in various peat and clay facies using the oedometer test. Since marine clays are deposited as a slurry at the water/mud interface, and because peats form at a waterlogged surface, Smith (1985) assumed an original (depositional) overburden effective stress of zero. He therefore was able to provide an approximation of original layer thicknesses by extrapolating the normal compression line obtained in the oedometer test back to the intercept (an effective stress of 1 kPa) on a plot of percentage height change against $\log_{10}$ vertical effective stress (Figure 3.9). This procedure yields a negative percentage value on the *y*-axis which represents a percentage layer thickness increase from the present day layer thickness. When this value (made positive) is multiplied by the thickness of the corresponding lithological unit in the field, the original height at 1 kPa (at the depositional surface) is obtained. When undertaken for each sedimentary unit and summed, the sequence can be fully decompacted. Despite reporting them, Smith (1985) did not make use of the values of the compression index, C_c, and so 'continuous' prediction of height changes in response to effective overburden stress was not achieved. As a result, lithological units were reported as fully compacted or fully decompacted with no mention of intermediate conditions between these states, and so the value of the oedometer test was not fully utilised.

Paul and Barras (1998) did make use of the compression index and were able to replicate the 'complete' compressive stress-strain behaviour of the sediments according to Terzaghi's compression law (Equation 3.4). The values for the compression index are obtained from a correlation, initially identified by Skempton (1944), with the experimentally-obtained liquid limit (LL) of the sediment:

$$C_c = 0.009(LL - 10\%) \tag{3.7}$$

However, the liquid limit test should only be determined for generally fine-grained (<425 μm) soils. Similarly, the correlation of the liquid limit with the compression index is only applicable to fine-grained plastic soils. Nonetheless, the correlation can be invaluable due to the difficulty in obtaining an undisturbed sample from depth for geotechnical testing and direct measurement of the compression index.

Tovey and Paul (2002) used the same rheological model (Terzaghi's compression law) but in this case defined a forward-iterative model. Rather than starting with the compacted profile as the starting point, as was the case in the models of Smith (1985) and Paul and Barras (1998), Tovey and Paul (2002) started by analysing the consolidation process in the sediment layer that is deposited first, and tracked the changing voids ratio profile in the sediment column as each new increment of sediment is added at the depositional surface. Thus, Tovey and Paul (2002) modelled the early consolidation process, but on the assumption that Terzaghi's compression law applies to very thin layers (≤ 1 mm).

Tovey and Paul (2002) also provided a means of estimating the voids ratio at an effective stress of 1 kPa. This allows the gradient of the compression index to be 'fixed' to a given intercept value and permits prediction of e using Equation 3.4 if the effective stress is known. Tovey and Paul (2002) pointed out that geotechnical investigations rarely report a value of e_1 determined from the backward extrapolation of the virgin compression line to 1 kPa. By plotting experimentally-derived values of e_1 against C_c, the following relationship (r^2 = 0.971) is obtained:

$$e_1 = 0.8154 + 2.8473C_c \tag{3.8}$$

Although there is no obvious scientific rationale for this relationship, also identified by Skempton (1970) and Burland (1990) and itself requiring investigation (Tovey and Paul, 2002), it means that only a value of C_c needs to be (directly or indirectly) empirically obtained for use in (de)compaction algorithms.

3.5 TIME-DEPENDENCY OF PRIMARY CONSOLIDATION

Contrary to the aforementioned findings of Gibson (1958), Tovey and Paul (2002) suggested that excess pore pressures may develop in sequences thicker than approximately 2 metres. Thus it may be necessary to consider the time-dependency of primary consolidation in models of autocompaction/consolidation.

Over conventional civil engineering timescales (decades to centuries) and stresses (e.g. 10 kPa – 1 MPa), primary consolidation is the dominant settlement process in cohesive clastic sediments (clays and silts) (Head, 1988) and it has hence received considerable analytical attention. Terzaghi (1943) proposed a mathematical approach to the consolidation process in lithologically homogenous saturated clay strata. This approach allows the calculation of the rate of soil compression and the time-period over which consolidation settlement will take place.

For a given increase in total stress, the extent to which the transfer from total to effective stress has progressed is known as the degree of consolidation, U (expressed as a percentage). For one-dimensional consolidation of a clay layer subjected to uniform normal loading, the process is described by the following differential equation:

$$\frac{\partial u_e}{\partial t} = \frac{k}{\rho_w g m_v} \frac{\partial^2 u_e}{\partial z^2} \tag{3.9}$$

where:

u_e = excess pore water pressure at time, t, at a given point
z = vertical height of that point
k = coefficient of permeability of the clay
m_v = coefficient of volume compressibility of the clay
ρ_w = mass density of water
g = gravitational acceleration

In Equation 3.9, it is possible to replace the compound coefficient on the right-hand side with c_v, the coefficient of consolidation, where:

$$c_v = \frac{k}{\rho_w g m_v} \tag{3.10}$$

and so Equation 3.9 becomes:

$$\frac{\partial u_e}{\partial t} = c_v \frac{d^2 u}{\partial z^2} \tag{3.11}$$

The coefficient of consolidation, c_v, describes the combined effects of permeability and compressibility of a soil on the rate of volume change (Punmia, 1994). It relates the change in excess pore pressure (with respect to time) to the amount of water draining out of the voids of a soil sample during the same time as a result of consolidation (Head, 1988).

Equations 3.9 and 3.11 are expressed in terms of *U* (%) as a function of c_v, *h* and *t*, where *h* is the longest drainage path, i.e.:

$$U\% = f\left(\frac{c_v t}{h^2}\right) \tag{3.12}$$

Importantly, from Equation 3.12 it can be seen that in addition to the permeability of the sediment at a given effective stress, the degree of consolidation reached after a certain time, and thus the time taken for primary consolidation, is inversely proportional to the square of the length of the drainage path (Head, 1988; Lambe and Whitman, 1979).

Terzaghi's consolidation theory can be applied empirically when used in conjunction with data obtained from oedometer tests. At the end of each load increment stage when excess pore pressures have dissipated and returned to equilibrium values (i.e. when primary consolidation has ended), a graph may be plotted of settlement against time. Time is plotted either on a logarithmic scale ('log-time settlement curve', Figure 3.10) or the square-root of time is taken and used as the abscissa ('square-root time settlement curve', Figure 3.11). Both of these curves are typical of normally consolidated argillaceous deposits during a load increment duration of 24 hours. They are characterised by a straight line portion which corresponds to the operation of primary consolidation processes (Head, 1988). Following this stage, there is a well-defined decrease in the gradient of the slope, indicating that excess pore pressures have reached a very small, residual value (Leonards and Girault, 1961). Through analysis of the straight line (primary consolidation) portion of time-settlement curves, and by relating this to the theoretical time-frame provided by Terzaghi's consolidation equation (Equation 3.9), the coefficient of consolidation, c_v, can be determined. When used in conjunction with additional parameters estimated from laboratory and/or field data describing the permeability and compressibility of a soil for a given load increment, the time taken for various degrees of primary consolidation under field conditions can be estimated.

3.5.1 Application of consolidation theory to intertidal sediments

In order to address the time-dependency issue, Pizzuto and Schwendt (1997) used a more complex numerical method to calculate the rate and magnitude of consolidation in Holocene coastal deposits at Wolfe Glade, a Delaware saltmarsh. Finite-strain consolidation theory developed by Gibson *et al.* (1967; 1981), a modified version of Terzaghi's consolidation theory, was used to determine the voids ratio as a function of time, *t*, by solving the following partial differential equation:

$$\frac{\partial^2 e}{\partial z^2} - \lambda(\gamma_s - \gamma)\frac{\partial e}{\partial z} = \frac{1}{G}\frac{\partial e}{\partial t} \qquad (3.13)$$

where:

$(\gamma_S - \gamma)$ = the submerged unit weight of a sediment layer
G = the finite-strain coefficient of consolidation (a modified c_v)
z = the volume of solid material per unit area above an arbitrary datum.

The parameter λ in Equation 3.13 describes the compressibility of the sediment layer and so partly governs the (assumed) negative exponential relationship between e and σ':

$$e = (e_i - e_f)\exp(-\lambda\sigma') + e_f \qquad (3.14)$$

where e_i and e_f are the voids ratios at the beginning and end of primary consolidation respectively. Equations 3.13 and 3.14 therefore not only describe the stress-strain behaviour of an individual sediment layer, but also the degree of change in voids ratio for a given time step in the model (i.e. a consideration of consolidation rate). The method was coded as a FORTRAN program ('SQUISH3') that accumulates layers of specified voids ratios at specified time intervals (determined by the chronostratigraphy of the sediment column under analysis). SQUISH3 calculates the changes in voids ratio in each layer resulting from overburden stresses caused deposition of new layers during each time step.

Three main lithological units were preserved in the compacted Wolfe Glade stratigraphic sequence; a basal organic-rich mud (indicative of freshwater wetlands) and an intermediate muddy subtidal estuarine unit are overlain by a contemporary organic-rich saltmarsh deposit. The forward-iterative model requires values of e_i, e_f, G, λ and z (at the contacts between layers) for each of these materials. For the muddy subtidal unit, they were obtained by oedometer consolidation testing of an undisturbed sample. However, owing to the abundance of wood fragments in the organic facies, oedometer tests of this unit were not undertaken. Instead, material properties for the organic units were varied until:

1. The compacted layers obtained from SQUISH3 were equal in thickness to those observed in the core.
2. Calculated voids ratios were similar to those measured in the core samples.
3. The decompacted age-altitude graphs were consistent with a compaction-free basal peat sea level curve for Delaware produced by Belknap and Kraft (1977).

This calibration procedure limits the use of the model to a solely exploratory capacity. Pizzuto and Schwendt (1997) admitted that other combinations of calibrated material properties could have reproduced the sea level history equally as well. Although SQUISH3 provides a quantitative evaluation of consolidation at Wolfe Glade, it adds little to wider principles of the geotechnical response of intertidal sediments to loading. Current understanding of the relevant processes remains limited (Pizzuto and Schwendt, 1997). Therefore, despite being arguably the most sophisticated autocompaction model to date, by considering both the stress-strain relationship and the time-dependency of strain developments, SQUISH3 is unreliable and ineffective without sufficient geotechnical input. Application of Terzaghi's consolidation theory to intertidal (particularly organogenic) sediments without a sensible appraisal of its geotechnical and geomorphological applicability can lead to incorrect conclusions being drawn regarding rates of autocompaction and sea level change.

3.6 Assumptions of existing models of autocompaction

The rheological models employed in studies of autocompaction in intertidal environments are based on the 'conventional wisdom' of Terzaghi's compression law and consolidation theory (Massey *et al.*, 2006; Paul and Barras, 1998; Pizzuto and Schwendt, 1997; Tovey and Paul, 2002). Both of these approaches are based on a number of assumptions that are rarely satisfied in practice (Head, 1998).

For valid application of Terzaghi's compression law (Equation 3.4), the lithological and geotechnical conditions in the specific environment must be the same as, or at least very similar to, those for which the law was initially developed. These assumptions are:

1. the materials are horizontal, lithologically homogeneous, laterally confined fine-grained materials of uniform thickness and low organic content (<5% loss on ignition; Hawkins, 1984);
2. the sediments are completely saturated (i.e. all voids are completely filled with water).
3. the individual soil particles are incompressible;
4. drainage of pore water and compression are one-dimensional (vertical) only;
5. volumetric deformation is a result of primary consolidation only i.e. compression is stress-dependent only; specifically, voids ratio is a unique function of effective stress;
6. primary consolidation is the result of increasing vertical total stresses resulting from the deposition of superincumbent material and no other processes;
7. the sediments are normally consolidated, and hence follow the normal compression line, from *c.* 1 kPa (i.e. at 'structural density' when effective stresses first develop; Sills, 1998) to a final value (zero porosity);
8. throughout all depths of burial, only mechanical modification (bulk density increase) to the soil matrix is occurs: diagenesis is unimportant and so does not influence the general form of a one-dimensional, voids ratio-effective stress ($e\log_{10}\sigma'$) plot.

These assumptions also apply to Terzaghi's consolidation equation (Equation 3.9), which has the following additional assumptions (summarised from Das, 1998; and Head, 1988):

1. Darcy's law for the flow of water through the soil is valid;
2. soil properties, such as the coefficients of permeability and volume compressibility, remain constant during any one increment of applied effective stress;
3. the initial excess pore pressure due to the application of load is uniform throughout the depth of the soil layer;
4. the extended duration of the consolidation period results entirely from the low permeability of the soil;
5. one or both of the strata adjacent to the consolidating soil layer are perfectly free draining in comparison with the clay.

3.7 Modelling the effects of secondary compression ('creep')

The assumption of Terzaghi's compression law that all settlement is a result of primary consolidation only is simplistic. Examination of the time-settlement laboratory curves in Figures 3.10 and 3.11 allows two further settlement processes to be identified. Before drainage of pore water begins in response to non-equilibrium pore pressures, the application of a load increment results in a simultaneous phase of 'initial compression'. This results partly from the 'bedding down' of contact surfaces in laboratory equipment and partly from the compression of small pockets of gas which may be present despite initial saturation stages (Head, 1988). Some proportion of the initial compression may be due to elastic compression of the soil particles. This process is represented graphically as a deviation from the straight line consolidation phase (Figures 3.10 and 3.11) (Head, 1988).

More importantly, towards the end of primary consolidation phase when U% approaches 100%, the gradient of both curves begins to rapidly decrease and secondary compression ('creep') begins to occur. The smooth nature of the transition between the two gradients suggests that these time-dependent components of sediment compression occur simultaneously (Figures 3.10 and 3.11). Creep results from the continued movement and rearrangement of soil particles as the soil structure adjusts itself to the ambient stress conditions. Since excess pore water pressures have dissipated, it is clear that creep deformation is occurring at a constant effective stress.

In both laboratory and field experiments, creep is evident as a linear relationship on a log-time settlement plot (Cooper and Rose, 1999; Das, 1998; Head, 1988; Hobbs, 1986; Lefebvre *et al.*, 1984). The rate of creep is typically described by the slope of this line in terms of strain per log cycle of time. The linear portion of the secondary compression 'tail' is extended so that it covers one complete log cycle of time (Figure 3.12). The vertical displacement that occurred during this log cycle is then divided by the initial height of the test specimen to calculate the coefficient of secondary compression, C_α (Figure 3.12). Values of C_α can be used to predict the magnitude of settlement in field scenarios occurring as result of creep (Das, 1998; Nash *et al.*, 1992). Values of C_α have been found to be independent of applied effective stress for mineralogenic sediments (Head, 1988).

In argillaceous deposits, creep deformation at constant effective stress usually occurs at a slower rate than that of primary consolidation (Powrie, 2004). It is also generally considered to be a minor component of total compression in mineralogenic soils (Head, 1988). However, it can still be a significant process, contributing appreciable additional volumetric strains over long periods. Crawford and Morrison (1996), for example, analysed 22 years of settlement and piezometric data obtained from a deep (200 m thick) stratigraphic sequence of interlayered sands, silts and clays of the Fraser River delta, Vancouver, Canada. The data were collected as part of a construction program in the region. The sediments were 'preloaded' with dredged sand prior to the main phase of construction to minimise post-construction settlement and the consequent damage to buildings and other structures. By comparing pore water pressure readings obtained from the piezometers with the total stress applied by the overburden of dredged sand, Crawford and Morrison (1996) were able to determine the time after the initiation of preloading that excess pore water pressures dropped to equilibrium values (i.e. the time at which primary consolidation ceased). This generally occurred very quickly, within 21 days of the application of the preload. However, following the cessation of primary consolidation, settlement continued. When plotted against log time, the straight line indicated the operation of creep processes. Following preloading, the dredged sediment was removed and construction began; the new structures created a total stress less than that of the dredged material but creep continued to occur at a similar rate. Nineteen years after the completion of the construction project, the magnitude of creep had equalled that of primary consolidation (Crawford and Morrison, 1996). When considered over a longer, Holocene timescale, it is clear that creep may be the dominant compression process.

The consequences for modelling the volumetric behaviour of soils subjected to prolonged loading after excess pore-pressures have dissipated is that the volumetric state is not solely a function of effective stress; it also has a direct dependence on time (Berre and Iversen, 1982; Garlanger, 1972; Leroueil *et al.*, 1985; Sills, 1998). The relationship between vertical effective stress, voids ratio and time of sustained loading for normally consolidated clay soils during one-dimensional compression has been examined by Bjerrum (1967) and Hobbs (1986). They found that identical soils subjected to sustained loading (i.e. beyond 24 hours) during each load increment display identical compression indices, C_c, but with different graphical intercept values. Hence, the compressibility characteristics cannot be described by a single curve in $e\log_{10}\sigma'$ space. Rather, a system of lines is required ('isochrones'), each representing the equilibrium voids ratio for different values of effective stress at a specific time of sustained loading (and hence operation of creep processes) (Bjerrum, 1967). This basic relationship is displayed in Figure 3.13, in which the times of sustained loading have been arbitrarily selected at subsequent log cycles. Such a diagram is idealised; real-world situations would not display such convenient loading stages or durations. Nonetheless, Figure 3.13 illustrates how creep can significantly alter the magnitude of compression behaviour and, hence, the shape of an EOP $e\log_{10}\sigma'$ curve (Bjerrum, 1967). It also exemplifies how Terzaghi's compression law is unlikely to accurately describe *in situ* behaviour if creep processes are operating.

3.8 ADAPTATION OF TERZAGHI'S LAWS TO SPECIFIC AUTOCOMPACTION CONDITIONS

Terzaghi's compression law and consolidation theory offer a framework of autocompaction behaviour (dominated by primary consolidation) that is often too simplistic and rigid for use in most practical problems. However, this is not cause for rejection of the theories; rather, using Terzaghi's consolidation theory as a starting point, some authors have adapted and modified the equations to suit particular conditions.

3.8.1 Application to specific sedimentary configurations

By accounting for variations in the coefficients of permeability and compressibility during a load increment, non-linear models of consolidation were developed by Gibson *et al.* (1967) to describe the time-dependency of consolidation in thin, homogenous layers of saturated clays. Gibson *et al.* (1981) expanded the original model to account for self-weight of a soil skeleton and pore water in thick homogenous clay layers. Both models aim to provide more realistic estimates of the time taken for excess pore water pressures to dissipate for specific stratigraphic conditions.

3.8.2 Adaptation to low stress environments

Been and Sills (1981) investigated very low stress (0.01-1 kPa, two to three orders of magnitude lower than in conventional soil mechanics) behaviour of muds settling from a suspension and then consolidating. This stress range covers the transition from a fluid-supported suspension to a soft soil; this occurs when pore water pressures drop from being equal to the total vertical stress (caused by the self-weight of the soil) to being lower that then total vertical stress – i.e. when effective stresses first develop at structural densities. Previously developed consolidation theories were adapted to the experimental conditions to adequately describe the large strain behaviour that was observed in an estuarine mud from Combwich in Somerset in 100 mm settling columns. An interesting outcome of this work for compressive stress-strain ($e\log_{10}\sigma'$) decompaction analysis, extended by Sills (1998) for different fine-grained materials, is that normal compression lines may 'bend upwards' at low effective stresses, rather than extrapolating back linearly (Figure 3.14). In addition to the linear correlation between voids ratio and the common logarithm of effective stress for stresses greater than 0.8 kPa, there may also be a quasi-linear correlation within the effective stress range 0.01-0.8 kPa, but with a steeper gradient. This means that a backward extrapolation of the virgin compression line to 1 kPa, as is routinely carried out in the models outlined in Section 3.4, may underestimate the original layer thickness. Furthermore, the low stresses (< 1 kPa) question the validity of the use of 1 kPa as the reference voids ratio at the depositional surface.

3.8.3 Diagenetic processes in high stress environments

Whereas autocompaction refers to the sub-set of diagenetic processes that reduce the volume of a sediment, diagenesis describes the sum of all physical, chemical and biological post-depositional processes that occur prior to the onset of metamorphism (Nygard *et al.*, 2004). Physical ('mechanical') processes are mainly controlled by effective stress variations, sediment (both matrix and grain) strength and material compressibility. These primarily lead to volume changes due to a reduction in porosity (voids ratio) (Gutierrez and Wangen, 2005). Skempton (1970) estimated that mechanical processes dominate at burial depths less than 2 – 3 kilometres (effective stresses of *c.* 25 – 50 MPa). Indeed, at these depths and effective stresses, Terzaghi's compression law (Equation 3.4) has been shown to accurately represent the compression behaviour of argillaceous soils (Gutierrez and Wangen, 2005; Nygard *et al.*, 2004).

However, as burial depths increase in sedimentary basins of large lateral extent, chemical processes begin to dominate (Skempton, 1970); the exact depth of this transition depends on the mineralogy and petrophysical properties of the sediment, temperature variations and the flow of transport of precipitating minerals (Nygard *et al.*, 2004). Chemical processes that affect the autocompaction of sediments include chemical dissolution and precipitation. Dissolution of mineral phases causes a decrease in voids ratio and a closer packing of the grain structure with or without a change in total sediment volume (Bjorlykke and Hoeg, 1997). Chemical precipitation of minerals at grain contacts has been shown to reduce voids ratios and act as grain 'cement', reduce the compressibility of the sediment (Greensmith and Tucker, 1986). Nygard *et al.* (2004) illustrate these mechanisms by testing samples of Kimmeridge clays for their one-dimensional compression behaviour under high vertical effective stresses (up to 150 MPa, 9 km in burial depth). The first of these, the Kimmeridge Westbury Clay (KWC), is an unlithified, uncemented clay obtained from Westbury Quarry in Wiltshire, UK. The second batch of samples, the Kimmeridge Bay Clay (KBC), was obtained from Kimmeridge Bay, Isle of Purbeck, Dorset, UK. KBC is a lithified and cemented shale. X-ray diffraction (XRD) analysis of the two samples revealed that KBC has higher carbonate (aragonite, calcite and dolomite) and quartz contents than KWC. The two samples therefore differ in their diagenetic histories but are from a largely identical parent material (Nygard *et al.*, 2004).

After compression of both samples to an effective stress of 22 MPa (an effective stress greater than the previous maximum effective stress experience by both samples), a difference in porosity of 15 % was still observed between KBW (*c.* 35 %) and KBC (*c.* 20 %). Also, the compressibility of the samples in the virgin range was dramatically different between the two samples; the compression index of the cemented KBC (C_c = 0.06) is about a quarter of that of the uncemented KWC (C_c = 0.22). The reduction in porosity and decrease in compressibility are attributed to the precipitation of minerals in the sediment pores. Nygard *et al.* (2004) calculated that the porosity reduction in KBC due to chemical precipitation is equivalent to a mechanical compression under an effective stress of more than 200 MPa (more than 16 km of burial). These results generally imply that once sediments have undergone chemical diagenesis, they become more difficult to mechanically compact (Nygard *et al.*, 2004). Therefore, although mechanical compression continues to occur due to effective stress increases caused by overburden sedimentation, voids ratio is no longer uniquely related to effective stress. Since Terzaghi's compression law does not relate volume and compressibility changes to chemical processes, increasingly complex models of autocompaction would be required to adequately describe the structural and volumetric evolution of a stratigraphic column.

Physical and chemical factors can combine in deeply buried sediments to lead to the generation of 'overpressures' (i.e. pore water pressures greater than those obtained from hydrostatic profiles) (Gutierrez and Wangen, 2005). From Equation 3.2, this can reduce the effective stress in these sediments to zero or negative values, typically leading to underconsolidation and openly-structured sediments with high voids ratios (Gutierrez and Wangen, 2005). Audet (1995), for instance, modelled the thermally-activated dehydration of expandable smectite clay species to form dehydrated, non-expandable illite. This dehydration reaction, which occurs at significant burial depths (≈ 4 km) has significant implications for excess pore pressures, elevating them by as much as 30% above equilibrium (hydrostatic) values. Without dissipation of these pore water pressures, the volume of these sediments cannot be reduced despite considerable total stress overburdens. Furthermore, the presence of overpressures significantly complicates the use of Terzaghi's consolidation equation; more complicated relationships exist between permeability and consolidation in deep, overpressued systems. Significant changes in the coefficients of permeability and compressibility can occur during different loading stages (Audet and Fowler, 1992; Gutierrez and Wangen, 2005). In order to overcome such problems, Audet and Fowler (1992) present more rigorous analytical treatments of Terzaghi's consolidation theory; importantly, they introduce a dimensionless 'gravity number' which describes the ratio of the initial sediment permeability to the sedimentation rate, where a high gravity number indicates small overpressures and a low gravity number signifies large overpressures and underconsolidation. Whilst the exact mathematical descriptions of such models are beyond the scope of this chapter, they serve to illustrate how Terzaghi's consolidation equation is highly simplified and requires significant adaptation for successful use in different diagenetic and burial environments.

3.9 The dynamic intertidal environment and intertidal materials

Primary consolidation is a widely recognised and well-understood process in geotechnical engineering. However, its application to natural sediment burial needs careful consideration. Although sediments experience primary consolidation as they are buried, the research discussed above illustrates how the process is complicated by other mechanical and diagenetic processes. Hence, the established theories of Terzaghi have been carefully considered in the context of a particular stress range (e.g. Been and Sills, 1981; Nygard *et al.*, 2004) and/or diagenetic and burial environment (e.g. Audet, 1995). Without such critical evaluation and adaptation of theory, the predictive capacity of Terzaghi's basic models is likely to be dramatically decreased.

In contrast to the studies of autocompaction and diagenesis in high-pressure (1 – 200 MPa) environments outlined in Section 3.8.3 in which the idiosyncrasies of the particular stress environment have been explicitly addressed, similar studies in low stress (0 – 1000 kPa), near-surface sediments are conspicuous by their absence. As a result, Terzaghi's compression law and consolidation equation have been applied to intertidal areas without sufficient thought regarding the transferability of such models. Indeed, Pizzuto and Schwendt (1997) explicitly state that their model of autocompaction is based on number of assumptions; namely, that the sediments are not overconsolidated, that diagenesis is unimportant and that the principle of effective stress is applicable to organogenic sediments. Similarly, Paul and Barras (1998), Tovey and Paul (2002) and Massey *et al.* (2006) acknowledged that Terzaghi's compression law was developed for normally consolidated argillaceous sediments. However, none of these studies examines the characteristics and associated implications of the intertidal environment in which accurate predictive models of autocompaction and stratigraphic volume change are so desperately required. This section therefore examines the geomorphological, geotechnical and diagenetic idiosyncrasies of the intertidal environment and the materials that form there to determine whether they meet the assumptions of Terzaghi's laws.

3.9.1 Intertidal materials

Intertidal energy and ecology gradients (Section 2.2, Figure 2.1) produce a range of materials that are a great deal more diverse than the idealised clay lithologies upon which Terzaghi's models are based. Stratigraphic sequences in intertidal areas comprise successions of interbedded silt-rich soils (representative of tidal mudflats), organic silts ('low', mineralogenic salt marshes) and various types of peats that formed autochthonously in mid-high marsh (marine peats) or supratidal environments (freshwater peats). Coarser-grained lithologies are occasionally deposited by high energy events (e.g. storm surges) or by the migration of tidal channels (sands) and beach barriers (sands/gravels). Each lithology has individual physical and mechanical characteristics. Furthermore, these materials are laterally and stratigraphically variable (Allen, 2000b). Greensmith and Tucker (1986) quoted compaction values of up to 90 % of original volume for peat, 75 – 90 % for clayey silt and 25 -35 % for sands. However, these figures are maximum values that are unlikely to be obtained in Holocene coastal stratigraphies. Classifying the range of materials present into a few discrete groups is also a dramatic oversimplification of the materials found in intertidal areas (Allen, 2000a). The presence of varying organic contents, both detrital and autochthonous, in combination with varying degrees of peat decomposition results in a large variety of material types.

3.9.2 Depositional processes

The presence of vegetation changes the processes of deposition on a marsh surface, indicating that the models of low stress consolidation developed by Been and Sills (1981), for example, are not applicable to intertidal environments. Most obviously, organogenic sedimentation is a dramatically different process of sedimentation to the suspension/settling processes modelled by Been and Sills (1981). In terms of mineralogenic sedimentation in intertidal areas, direct settling of flocs from suspension does occur, but it is not the only process. In the Tay Estuary, Alizai and McManus (1980) found significant amounts of silt adhering to the upright *Phragmites* stems. Similarly, on Hut Marsh, North Norfolk, up to 8.5% of the total sedimentary deposit is trapped on stems, leaves and branches (French and Spencer, 1993). Allen (2000b) suggested that thicker, silt-rich 'crusts' may flake off the plant surfaces and fall directly on to the depositional surface. Even on the vegetation-free mudflat surfaces, depositional processes are not entirely analogous to those studied by Been and Sills (1981) due to the diurnal tidal ebb and the consequent exposure to subtidal conditions, the effects of bioturbation and the sediment trapping afforded by the presence of biofilms (Austen *et al.*, 1999). Such a combination of depositional factors is likely to result in a degree of stochasticity in soil structures found at the depositional surface. Indeed, the effects of intertidal depositional processes on structure and subsequent autocompaction behaviour are unquantified.

3.9.3 Post-depositional changes in organic facies

The presence of organogenic facies in particular significantly complicates autocompaction analysis and modelling. In a solely mechanical sense, peats are one of the most compressible soils that exists (den Haan *et al.*, 1994). Instead of consisting of individual mineral particles, organic soils consist of plant remains in the form of rhizomes, roots, leaves and stems. This violates one of the fundamental assumptions of Terzaghi's consolidation and compression laws, since the soil particles themselves are compressible and effective stress changes are not required for soil deformation. As a result, secondary creep processes are a much more important factor in the compression of these soils (Berry and Poskitt, 1972; Head, 1988; Hobbs, 1986; Lefebvre *et al.*, 1984).

The living root network creates an initial high porosity, 'open' soil structure. In *Spartina alterniflora*, for example, the below-ground biomass contains a series of lacunae that form a continuous gas-filled system that occupies a significant amount of root volume (Delaune *et al.*, 1994). Arenovski and Howes (1992, cited in DeLaune *et al.*, 1994) stated that lacunae occupy approximately 30% of root cross sectional area in live, turgid *Spartina alterniflora* roots. Upon death, loss of plant turgor leads to a significant structural collapse (Delaune *et al.*, 1994) that is unrelated to changes in effective stress. Such collapse would manifest itself as an overconsolidated horizon.

Following the death of living matter, organic soils are also prone to decomposition. Hackney and de la Cruz (1980) investigated the rate of *in situ* decomposition of the roots and rhizomes of *Juncus roemerianus* and *Spartina cynosuroides* – two plant species that dominate the belowground productivity in a saltmarsh located on the western side of St. Louis Bay, Hancock County, Mississippi Gulf Coast, USA. They found that the decomposition rate was greatest (20% organic mass loss per year) in the top 10 cm of the soil profile. They also found no apparent decomposition below 20 cm depth and suggested a relatively constant decomposition rate throughout the year. However, these tests were only undertaken on two species in a particular environment and are too short-term (up to 12 months) to offer any quantitative predictive insights into rates of humification.

Humification results from a complex set of physical, chemical and biological processes including abiotic and biotic fragmentation, microbial decay and chemical transformation (Lillebø *et al.*, 1999). At a general level, humification occurs when dead plant matter is exposed to air or aerated groundwater (Smith, 1985). The rate of decay is dependent on the aerobic state; waterlogged soils lead to anaerobic conditions and slower decomposition rates (Clymo, 1965). The rate of decomposition in these areas is decreased further by slower heating of the waterlogged soils (Martin and Holding, 1978). The type of vegetation present also influence the rates of decomposition through the effect of the texture and fibre strength of the rhizospheric tissue (Hackney and de la Cruz, 1980) and the tendency of some plant species (e.g. *Spartina* spp.) to pump oxygen through the soil to their roots, promoting decay at normally reducing depths.

The effects of humification can be profound, altering a coarse fibrous newly deposited material showing little or no signs of decomposition to an amorphous, highly humified black peat with a granular appearance. Humification not only changes the bulk density of the material without a change in effective stress; it potentially alters the mechanical properties and hence future behaviour of the material under an imposed compressive stress. Such variation in mechanical properties between a newly deposited unhumified peat and a semi-/fully-decomposed peat are, as yet, unquantified, particularly for wholly organic high marsh peats and silt-rich low marsh organogenic deposits.

3.9.4 The dynamic intertidal environment

Two critical assumptions for the application of Terzaghi's compression law in its unaltered format are that sediments remain permanently and fully saturated and that effective stresses only occur as a result of overburden sedimentation. For these assumptions regarding the constancy of hydrological/hydraulic and subaerial conditions to be valid, the overlying water body from which deposition occurs by settlement from a suspension (Lintern, 2003; Sills, 1998) must remain in place. However, the depositional and near-surface environment in intertidal areas varies (semi-) diurnally in response to the tidal cycle. The direct (e.g. mechanical loading of sediments by tidal waters) and indirect (effective stress changes as a result of groundwater variations; subaerial exposure; and diagenetic processes) effects of the tidal cycle have the potential to cause significant deviations from the form of a conventional Terzaghi (straight line) $e\log_{10}\sigma'$ plot.

3.9.5 Mechanical loading of sediments by tidal waters

In the majority of hydrostatic situations, above-ground surface (tidal) water does not have any measurable influence on the voids ratio of the underlying strata (Punmia, 1994). This point can be illustrated with reference to Figure 3.15. At level B, the total pressure is equal to $h_1\gamma_w$. Similarly, the pore pressure is equal to $h_1\gamma_w$. By Equation 3.2, the effective stress is equal to zero;

i.e. $$\sigma' = \sigma - u = h_1\gamma_w - h_1\gamma_w = 0 \qquad (3.15)$$

Similarly, the tidal load has no influence on the effective stress at level C:

$$\begin{aligned}
&\sigma = h_2\gamma_{sat} + h_1\gamma_w \\
&u = (h_1 + h_2)\gamma_w \\
&\sigma' = \sigma - u \\
&\sigma' = (h_2\gamma_{sat} + h_1\gamma_w) - ((h_1 + h_2)\gamma_w) \\
&\sigma' = h_2(\gamma_{sat} - \gamma_w) \\
&\sigma' = h_2\gamma' \qquad (3.16)
\end{aligned}$$

Therefore, the effective stress at level C is equal to the thickness of the soil multiplied by the submerged unit weight of the soil, γ', which is equal to the saturated unit weight of the soil, γ_{sat}, minus the unit weight of water, γ_w. It does not depend on the overlying column of surface (tidal) water of depth h_1 (Punmia, 1994).

This general principle is valid as long as the water is in continuous contact in the pores throughout the entire depth of the stratigraphic column and across the soil-water interface at the sediment surface (Powrie, 2004). However, hydraulic disconnectivity may occur within the uppermost part of the stratigraphic profile as a result of the following, particularly when they occur in synergy:

1. a severe desiccation event (Greensmith and Tucker, 1971a, 1971b; Hawkins, 1984) leading to the formation of an overconsolidated, impermeable horizon and/or a pronounced unsaturated zone;
2. deflocculation of clay aggregates following 'immersion' in meteoric waters, resulting in dense, low permeability or impermeable overconsolidated horizons (Crooks, 1999; Crooks and Pye, 2000);

3. the presence of pronounced horizontal lamination within the soil, resulting in strong anisotropic (im)permeability (Hawkins, 1984);
4. the formation of diagenetic aquicludes/aquifuges as a result of the post-depositional redistribution of redox sensitive elements (Lascelles *et al.*, 2000; Thomson *et al.*, 2002).

The operation of these processes would modify the vertical distribution of pore water pressures or completely decouple pore pressures above and below an hydraulic discontinuity. If such hydraulic discontinuity were to occur, the application of a tidal load on the surface of the sediment column would be analogous to a surcharge of $h_c\gamma_w$ placed on the saturated soil mass, where h_c is the height of the water column above the ground surface and γ_w is the unit weight of water (9.81 kN/m^3). The loss of hydraulic connection would prevent pore pressures from counteracting the application of the tidal total stress. Hence, the flooding of the mudflat and/or saltmarsh surfaces would cause an increase in compressive total stress. Given the diurnal nature of the tidal cycle, the volumetric strains may be considerable over time. Tidal loading may then be recorded in the sediments as overconsolidated strata.

3.9.6 Groundwater variations and hydrostatic stresses

Further changes in effective stress may be caused by variation in groundwater levels, again in response to tidal conditions. In conventional soil mechanics analysis, the required calculations depend upon whether the free (i.e. non-adsorbed) water is stationary (hydrostatic conditions) or in motion (hydrodynamic conditions).

At the level of the water table, pore water pressures are atmospheric (0 kPa). Under hydrostatic conditions, pore water pressures (u) are positive and increase linearly beneath the water table at a rate of 9.81 kPa m^{-1}. If the total stress (σ) at a particular depth is known (Equation 3.1), effective stress (σ') can be calculated (Equation 3.2) (Figure 3.16).

Above the water table, sediments can remain saturated due to the rise of capillary water. This phenomenon results from the existence of surface tension in the water which pulls it upwards against gravitational force (Powrie, 2004; Punmia, 1994). In these circumstances, pore water pressures are less than atmospheric (negative suction pressures) (Das, 1998; Powrie, 2004; Punmia, 1994). Therefore, by Equation 3.2, the effective stress is greater than the total stress and so materials in the capillary zone tend to be overconsolidated. Pore water pressures decrease above the water table at a rate of 9.81 kPa/m. Therefore, a drop in the water table of 1 m would result in an effective stress at the surface of 9.81 kPa (Figure 3.17).

Capillary water will continue to rise within sediments above the water table until a critical value, the air entry value (U_e), is reached (Powrie, 2004). The height of capillary rise above the water table depends on the diameter of the soil pores (a function of particle size and sorting) and the value of surface tension in the water (Punmia, 1994). The negative pore water (suction) pressure at air entry (measured relative to ambient atmospheric pressure) can be estimated by considering the equilibrium of a hemispherical water meniscus in a circular pore of diameter, *d*:

Force due to surface tension around the rim of the meniscus	=	Force due to difference in pressure between the pore water and the air

$$\pi dT = \left(\frac{\pi d^2}{4}\right)U_e \tag{3.17}$$

or

$$U_e = \frac{4T}{d} \tag{3.18}$$

where *T* is the surface tension of the water/air interface (= 7 × 10^{-5} kN/m at 10°C) (Powrie, 2004). If the air entry value is exceeded by the capillary suction pressures in the pore water above the water table, the soil will draw in air through any surface that is exposed to the atmosphere. The parameter *d* can be approximated by calculating the 10th percentile of the cumulative particle size distribution of a soil, D_{10} – the effective pore size (Powrie, 2004; Punmia, 1994). From Equation 3.18, it can be seen that coarser grained soils (higher D_{10} value) have lower air entry values than fined grained cohesive soils. As a result, coarser sandy materials above the water table are generally unsaturated (Powrie, 2004). Conversely, silts and clays have greater air entry values and so remain saturated above the water table for several metres (Powrie, 2004).

3.9.7 Hydrodynamic conditions, groundwater flow and seepage pressures

Any change in pore water pressure from the hydrostatic value causes free water to flow through the voids between soil particles (Powrie, 2004). If water is flowing, effective stresses differ from those experienced under hydrostatic conditions (Das, 1998; Lambe and Whitman, 1979).

When considering fluid flow, potential and kinetic energy are conventionally expressed in terms of 'head' (unit of energy per unit of mass) (Lambe and Whitman, 1979). Three types of head are used in analysis of flow of groundwater through permeable soils (Figure 3.18):

1. pressure head, h_p: the pore water pressure, u, at a point divided by the unit weight of water, γ_w;
2. elevation head, h_e: the vertical distance of a given point above or below an arbitrary, yet constant, datum;
3. total head, $h = h_p + h_e$: the sum of the pressure head and elevation head.

Each of these parameters has the dimension of length. Both pressure and elevation heads contribute to movement of water through soils. It is total head that determines flow; if a total head gradient exists between two points, flow will occur through the soil (Lambe and Whitman, 1979). The change of total head with distance can be expressed in non-dimensional form as:

$$i = \frac{\Delta h}{L} \qquad (3.19)$$

where i is the hydraulic gradient and L is the length of flow over which the loss of head, h, occurred (Das, 1998) (Figure 3.19).

As water flows through a soil and total head is lost, an energy transfer occurs as a result of the frictional drag of water on soil particles (Terzaghi and Peck, 1948). The stress corresponding to this energy transfer is called the seepage pressure, p_s (Lambe and Whitman, 1979; Punmia, 1994). It is the pressure exerted by water on the soil through which it flows. If h is the total head lost due to the frictional drag of water on a soil mass of thickness z, the seepage pressure is given by (Das, 1998; Punmia, 1994):

$$p_s = h\gamma_w$$

or

$$p_s = \left(\frac{h}{z}\right) z\gamma_w = iz\gamma_w \qquad (3.20)$$

where:

z = the vertical distance over which the head, h, is lost and
i = the hydraulic gradient.

The vertical effective stress may be increased or decreased due to the seepage pressure depending on the direction of flow. Thus the effective stress in a soil mass subjected to an additional seepage pressure is given by:

$$\sigma' = z\gamma' \pm p_s = z\gamma' \pm iz\gamma_w \tag{3.21}$$

In an isotropic soil, the seepage pressure always acts in the direction of flow (Lambe and Whitman, 1979). If the flow occurs in a downward direction, the effective stress is increased ($+ p_s$). If flow occurs in an upward direction, the effective stress is decreased ($-p_s$).

3.9.8 Subaerial processes: diurnal exposure and desiccation

Intertidal sediments are also prone to a range of subaerial conditions that can potentially affect their mechanical strength (Tovey and Yim, 2002). In particular, intertidal sediments may be desiccated if exposed for significant durations, especially during hot, dry weather. This desiccation process decreases the moisture content of a soil. In the intertidal fine silty clays of Avonmouth in the Severn Estuary, UK, Hawkins (1984) recorded moisture content decreases from 80% to 64% after two hours of exposure following tidal recession. This figure decreased to 56% after four hours of exposure. Such desiccation has been shown experimentally to increase the effective suction pressures acting on the soil skeleton. During laboratory experiments in which London Clay mixed with sand in varying proportions was allowed to dry naturally, Marinho and Chandler (1993) found an inverse negative relationship between water content (%) and common logarithm of suction stress (kPa). Although less well defined, a similar inverse linear relationship was observed between voids ratio and the common logarithm of suction stress (kPa). The desiccation of clay soils may lead to significant overconsolidation since they can maintain suction pressures varying between 65 kPa (30% clay, 70% sand mixture) and 7 MPa (pure clay) (Marinho and Chandler, 1993).

At a critical point, the limiting suction stress that a soil can maintain is reached – the air entry value. This value increases with decreasing pore size. If the air entry value is exceeded, the soil will begin to draw in air and desaturate (Powrie, 2004). Unsaturated conditions are of particular importance. The presence of air in the soil pores renders the pore 'fluid' compressible and leads to permeability changes under the influence of applied stress. This results in different characteristic features of the time-settlement curves to those illustrated in Figures 3.10 and 3.11. These features are (Fredlund and Rahardjo, 1993b; Head, 1988):

1. a large initial compression;
2. a log-time settlement curve that is flatter than the theoretical saturated curve in the primary consolidation phase;
3. a square-root time relationship which is continuously curved, instead of showing an initial linear portion;
4. a steeper secondary compression line.

Essentially, unsaturated soils operate in different ways to saturated soils and hence cannot be considered within the 'classical' framework of Terzaghi's theories (Fredlund and Rahardjo, 1993a; 1993b). Barden (1965) conducted a detailed theoretical study, but his efforts did not yield a standard curve fitting procedure for the time-settlement analysis of unsaturated clay soils.

3.9.9 Subaerial exposure of intertidal sediments over longer timescales

Subaerial exposure as a result of falls in sea level influence the geotechnical properties of intertidal sediments over centuries to millennia. Greensmith and Tucker (1971a; 1971b; 1973; 1976) tentatively suggested five sea level 'regressions' superimposed on a long-term transgressive trend in the Holocene sediments of southern Essex and the inner Thames Estuary. This interpretation was influenced by the identification of overconsolidated strata in silts and silt-rich clays. These layers had significantly lower moisture contents and higher densities granulometrically similar yet softer, normally consolidated sediments that over- and under-lie them. The shear strength (τ_f) of the overconsolidated layers varied between 43.2 and 121.7 kPa. In stark contrast, the surrounding normally consolidated sediments possessed shear strengths of 13.7 to 29.4 kPa. Greensmith and Tucker (1971b) point out that the shear strength of the overconsolidated layers equals that of the London Clay (τ_f = 49.1 to 147.4 kPa) – a sediment heavily overconsolidated under at least 150 m of late Tertiary sediment, whereas the maximum possible previous overburden experienced by the overconsolidated layers was *c.* 35 m.

Greensmith and Tucker (1971a; 1971b) inferred that these overconsolidated beds, up to 4 m thick, reflect periods of emergence well above Highest Astronomical Tide and a consequent desiccation during each 'regressive' episode. As supporting evidence to periods of exposure and desiccation, Greensmith and Tucker (1971b) discussed the physiographic nature of the overconsolidated deposits: a sharp vertical lithological change at the top of the overconsolidated unit (from silty clay to sand, for example); a change in microfaunal assemblage indicating an increasing marine influence (a change from brackish species to open marine, for instance); an increase in organic carbon content, often as traces of peat, reflecting emergence from the influence of mineralogenic deposition; a mottled colour and texture caused by penetration by plant rootlets or bioturbation; sedimentary structures reminiscent of desiccation cracks; and changes in colour, possibly reflecting a change in oxidation state of the elements present. On the basis of the stratigraphic architecture of the area, Greensmith and Tucker (1971a) also suggested synchronous widespread river channel incision associated with a lowered erosive base level during regressive phases.

Hawkins (1984) rejected the hypothesis put forward by Greensmith and Tucker (1971a; 1971b; 1973; 1976) that the overconsolidated horizons are indicative of falls in relative sea level. Given the geographical proximity of the Severn Estuary marshes and the marshes of southeast England (240 kilometres), Hawkins (1984) suggested a regional sea level oscillation should lead to overconsolidated horizons in both areas. However, following investigations into the geotechnical properties of the intertidal Holocene sediments of the Severn Estuary, Hawkins (1984) found no evidence of overconsolidated strata in the Severn Estuary sediments. Hawkins (1984) therefore advocated a steady, rather than oscillatory, rise in sea level in southern Britain during the Holocene.

By invoking clay dispersion theory and by using sedimentological and geotechnical analysis of short cores from natural, reclaimed and reactivated Essex and Severn Estuary marshes, Crooks (1999) suggested a further mechanism for the formation of overconsolidated horizons in these clay-rich intertidal sediments. He illustrated that the presence of divalent calcium ions (derived from detrital calcium carbonate) in clay-rich saltmarsh substrates was sufficient to displace monovalent, more weakly bonded sodium ions from the exchange sites on clay particle surfaces. Calcium cations have a more compact diffuse double layer then sodium cations and this allows closer particle interaction and stronger attractive van der Waals forces (Rowell, 1994). The dimensions of the diffuse double layer are also influenced by the osmotic potential between the fluid at the clay particle surface and the 'bathing' pore fluid. When the concentration of cations in the bathing solution is lower than that of the diffuse double layer water film, water molecules migrate from the former to the latter, increasing its thickness and reducing inter-particle van der Waals attractive forces until they are so weak that the particles disperse (deflocculate). Therefore, when the saline water table falls in calcium-deficient soils (such as those in the Essex marshes) during a regressive phase of relative sea level, it is replaced by non-sodic meteoric water and the clay particles disperse. This leads to a loss of soil structure, and the creation of dense, low permeability overconsolidated horizons with high shear strength. The calcium-rich Severn Estuary marsh soils therefore do not disperse during a sea level fall. The detrital calcium carbonate content 'protects' the clay aggregates from dispersion under low-salinity conditions due to stronger inter-particle attractive forces creating a denser initial structure and a negligible osmotic gradient when the soil is submerged in fresh water (Crooks, 1999; Crooks *et al.*, 2002; Crooks and Pye, 2000).

The identification of clay dispersion as a mechanism for the formation of overconsolidated horizons and illustrating that it is geochemically controlled and regionally variable means that historical oscillations of sea level during the Holocene in the Severn Estuary can no longer be ruled out on the basis of geotechnical data. Indeed, van de Plassche (1986) advocates the use of the full range of bio-, litho, chemo- stratigraphic and geotechnical data when undertaking reconstructions of former sea level. Nevertheless, the studies of Greensmith and Tucker (1971a; 1971b; 1973), Crooks (1999) and Crooks and Pye (2000) provide additional mechanisms by which sediments can become overconsolidated without an increase in effective overburden stress. They also therefore question the validity of a universal model of autocompaction, since changes in exposure time and sedimentology can lead to significant departures from the compression behaviour described by Terzaghi's compression law.

3.9.10 Diagenetic remobilisation of concretionary materials

Over monthly, annual and decadal timescales, early stage diagenetic processes can alter the virgin geochemical state of intertidal materials. Reduction-Oxidation ('redox') reactions have been observed in geochemical studies of marine (e.g. van der Weijden, 1992) and intertidal sediments (e.g. Cundy and Croudace, 1995b, , 1996; van Huissteden and van de Plassche, 1998). Such aquatic redox processes are associated with the remineralisation of organic carbon in decaying marsh vegetation and are bacterially-driven (Stumm and

Morgan, 1981). The dynamism of the water table in intertidal environments results in periodic shifts from oxidising (above the water table) to reducing (beneath the water table) conditions. These processes lead to geochemical zonation in the stratigraphy due to the postdepositional relocation of redox sensitive elements (Thomson *et al.*, 2002) (Figure 3.19). The geochemical boundaries that develop are often visually evident (Williams *et al.*, 1994). In saltmarsh stratigraphies, upper oxidised mottled red/grey zones are typically enriched in iron (Fe) and manganese (Mn) oxyhydroxides (Croudace and Cundy, 1995; Cundy and Croudace, 1995a; 1995b). Beneath this zone lies a grey/black layer with no mottling. The colour of this layer indicates the presence of iron sulphides and more reducing conditions than in the layer above (Spencer *et al.*, 2003). The upper boundary of this layer may represent the lowest position of the permanent water table, below which conditions are permanently reducing (Cundy and Croudace, 1995b). In mudflat environments, the permanent water table is less deep, leading to a thin, brown oxic surface layer underlain by a uniformly grey/black zone devoid of any sedimentary structures (Cundy and Croudace, 1995b). The enrichment of redox sensitive elements (particularly Fe) in specific stratigraphic locations may have an influence as a diagenetic point contact cement (Hawkins, 1984; Tovey and Yim, 2002), increasing bulk density resulting in overconsolidation, changing the material compressibility and therefore altering future mechanical behaviour.

However, unlike the analysis of feedbacks between diagenetic processes, geotechnical structure (voids ratio/density) and compressibility at high stresses undertaken by Audet (1995) and Nygard *et al.* (2004), no such data exist for intertidal materials. Hence, the effect of diagenetic alteration on compression behaviour and any resultant departures from that described by Terzaghi's compression law remains unquantified in these low stress environments.

3.9.11 The effects of land reclamation

The reclamation, embankment and drainage of estuarine saltmarshes for agricultural purposes throughout the last millennium but particularly the last 200-300 years (Allen, 2000b), has resulted in the operation of processes that are analogous to a relative sea level fall in active, natural saltmarshes: a cessation of tidal flooding, a lowering of the saline water table and prolonged exposure to subaerial conditions and desiccation (Crooks, 1999). Barras and Paul (2000) examined the post-reclamation changes in the Carse Clays of Bothkennar in the Forth estuary, Scotland; an area first embanked in the late 1700s. Barras and Paul (2000) described a range of geotechnical and geochemical changes similar to those outlined above. Rapid effective stress increases caused by artificial drainage and desiccation have led to overconsolidation by approximately 150-200 kPa. The post-reclamation chemical development of the soil profile has involved desalinisation, oxidation and acidification. Barras and Paul (2000) suggested that the geotechnical changes occurred rapidly and are now complete, but the geochemical alteration is ongoing – the chemical soil profile is still incomplete after 200 years.

3.10 SUMMARY, THESIS FOCUS AND KEY RESEARCH ISSUES

The models outlined in Section 3.8, adapted from Terzaghi's theories, illustrate that an autocompaction model is a highly idiosyncratic tool which must be modified and tailored to suit the stress range and the associated diagenetic processes operating in a particular burial environment. However, existing Terzaghi-based rheological models of compression and consolidation have been applied to problems of intertidal autocompaction with no regard for the specific conditions of the intertidal environment. The applicability and accuracy of these models have been entirely assumed. Furthermore, the dynamics of the intertidal environment and the material complexities outlined in Section 3.9 strongly suggest that Terzaghi models are unlikely to be valid (Table 3.1). Indirectly obtained geotechnical parameters (Equation 3.7) are not applicable to organic-rich intertidal sediments. Similarly, obtaining such parameters by calibration (Pizzuto and Schwendt, 1997) contributes nothing to the formulation of a phenomenologically correct conceptual model of intertidal autocompaction. Furthermore, extrapolation of the normal compression line across all stress states has been shown to be invalid in clay materials (Sills, 1998), let alone in geotechnically anomalous materials and environments (cf. Smith, 1985; Paul and Barras, 1998; Tovey and Paul, 2002).

Put simply, there is a lack of empirical data obtained from field and laboratory investigations into the basic one-dimensional autocompaction behaviour of intertidal sediments and the modifying effects of subaerial and diagenetic processes. A framework of intertidal sediment autocompaction behaviour has never been formulated. However, such a framework is critical to the development of accurate and realistic conceptual representations and mathematical descriptions of autocompaction behaviour in intertidal areas.

Given the large stress range, a multitude of sedimentary environments and lithologies and a range of different diagenetic environments all operating at specific depths of burial, there is considerable scope for research into the mechanics of autocompaction in intertidal environments and sediments. A convenient starting point in the deconvolution of the potentially complex autocompaction behaviour in these environments is the depositional surface. This is the initial boundary condition from which all sediments begin to autocompact, and to which all sea level index points must be decompacted. Indeed, relying on the principle of uniformitarianism, contemporary sediments are the modern analogue to fossil sediments and will provide clues to their early stage autocompaction behaviour. Early stage autocompaction is also of particular interest because it is contemporary and recent sediments that are subjected to the dynamics of subaerial and vadose zone processes; it is in this geotechnical and diagenetic sub-environment that substantial deviations from Terzaghi's theory may occur.

The complexities associated with predominantly organogenic sediments have been discussed in Section 3.9.3. In such sediments, volumetric reduction is equally likely to be caused by biochemical degradation of organic compounds; effective stress variations, although still a controlling factor in the volume of peats, are potentially a secondary control. To allow a sufficient level of detail and quality in the datasets produced regarding autocompaction processes and their controls, it is therefore necessary at this initial inductive stage to focus the research in this study on predominantly mineralogenic materials. Such materials are defined arbitrarily as those composed of less than 50 % organic matter. The autocompaction behaviour of mineralogenic materials is likely to be dominated by variations in effective stress. Following the discussion of existing theories of autocompaction and the complexities of the intertidal environment throughout this chapter and with the research sufficiently focused to concentrate on near-surface, mineralogenic intertidal materials, improvements to current understanding of the autocompaction behaviour of these sediments will be made by testing the following hypotheses:

1. Mineralogenic intertidal sediments show no variability in structure and/or compression behaviour.
2. Existing geotechnical laboratory methods sufficiently represent intertidal field conditions.
3. Near-surface mineralogenic intertidal sediments are normally consolidated.
4. Terzaghi's compression law and consolidation theory are applicable to mineralogenic intertidal sediments.

5. Primary consolidation is the principal settlement process in mineralogenic intertidal sediments; creep is unimportant.
6. Diagenetic changes are unimportant and do not significantly modify mechanical compression behaviour.

The following chapter introduces the field site and outlines the field and laboratory techniques used throughout the experimental phase of this study to test these hypotheses.

Table 3.1 The main assumptions of the application of Terzaghi's Compression Law and the reasons why these assumptions are not met in intertidal areas.

Assumption of Terzaghi's Compression Law	Reason(s) for Violation of Assumption	Reference(s)
Materials are lithologically homogenous saturated clays of low organic content	Range of materials present: clays, silts, sands, gravels and peats in varying proportions. Peats may be in various states of decay, from undecomposed to fully humified. Rapid lateral and vertical variation in lithology. Sediments may be unsaturated.	Allen (2000 b); Head (1988); Hackney and de la Cruz (1980); Lillebø *et al.* (1999); Smith (1985); Tovey and Yim (2002); Hawkins (1984).
Voids ratio is a unique function of effective stress	Geological timescale results in potential for operation of time-dependent creep processes; organic intertidal materials are particularly prone to creep and so effective stress is not the sole control on material compression.	Head (1988); den Haan *et al.* (1994); Bjerrum (1967).
Primary consolidation results from loading by superincumbent material only; sediments are normally consolidated	The diurnal flooding/exposure conditions of the intertidal environment create a range of processes that result in consolidation; tidal loading, groundwater changes, desiccation, desaturation etc. Longer-term sea level changes ('regressions') and anthropogenic land-claim are also likely to overconsolidate the materials due to tidal-/ground-water variation, clay dispersion, prolonged exposure/desiccation etc. Creep also creates an overconsolidation effect.	Powrie (2004); Marinho and Chandler (1993); Hawkins, (1984); Fredlund and Rahardjo (1993 a, b); Crooks (1999); Greensmith and Tucker (1971 a, b; 1973); Barras and Paul (2000).
Post-depositional material modification is unimportant	Death of vascular plants results in collapse of open peat structure and autogenic overconsolidation; organic sediments are subject to humification processes; diagenetic processes concentrate redox sensitive elements in specific stratigraphic locations, potentially resulting in overconsolidation and concretionary fabric strengthening.	DeLaune *et al.* (1994); Smith (1985); Lillebø *et al.* (1999); Cundy and Croudace (1995 a, b); Cundy and Croudace (1996); Thomson *et al.* (2002); Spencer *et al.* (2003).

CHAPTER 4: FIELD SITE AND RESEARCH METHODS

4.1 STUDY SITE

4.1 Study site location

The field site location for this study is Cowpen Marsh in the Tees Estuary, Cleveland on the northeast coast of England (National Grid Reference: NZ 500529; Figure 4.1). Cowpen Marsh lies on the west side of the Tees Estuary and the A178 road (Figures 4.1 and 4.2). North and south sea walls were built in 1740 and resulted in the conversion of 85% of Cowpen Marsh from salt- to fresh-water marsh (Sproxton, 1989). As a result, it is only the northeastern corner of Cowpen Marsh that remains actively tidal *via* Greatham Creek, which floods and drains through Seal Sands (Figure 4.1). It is therefore the area of the marsh between the sea walls (Figures 4.2 and 4.3) that is the focus of this study, which is concerned with autocompaction in actively accreting intertidal sediments.

Despite the intensity of the surrounding land reclamation and the associated agricultural and industrial development (Shennan and Sproxton, 1990; 1991), Greatham Creek itself is the least disturbed of the tidal channels within the Tees (Berry and Plater, 1998) and is located at sufficient distance (*c.* 4 km) from the effects of human disturbance, mainly in the form of dredging (Plater *et al.*, 2000), in the main Tees channel. Furthermore, Greatham Creek has been a Site of Special Scientific Interest (SSSI) since 1966 due to its international importance as a wintering site for migratory wildfowl and wading birds (http://www.english-nature.org.uk/Special/sssi/unitlist.cfm?sssi_id=1000036). Hence, public access to the active intertidal zone is restricted. This minimised general disturbance to the sediments and provided a safe environment for the long-term installation and use of field equipment such as piezometers and barometers.

4.2 General character of sediments and rationale for site selection

The Tees Estuary has a tidal range of 6.1 m (spring tidal range of 4.6 m) (Table 4.1). The sediments of the upper intertidal zone (i.e. elevations above mean sea level, 0.35 m OD) at Greatham Creek display elevational zonation, from a tidal flat setting at lower elevations through low and middle marsh environments up to a high marsh floral assemblage. The altitudinal and elevational ranges of each floral zone and the species present therein are displayed in Table 4.2. However, these ranges are based on point estimates; in reality zones are transitional, occurring over a *c.* 0.10 to 0.30 m elevation range. A variable 0.20 – 0.30 m 'cliff' marks the transition between mudflat and saltmarsh zones.

Table 4.1 Tide levels (m OD) for the Tees Estuary (Source: Admiralty Tide Tables, 2005).

Lowest astronomical tide (LAT)	Mean low water spring tide (MLWST)	Mean low water neap tide (MLWNT)	Mean sea level (MSL)	Mean high water neap tide (MHWNT)	Mean high water spring tide (MHWST)	Highest astronomical tide (HAT)
-2.85	-1.95	-0.85	0.35	1.45	2.65	3.25

Greatham Creek is an ideal location to study the autocompaction behaviour of mineralogenic sediments due to its representivity of similar low energy coastal deposits found throughout the seaboard of northwest Europe. The primarily mineralogenic sediments forming at Greatham Creek reflect the generalised lithofacies model of late Holocene sediments of northwest Europe presented by Allen (2003) on the basis of a forty years of research into low energy coastal sediments and stratigraphies in this region. Hence, conclusions obtained regarding autocompaction from the sediments at Greatham Creek can be considered broadly transferable to similar deposits throughout coastal northwest Europe.

In addition to its general representivity, locating the study at Greatham Creek also benefits from a wealth of knowledge generated by previous research into the area. In particular, the large database of contemporary microfossil sea level indicator distributions on the active saltmarshes and mudflats of Greatham Creek presented by Horton (1997; 1999), Horton *et al.* (1999) and Zong and Horton (1999) will be critical to application of the model.

Following analysis of the intertidal lithology gradient in relation to altitude/elevation, two sediment types have been selected for this study on the basis of their predominantly mineralogenic lithologies (full analysis of the material selection investigation is undertaken in Chapter 5). These are a low marsh material, for which the sampling altitude for analysis is 2.26 m OD, and a mudflat material, which is sampled from an altitude of 1.06 m OD. The locations of these sites within the intertidal zone at Greatham Creek are indicated on Figure 4.3. Photographs of the sampling sites are shown in Figures 4.4 (low marsh) and 4.5 (mudflat).

Table 4.2 Description of the contemporary vegetation at Greatham Creek and its zonation by altitude and elevation above mean sea level.

Floral zone	Vegetation	Approximate altitudinal range (m OD)	Elevation range (m above mean sea level)
Mudflat	Minerogenic mudflat substrate with no *in situ* growth of vascular plant species.	0.34 to 1.77	0.00 to 1.42
Upper tidal flat/ pioneer marsh zone	Mudflat substrate covered by a thin algal mat, with a scattered presence of *Salicornia europea* (1-5% coverage).	1.89 to 1.92	1.54 to 1.57
Low marsh	100% coverage of substrate dominanted by *Puccinellia maritima* (c. 50%) and *Salicornia europea* (c. 40%), with occasional *Aster tripolium*, *Limonium vulgare*, *Sueda maritima*, *Spergularia*, and *Plantago maritima.*	2.20 to 2.31	1.85 to 1.96
Mid marsh	Characterised by an increase in the dominance of *Sueda maritima, Aster tripolium* and *Limonium vulgare* and a corresponding decrease in *Puccinellia maritima.* Towards the transition to high marsh, further species are introduced, namely *Festuca rubra* and *Festuca* ovina.	2.31 to 2.49	1.96 to 2.14
High marsh	Dominated by *Elymus pycnanthus* accompanied by *Festuca rubra*, *Festuca ovina*, and *Limonium vulgare.*	2.49 to 3.25	2.14 to 2.90

4.2 Field Methods

4.2.1 Levelling

Locations of interest in this thesis were levelled to Ordnance Datum Newlyn (m OD) using a Leica TC1010 combined Electromagnetic Distance Measurement ('EDM')/theodolite 'Total Station'. Differences in altitude between the Total Station and a pole-mounted prism are calculated by the instrument on the basis of their time-of-flight separation and angular distance from the vertical. An Ordnance Survey (OS) benchmark on the A178 road provided an initial starting altitude (1.46 m OD). Intermediate temporary benchmarks were utilised when a clear line of sight was not possible. The distance between intermediate benchmarks never exceeded 100 m, since altitude precision is inversely proportional to the distance between the Total Station and the reflective prism.

The difference in altitude between locations of interest and the OS benchmark was recorded on both 'outward' and 'inward' legs, allowing the precision of field altitudes to be determined *via* a 'closing error'. The maximum closing error recorded was 0.016 m, which is within the acceptable levelling error recommended by Shennan (1982) (Table 2.2).

4.2.2 Sample disturbance and collection

Because samples obtained from the field vary in their degree of disturbance, the sampling procedure used depends on the laboratory test for which each sample is required. Various classes of sample have been recognised (BSI, 1981), ranging from high quality Class 1 samples to poor quality Class 5 depending on the properties which can be reliably determined (Table 4.3). The quality of sample depends upon the sampling method employed. Table 4.4 provides a summary of the applications, advantages and disadvantages of typical geotechnical field sampling equipment used in this study.

Table 4.3 Categories of soil sample based on quality. Class 1 and 2 samples are generally referred to as 'undisturbed', Classes 3, 4 and 5 as 'disturbed'. Source: BSI, 1981.

Quality	Application
Class 1	Index tests, moisture content, density, strength and deformation characteristics
Class 2	Index tests, moisture content, grading, density and remoulded strength in some clays
Class 3	Index tests and moisture content
Class 4	Index tests
Class 5	Strata identification only

Table 4.4 Summary of the applications, advantages and disadvantages of geotechnical field sampling methods employed in this study (Adapted from Weltman and Head, 1983).

Method	Application	Advantages	Disadvantages
100 mm diameter open-tube sampler (U100)	Firm to stiff clays, insensitive or stoney clays, clayey silts, some weathered rocks or weak rocks. Class 1 or 2 samples in suitable soils, otherwise 3 to 4 and possible 5 in weak rocks	Simple robust equipment, usually dynamically driven. Types with detachable liner convenient for sample storage. Provide a reasonably large sample. Inexpensive. Rapid. Widely accepted and used.	Friction on inside of tube produces sample disturbance, particularly if driven at speed by drilling equipment. Disturbed material at base of borehole passes into sampler. Accurate control of sampler penetration is difficult. Quality often dependent on care taken by driller. Produces disturbed samples in soft and hard soils.
		Effects of sample disturbance significantly reduced on near-surface samples when tube is driven by hand (i.e. no dynamically driven drill equipment used).	Time consuming and great care required when driven by hand.
Block Sampling	Retrieval of representative volumes of fissured or otherwise structured material, including weak and weathered rock for representative laboratory strength tests, or analysis of fabric such as stereograph plot of fissure orientation. High quality Class 1 samples obtained.	Large, orientated, representative sample in mechanically undisturbed condition.	Great care required. Time consuming. Can suffer from severe disturbance from stress relief.

Block sampling (Class 1 samples) provides the highest quality of undisturbed sample. This sampling method provides the best means of obtaining undisturbed samples for geotechnical tests, particularly in the (near-) surface sediments required for this thesis.

Block sampling involves the use of a 13,500 cm^3 (30 cm x 30 cm horizontal cross-sectional area by 15 cm depth) iron sampling frame (no lid or base) to laterally constrain the sediment. The lower edge of the frame has a sharp cutting edge to assist in insertion of the frame into the sample sediment. To begin the sample collection, this frame is initially placed on the sediment surface. The internal areal dimensions are then marked on the ground surface using a sharp knife/machete. The sampling frame is then temporarily removed and a trench around the marked area is carefully excavated, though with an additional surrounding 'margin' of 5 cm to prevent sample disturbance. The trench is approximately 20 cm deep to allow access to the base of the sample. The frame is then placed back on the marked area and slowly pushed into the ground surface; as the frame cuts deeper into the ground, the 5 cm sediment margin is gradually trimmed away. When the frame is at the required depth, the sample base is cut from the underlying sediment using a wire saw. The laterally-confined block sample is then further confined on its top and base with metal sheets to prevent vertical movement and disturbance. The now fully confined sample is immediately wrapped in several layers of 'cling-film'. In order to prevent the 'cling-film' from being torn, the sample was then sealed with strong 'gaffer' tape. Samples were stored in a refrigerator to prevent bacterial decomposition, and in their fully confined, air-tight condition to prevent sample disturbance due to stress relief and/or loss of moisture. Typical block samples are shown in Figure 4.6.

One block sample typically yielded ≤ 4 of samples for geotechnical testing. In addition to these tests, the trimmings and other unused parts of the block sample were used to determine the physical properties of the sediment.

For the high resolution voids ratio/density measurements using the non-destructive x-ray scanning technique developed by Been (1980), high quality (Class 1) undisturbed sediment cores were required, but the block sampling technique was not sufficient. Instead, plastic tubing (25 cm length, internal diameter 10.3 cm) with a bevelled lower cutting edge was vertically hand-driven into the sediment surface (Figure 4.7 – 4.9). Each tube was only driven to approximately 18 cm to prevent disturbance of the most recent sediments so critical to this investigation. A comparison of distances from the sediment surface to the top of the plastic tubing on both the interior and exterior of the tube allowed an estimation of any resulting compaction. These distances were equal in every core taken, indicating negligible vertical compaction.

To allow gentle removal of the core from the ground, a pit was dug around one side of the core and the base was cut from the underlying sediment using a machete (Figures 4.7 – 4.9). The cores were sealed with water tight end caps, sealed with several layers of 'cling-film' and protected against tearing using 'gaffer' tape. Samples were stored upright to maintain conditions as representative of those in the field as possible and to prevent unwanted material deformation. Again, samples were refrigerated until required.

Since the core removal process and machete use slightly disturbs the underlying sediment, cores obtained from greater depths were not taken from locations directly beneath overlying cores. Rather, a series of cores of 'overlapping' depths were taken adjacently in the same excavated pit. The overlapping nature of this core sampling method also provided a means of examining the magnitude of any lateral variation in the sediment structure.

For basic sedimentological and lithostratigraphic analysis (particle size analysis and organic content determination *via* loss on ignition), surface sediment samples of 100 cm^3 (100 cm^2 by 1 cm deep) were obtained using a sharp knife or trowel, immediately placed in air-tight bags to prevent moisture loss and were stored in a refrigerator.

4.2.3 Piezometer and local tide gauge installation

Both the groundwater depths in the low marsh environment and tidal water depths through time were monitored using piezometers (Van Essen 'Divers') with built-in data logging capacity. The piezometers record pressure as centimetres of water column above the pressure sensor at 4°C. A semiconductor sensor also measures and records the ambient groundwater temperature. This data is used by the internal computer of the instrument to compensate the water depth measurements for temperature, enhancing the accuracy of water depth readings.

The piezometers are calibrated by the manufacturer (Van Essen Instruments). This calibration procedure involves fully submerging the instrument in distilled fresh water (specific gravity of 1.00). The temperature of the bath is stabilised at two different calibration temperatures (15°C and 35°C) at which the water pressure is varied. Pressure variations consist of increasing and decreasing pressures of 10, 30, 50, 70 and 90% of the useable pressure range. The maximum calibrated range of both instruments is 10 m of water column.

To monitor groundwater depths, a piezometer was installed into the low marsh surface (2.26 m OD). This was done by hand-augering a 10 cm diameter borehole into the ground surface to a depth of 2 m. In order to prevent the borehole from collapsing, a 2 m long, 5 cm diameter PVC well casing was inserted into the borehole. The bottom 1.5 m of the casing was vertically 'slotted' to allow free movement of water in and out of the well casing, which itself was wrapped in a geotextile filter to prevent the piezometer well from filling with sediment. The base of the well casing was sealed with a water tight cap. Once the well casing was in place, the surrounding gap in the borehole was filled with a coarse-grained sand to act as an additional filter and to hold the well casing securely in place (Figure 4.10). The top *c.* 5 cm of the well casing was further secured in place by a bentonite seal. The piezometer was suspended to a known depth/altitude in the well casing by a stretch-proof stainless steel wire attached to the removable well cap, the top of which was level with the ground surface. The watertight well cap and the bentonite prevented rainwater, surface runoff and tidal waters from entering the piezometer well directly.

The local tidal gauge was installed as low in the intertidal frame as was logistically possible (surface altitude of -1.045 m OD, approximately 0.2 m below Mean Low Water Neap Tides, Admiralty Tide Tables, 2005) in order to record as much of the total tidal range as possible. Although the lowest spring tides were not therefore recorded, it is flooding frequency and magnitude above Mean Sea Level that are of primary interest in this study. A 50 cm long, 5 cm diameter PVC pipe (non-slotted) with a water tight cap on the base was driven into the lower mudflat sediments near to the permanent (sub-/non-tidal) Greatham Creek channel. Six 3 mm holes were drilled through the top cap to ensure hydraulic 'connection' with tidal waters. Stainless steel wire was then tied to the top cap through these holes and the piezometer was suspended in the PVC pipe to a known depth/altitude.

The overlying water depth measurements recorded by the instrument are absolute and so each depth measurement is the sum of the water pressure and the barometric pressure. In order to compensate for the barometric pressure, a self-logging barometer (Van Essen 'BaroDiver') was also installed at Greatham Creek in a location that ensures that the tip of the instrument is never submerged. The barometer is also equipped with a temperature-corrected pressure sensor to reduce the influence of temperature on pressure readings. Barometric compensation is a simple matter of subtracting the barometric pressure from the water pressure for a particular time-referenced reading.

The piezometer used to measure and record the depth of the groundwater beneath the low marsh surface is accurate to ± 1 cm and precise to the nearest 2 mm. The Diver used as a tidal gauge is accurate to ± 1 cm and is precise to the nearest 1 mm. Before being installed in the field, the validity of these figures was tested in the laboratory by lowering the piezometer into columns of water of known depths in a measuring cylinder. In each case, the (barometer compensated) output readings were identical to the observed depths of water. Further tests of the accuracy of the piezometer were taken on a monthly by using a graduated measuring ruler as a 'dipstick' to manually observe groundwater levels.

The piezometers were connected directly to a 'ruggedised' laptop PC for programming (a sampling interval of 5 minutes was selected) and for downloading results. Such data downloads took place at fortnightly intervals. These field visits were also used to remove any sediment from the PVC tubing in the lower mudflat tidal gauge.

4.3 Laboratory methods

4.3.1 Sedimentological and lithostratigraphic analysis

The lithological characteristics of the sediments in this study were characterised using two main methods: particle size analysis and organic content.

For particle size analysis, sub-samples of wet mass between 0.5 g and 1.5 g (depending on the organic content) were digested in 20 ml of 20% hydrogen peroxide (H_2O_2), an oxidising agent, in a hot water bath (*c.* 80°C) for 2 to 4 hours. When all organic content was removed (when no further gas was evolved from the sediment), which may have required repeat application of the oxidation procedure if sediments were particularly organic, the sediments were washed with distilled water and centrifuged at 4,000 rpm for four minutes to remove the hydrogen peroxide. The supernatant liquid was decanted, 20 ml of distilled water was added to the sample along with 2 ml of aqueous sodium hexametaphosphate (3.3 wt %) buffered with sodium carbonate (0.7 wt %). Sodium hexametaphosphate is a dispersant that causes the separation and dispersion of flocculated particles. The sample was then examined using a Coulter LS 230 laser granulometer with PIDS (polarisation intensity differential scattering). This equipment generates highly reproducible grain size distributions over the size range 0.04 – 2000 μm. During loading, samples were sonicated to further aid particle dispersion.

Organic content was determined by loss on ignition. This technique involves firstly oven-drying of a sediment sub-sample to constant mass (typically 24 hours at 105°C). This sediment is then placed into a crucible of known mass. Crucibles are then placed into a muffle furnace at 550°C for 4 hours. During this time, organic matter is combusted to ash. The loss on ignition is then calculated using the following equation:

$$LOI_{550} = \frac{(DW_{105} - DW_{550})}{DW_{105}} \times 100 \tag{4.1}$$

where:

LOI_{550} = loss on ignition at 550°C (%)
DW_{105} = the dry mass (g) of the sample before combustion
DW_{550} = the dry mass of the sample after combustion (g).

The loss of mass on ignition is proportional to the amount of organic carbon contained in the sample (Dean, 1974). Sample weights are all accurate to 0.001 g. Between stages, samples and crucibles are stored in a desiccator to prevent uptake of moisture.

Although considered to be a simple and reliable method of estimating the organic content of a soil, Heiri *et al.* (2001) presented results which suggest that factors such as duration of heat exposure, the position of crucibles in the furnace, sample size and the idiosyncrasies of the laboratory undertaking the test can all have an influence on results. Accordingly, the time of heat exposure was kept constant for every sample (four hours) and dry sample weights were kept constant (5 g, although this was not always possible due to samples possessing different moisture contents). Furthermore, Heiri *et al.* (2001) recommend that small differences in loss on ignition should not be over-interpreted to allow for variations caused by unavoidable errors (a comparison of results from different laboratories following a standard method yielded a maximum error of approximately 2%).

Further, semi-quantitative analyses of lithology are undertaken throughout the study according to the Troels-Smith (1955) scheme of sediment description. The Troels-Smith scheme is based on the semi-objective classification of unconsolidated sediments (Long *et al.*, 1999). The scheme is independent of any knowledge of depositional processes and provides a structured approach that enables direct comparison of results collected by different investigators. Further evaluation of the approach is provided by Schnurrenberger *et al.* (2003). A summary of the Troels-Smith (1955) classification procedure is presented in Appendix I.

4.3.2 Physical property tests

Standard tests were used to determine the physical properties of sediment samples. Physical property data are required for description and classification of the sediments and for use in calculations that describe the structural state of the soils, such as the voids ratio.

The physical property tests undertaken in this thesis, their purpose and the test method employed are presented in Table 4.5. The exact details of the appropriate methods are outlined elsewhere (BSI, 1990; Head, 1980) and will not be repeated here.

Table 4.5 Details of the physical property tests undertaken in this study, their general definitions and the test method employed.

Physical Property	Definition	Test Method
Moisture content	The amount of water contained within a soil mass. Moisture content is expressed as a percentage of the dry mass of the soil.	Oven-drying method (110°C): BS 1377: Part 2 (1990): 3.2
Specific gravity	The ratio of the mass of dry solids to the mass of distilled water displaced by the dry solids.	Density bottle method: BS 1377: Part 2 (1990): 8.3
Bulk density	Mass of bulk soil, including solid particles, air and water, per unit volume. Expressed in g/cm^3.	Linear measurement method: BS 1377:Part 2 (1990): 7.2
Atterberg limits	A primary form of classification for fine-grained cohesive soils. The liquid and plastic limits define the moisture content boundaries between liquid, plastic and semi-solid states (in terms of decreasing moisture content and volume).	Liquid limit: cone penetrometer method: BS 1377: Part 2 (1990): 4.3 Plastic limit: BS 1377: Part 2 (1990): 5
Voids ratio	The ratio of the volume of voids to the volume of solids.	Height of solids method. Head (1980)

4.3.3 Geotechnical testing

Geotechnical consolidation testing was undertaken in two key apparatus: the oedometer compression apparatus (Figure 3.5) and the back-pressured shear box (Figures 4.11 – 4.18). Both of these apparatus laterally confine the soil by either the steel cutting ring (oedometer) or the sample frame vessel (back-pressured shear box). By preventing horizontal strains, the K_0 condition is achieved, where K is the earth pressure coefficient: the ratio of horizontal (σ'_h) to vertical effective stresses (σ'_v) (Garga and Mahbubul, 1991; Mesri and Hayat, 1993). In natural, undisturbed sedimentary systems in which zero horizontal strain is occurring, the subscript 0 (K_0) is added (Powrie, 2004). By preventing horizontal strains in these apparatus, one-dimensional consolidation is achieved, mimicking *in situ* ground conditions. K_0 conditions are not possible in triaxial cell apparatus without significant and often expensive hardware modifications (e.g. installation of a radial deformation gauge) and software systems. It was therefore decided that the oedometer and back-pressured shear box offer the best means of anisotropic, one-dimensional consolidation testing for this investigation.

Oedometer tests were carried out on fixed ring, front loading oedometers manufactured by Wykeham Farrance. Normal (vertical) load is applied by the application of weights to a loading lever arm. The steel cutting ring which laterally confines a sample is of 75 mm internal diameter and is 19 mm high. Settlement displacement was automatically monitored using LVDTs (10 mm travel) and logged using an Autonomous Data Acquisition Unit (ADU) and software (Ds7) manufactured by ELE International (Figure 3.5). The oedometer test procedure follows BS 1377 (BSI, 1990; Head, 1988).

The back-pressured shear box apparatus is a recently developed, custom-built geotechnical testing apparatus. The K_0 conditions it maintains and its hydraulic operation provide a unique opportunity to undertake dynamic (semi-diurnal) compression testing of soils. Such an approach is not possible in the oedometer apparatus due to its manual loading by slotted weights. The accurate computer control of vertical effective stresses allows field conditions to be exactly replicated. In the back-pressured shear box, a 100 mm x 100 mm x 20 mm (width x length x height) sample is located in a mounted sample chamber in a sealed pressure vessel flooded with de-aired, distilled water (Figure 4.11). The fluid pressure within this chamber can be varied, permitting control of the pore pressure within the sample. This fluid pressure is measured using two transducers, one mounted adjacent to the sample and the other on the fluid pressure control line. Normal load is applied through a hydraulically-controlled actuator. Sample settlement displacement is measured by an LVDT (10 mm travel) and automatically logged using computer software. The apparatus and the controlling software system (GDSLab) were manufactured by GDS Instruments Ltd. Dynamic loading was programmed on the basis of sine wave amplitude and period (as determined from field results).

The assembly procedure of the back-pressured shear box can be summarised as follows:

Figure 4.12 displays the components of the sampling vessel. The upper and lower sections of the sample vessel are assembled using the connecting screws. The lower porous disc is placed into the lower circular recess visible in on the lower section of the sample vessel. The sample cutting square is then lubricated with silicon grease and a specimen is carefully obtained from the main 'block' sample. The specimen and cutting square are then weighed and the pre-determined mass of the cutting square is subtracted to allow calculation of sample bulk density, moisture content and voids ratio. Filter papers, trimmed to the same dimensions as the specimen, are placed on the upper and lower sides of the specimen, which is then carefully extruded into the sample vessel (Figure 4.13).

The sample vessel is then placed and secured into the main back-pressured shear box apparatus (Figure 4.14). This consists of a lower section, and a removable upper section (Figure 4.14). The lower section contains the sample vessel platform, which itself is held securely in place by four locating screws and two externally-operated side supports (Figure 4.15). Once securely located on the sample vessel platform, the upper porous plate is placed on top of the specimen (above the upper filter paper) and a vertical 'spacer' is added to the sample vessel, from which the connecting screws have been removed. The upper section of the back-pressured shear box apparatus is then placed on the lower section; the load cell (Figure 4.16) fits directly onto the porous disc/sample and applies vertical stress. Twelve screws fasten the upper section of the apparatus to the lower section. A vertical beam is connected to the apparatus (Figure 4.17); this applies vertical stress to the load cell and is controlled by a hydraulic pressure controller. The LVDT is then fitted to monitor vertical displacement (Figure 4.17).

Once the apparatus was fully assembled (Figure 4.18), samples were fully saturated by initially flushing the sample/chamber with carbon dioxide for one hour to displace pore air, since carbon dioxide is more soluble in water than air. The sample chamber was then flooded with distilled water, fluid pressure transducers were reset to 0 kPa (under atmospheric pressure) and the LVDT reading was zeroed. The chamber fluid back pressure was then increased to 1000 kPa at a rate of 2 kPa per minute. Normal stress was increased at the same rate but with a differential pressure of 3 kPa to maintain positive effective stress. During the saturation stage, settlement displacement was continuously monitored.

Although a lower back pressure could have been used, this high value (1000 kPa) was used to allow full release of vertical effective pressures during the simulated 'tidal ebb'; pilot tests with low back pressures highlighted the fact that the normal load cell did not fully respond to decreases in effective stress, since the load cell itself is not physically attached to the loading yoke beam/normal load cell actuator. High back pressures ensured that, during periods of low effective stress, the load cell was fully lifted off the sample.

Throughout each test, the following parameters were logged at a rate appropriate to the test: time since start of test/stage (seconds), normal effective stress (kPa), back pressure (kPa) and settlement (mm). As in the oedometer tests, upon completion of the experiment, samples were carefully removed from the sample chamber, immediately weighed and dried for determination of initial and final moisture contents.

4.3.4 X-ray core scanning for density determination

Downcore density/voids ratio variations are indirectly measured using a non-destructive x-ray technique developed by Been (1980) on apparatus located in the consolidation laboratory of the Environmental Soil Mechanics Research Group in the Department of Engineering Science, Oxford University. The experimental apparatus is shown schematically in Figure 4.19. A highly collimated beam of x-rays is directed from an x-ray tube and collimator (Figure 4.20) through the sediment core (encased in plastic core tubing – Section 4.4.2) to a sodium iodide crystal and photomultiplier assembly (Figure 4.21). This produces an x-ray count rate, N, (counts per second) which can be related to the saturated sediment bulk density, ρ_b, through the following equation:

$$N = N_0 e^{-k\rho_b} \tag{4.2}$$

The parameters N_0 and k can be calculated from separate calibration samples which are made of the same plastic casing and sediment type from the main cores being analysed, since k depends on atomic number of the elements in the sample and plastic. Six calibration samples of uniform density contained within shorter (5 cm) sections of plastic tubing (10.3 cm internal diameter) were used (Figure 4.22). Calibration samples of densities from 1.0 g/cm^3 to 1.76 g/cm^3 were obtained from additional materials obtained from the field sampling locations. Since a number of lithologies were present in the cores, the calibration curve (Figure 4.23) was created from calibration samples of each lithology being analysed in the main cores. The fact that each calibration sample falls on the same calibration curve (Figure 4.23) indicates that individual calibration curves for each material were not required. Lower density calibration samples were determined by mixing mudflat samples with varying amounts of distilled water (Figure 4.24). One calibration sample consisted solely of distilled water (density = 1 g/cm^3). For each test, low density calibration samples were remixed to maintain uniform densities.

The x-ray tube, collimator and detector are mounted on a single arm to keep their relative positions fixed. This arm is part of a free-standing rig. An electric-driven lead screw moves the arm vertically up and down at speeds set by a sliding potentiometer. Height of the arm relative to a datum is determined to the nearest millimetre with a potentiometer geared to the lead screw. The accuracy of the density calculation depends on the traverse speed of the x-ray and counting assembly. A rate of 1 mm/second was used, providing an accuracy of the order of $\pm$ 0.002 g/cm^3 with a spatial resolution of c. 1 mm.

Since exposure to x-rays can be extremely dangerous, the room which contains the x-ray equipment was fully enclosed by lead shielding and concrete bricks during experiments to prevent radiation leakage. The x-ray tube could only be operated from outside the room with the door shut. Therefore, all equipment controls were outside the shielded room (Figure 4.25). Count rate and height parameters were recorded using the LabVIEW software package.

The x-ray and counter apparatus were positioned relative to the sediment core and calibration samples (which were placed on top of the core) with care to ensure that the fixed geometry of the density system was maintained (Figure 4.26). A typical x-ray session consisted of an upward followed by a downward sweep of the core/calibration samples to generate a profile of count rate against height. The data of each sweep can then be compared to determine whether there were any unwanted major shifts in x-ray output voltage caused by 'spikes' in the electrical mains supply. Even if no such shifts take place, small variations in count rate, by comparison with those caused by changes in sediment bulk density, still occur because of random and inevitable fluctuations in the x-ray output and photomultiplier assembly. A mean value of the upward and downward sweeps is therefore taken and a standard error is calculated. This standard error and the systematic accuracy error mentioned above ($\pm$ 0.002 g/cm^3) are combined to describe a total error using a root squared error calculation.

Using the calibration equation from Figure 4.23, the count rate, *N*, for each core sample can be converted to density (g/cm^3). Voids ratio, *e*, can then be calculated using the following equation:

$$e = \frac{G_s - \rho_b}{\rho_b - G_w} \quad (4.3)$$

where G_s is the specific gravity of the soil particles, ρ_b is the calculated bulk density and G_w is the specific gravity of the pore water (taken to be 1).

The non-destructive x-ray technique was preferred to a destructive linear (direct) measurement for determining density primarily on the basis of the spatial resolution of readings. The accurate high-resolution x-ray technique offers readings at c. 1 mm, compared to readings at 1 cm at best using direct measurement. Readings at 1 mm are of the same resolution as annual sedimentation rates and therefore may provide significant insight into (sub-)annual structural variability. Furthermore, this resolution allows the exact depths to be identified at which structural changes occur; in rapidly changing sequences this is of particular importance and is not possible with direct measurement.

Further details of the experimental set-up, the constraints and accuracies of the x-ray technique and its application to low stress consolidation of sedimenting soils can be found in Been (1980), Been and Sills (1981) and Sills (1998).

4.3.5 Geochemical analysis

Geochemical analysis was undertaken using X-ray fluorescence (XRF) spectrometry to obtain diagenetic element compositional data. XRF is a relatively rapid technique, providing simultaneous determination of major (SiO_2, TiO_2, Al_2O_3, Fe_2O_3, MnO, MgO, CaO, K_2O, Na_2O, P_2O_5, Cl) and trace elements (S, Cr, V, Ni, Cu, Zn, As, Bi, Pb, Ba, Rb, Sr, Y, Zr, Nb, Th, U, La, Ce, Ga, Mo, Sn, Sb, Br, I, Se) using a small sample size (3 g). Furthermore, unlike other geochemical analytical techniques such as inductively coupled plasma mass spectrometry (ICP-MS) or atomic absorption spectrophotometry (AAS), XRF analysis is relatively straightforward in terms of sample preparation since it does not require acid digestion before treatment and can analyse sediments in solid phase. XRF is also non-destructive in the sense that sediment samples are not consumed during measurement. This not only allows re-examination of element concentrations for determination of analytical precision, but it also means that samples can be used in other experimental procedures following XRF analysis (an important consideration in samples obtained from core sediments where sample abundance can be an issue). Although the detection limits are much lower with ICP-MS (parts per million with XRF, parts per trillion with ICP-MS), the diagenetic elements of interest in this study are naturally present at concentrations in excess of XRF detection limits (A. Cundy, *pers. comm.*). XRF methods are well-established and are further detailed in Croudace and Williams-Thorpe (1988), Croudace and Gilligan (1990) and Goudie (1998).

Samples were taken from sediment cores at 2 cm depth resolution, freeze-dried and ground to a fine powder in a ball mill. All samples were made into fusion beads for major element analyses and into pellets for trace element analyses. Samples were then analysed on an automatic sequential wavelength dispersive XRF instrument (Philips Magix-PRO WD-X-ray Fluorescence Spectrometer) at the Geosciences Advisory Unit, University of Southampton. Major element data are reported as % weight and trace element data are reported as ppm.

Accuracy was determined by analysing international standard reference materials (USGS MAG-1: Marine Sediment) and comparing the data obtained with published recommended data for the reference material. Both major and trace element concentrations were generally within 10% of such recommended data. For elements well above their detection limits, the precision of analysis is good: 1-2% relative standard deviation for major elements and 1-5% relative standard deviation for trace elements (Croudace and Cundy, 1995; Croudace and Williams-Thorpe, 1988; Cundy and Croudace, 1995a, 1995b; 1996).

4.3.6 Biostratigraphic (microfossil) analysis

The microfossil group used for quantitative palaeoenvironmental analysis in this study is foraminifera. This is because of the extensive dataset of contemporary altitudinal distributions of foraminifera at the selected study site (Greatham Creek) previously developed by Horton (1997; 1999). This annual dataset is based on fortnightly counts of surface foraminiferal assemblages throughout the intertidal zone at Greatham Creek. It therefore provides valuable insights into seasonal fluctuations in the distributions of foraminifera and determines the relative merits of life, death and total assemblages for use in palaeoenvironmental reconstruction. Furthermore, the altitudinal zoning of the foraminiferal assemblages confirmed by Horton (Horton, 1997; 1999) permits the development of quantitative transfer functions to allow reconstructions of reference water levels.

The issue of live, dead or total foraminiferal assemblages remains contentious (Murray, 2000; Scott and Medioli, 1980a). Some argue that total assemblages most accurately represent general environmental conditions because they integrate seasonal and temporal fluctuations (de Rijk, 1995; Scott and Medioli, 1980b). However, total assemblages combine data on living assemblages (which have not experienced taphonomic change) with dead assemblages (which have been taphonomically modified). Murray (1991; 2000) and others (Horton and Edwards, 2005; 2006) argue that the live component is variable and may not be transferred into sub-surface environments, therefore degrading the utility of the dataset. Therefore, foraminiferal data are expressed as a percentage of dead assemblages (following Horton, 1999). Furthermore, the training sets used in the transfer function of Horton and Edwards (2006) employ dead assemblages.

To differentiate between live and dead foraminiferal assemblages, the protein stain 'Rose Bengal' can be used (Murray, 1991). Rose Bengal has been used extensively to differentiate living from dead foraminifera (e.g. Horton, 1999; Scott and Medioli, 1980a; Scott *et al.*, 2001). Protoplasm is stained red while test walls are either unstained or lightly stained. Tests with protoplasm in the outer chamber are assumed to be live.

For foraminiferal analysis, 5 cm^3 sub-samples were taken from cores at specific depth intervals. In cores taken from the saltmarsh areas, foraminifera were sampled at 2 cm intervals due to the rapid lithological variation in these cores. In mudflat sediments where lithology was visibly uniform, a 4 cm sampling interval was used. These samples were then wet sieved through 500 μm and 63 μm sieves. The >63μm/<500μm fraction was retained. The >500μm fraction was examined for foraminifera before being discarded. Samples were stored in distilled water; a carbonate chip was added to prevent dissolution of any calcareous foraminifera present. Rose Bengal was also added to determine if any infaunal species were present; if so, these were noted but not reported since their use may result in anomalous reference water level reconstructions. Samples were counted wet under a binocular microscope at typical magnifications of x40 to x50. From each sample, all dead foraminifera present in the sample were counted. If after counting this initial sub-sample, a minimum of 100 dead foraminifera (the minimum number of counts considered necessary for reliable reconstruction - Horton and Edwards, 2006) had not been obtained, a further 5 cm^3 sub-sample was taken from the relevant section (depth) of the core and counted. Foraminiferal taxonomy follows Horton and Edwards (2006) and Murray (1979) and references therein.

4.3.7 Dating methods

Although radiocarbon dating is typically employed in dating coastal sequences and sea level index points, it is unsuitable for looking at changes in sedimentation rate on shorter (twentieth century) timescales in view of the high standard errors associated with even recent dates (perhaps ± 20 years), problems arising from atomic weapons testing in the 1950s and the preceding Suess effect (combustion of 'dead' carbon) (e.g. Pennington *et al.*, 1976). An alternative approach is the use of radionuclides with shorter half-lives than ^{14}C. Two widely used radionuclides have been used to develop age-depth models in this study: lead-210 (^{210}Pb, half-life 22.26 years) and caesium-137 (^{137}Cs, half-life 30 years).

^{210}Pb is a natural radioactive decay product, part of the uranium-237 (^{238}U) series. ^{210}Pb reaches sediments in two forms: as a 'supported' (background) component arising from the presence in the sediment of its parent isotope, radium-226 (^{226}Ra), which precedes ^{210}Pb in the decay series, and as an 'unsupported' rainout component derived from atmospheric ^{210}Pb. In order to derive dates from ^{210}Pb analysis, it is necessary to determine both the total ^{210}Pb component and the supported activity. Subtraction of the supported component from the total ^{210}Pb activity allows calculation of the unsupported ('excess') ^{210}Pb content in the sediment. This declines with depth in accordance with a negative exponential function derived from the short half-life of the radioisotope.

^{137}Cs is an artificial radionuclide introduced into sedimentary environments by atmospheric fallout from nuclear weapons testing (maxima in 1958 and 1963). Locally, within northwest Europe, deposition of ^{137}Cs resulting from the Chernobyl accident in April 1986 has provided an additional opportunity for dating. However, Croudace (*pers. comm.*) has stated that the influence of Chernobyl in the UK is small, particularly in northern areas where discharges from Sellafield may obscure any Chernobyl-related influence. Hence, two ^{137}Cs sub-surface activity maxima have been used as marker horizons to date the cores and from which sediment accumulation rates may be derived: the 1963 weapons testing peak and the ~1980 Sellafield discharge maximum.

Samples were taken from sediment cores at 4 cm depth resolution, freeze-dried and ground to a fine powder in a ball mill. Radionuclide activity determination was undertaken at the Geosciences Advisory Unit, University of Southampton. ^{210}Pb activity was determined by a proxy method through alpha spectrometric measurement of its granddaughter nuclide ^{210}Po. The method employed is based on Flynn (1968), using double acid leaching of the sediment with ^{209}Po as an isotopic tracer and autodeposition of the Po isotopes onto Ag discs. Detection limits are 0.01 Bq/kg. The excess ^{210}Pb activity was determined from ^{226}Ra activities using high resolution gamma spectrometric analysis using a Canberra P-type HPGe gamma ray spectrometer (excess ^{210}Pb = total ^{210}Pb - ^{226}Ra). Similarly, ^{137}Cs activity was determined using gamma spectrometry. Count times of 80,000 seconds were used and errors were typically 4% (1σ). Detection limits of 0.04 Bq/g were obtained.

4.4 SUMMARY

This chapter has outlined the methodological approach that this thesis will adopt to meet the objectives of this thesis. It has also introduced the field study site and discussed the field and laboratory techniques that will be used. Such techniques are drawn from different, yet related, subjects, including geomorphology, soil mechanics and palaeoenvironmental reconstruction, and reflect the multidisciplinary nature of this study. The data obtained using these techniques are presented and analysed in the following four chapters. The overall research framework is summarised as a flowchart in Figure 4.27; this figure also relates the main information sources used in this study to the research hypotheses set out in Section 3.10.

CHAPTER 5: THE CONTEMPORARY GEOTECHNICAL ENVIRONMENT AT GREATHAM CREEK

5.1 CONTEMPORARY MATERIALS IN THE INTERTIDAL ZONE AT GREATHAM CREEK

In order to select materials suitable for use in this study, it was firstly necessary to classify the full range of materials present in the contemporary upper intertidal zone at Greatham Creek. To determine variations in surface sedimentology, a transect was taken between mean sea level (MSL, 0.35 m OD – Table 4.1) and highest astronomical tide level (HAT, 3.25 m OD) at approximately 0.04 - 0.05 m altitude intervals (a total of 69 samples), which sampled all marsh floral zones (Table 4.2). The sedimentological transect allows the specific conditions of the intertidal zone at Greatham Creek to be related to the generality of northwest European saltmarshes displayed in Figure 2.1. Sedimentology/lithology is described on the basis of loss on ignition (to determine the organogenic component) and particle size analysis. These data are presented as the mean of three determinations. For clarity, standard errors are not presented graphically; their inclusion did not alter any of the observed trends in surface lithology.

5.1.1 Duration of tidal submergence

The data provided by the local tide gauge installed at Greatham Creek were analysed to allow observed variations in sedimentology to be considered in terms of the duration of tidal flooding (the hydroperiod). For each tide, the duration of tidal submergence at 5 cm altitude intervals along the transect was calculated (Figure 5.1). These durations were then summed for each month of observation and divided by the total time in each month to calculate monthly flooding durations, expressed as a percentage (after Gehrels *et al.*, 2001). These monthly data were then summed for each altitude to provide annual flooding durations, which again are expressed as a percentage of total time (November 2003 – January 2005). The relationship between altitude and flooding duration is curvilinear in graphical form (Figure 5.2). MSL is flooded for 50 % of total time. As altitude increases, this value decreases at a generally constant rate until approximately 2.4 m OD, which is submerged for *c.* 5 % of total time. Above this altitude, and particularly between mean high water spring tide (MHWST) level and HAT level, flooding duration decays asymptotically. At these levels, high in the intertidal frame, the saltmarsh surface is rarely submerged (< 2 % of total time, equivalent to brief time periods of less than two hours on one or two tides per month).

5.1.2 Organic content variation

Figure 5.3 displays the variations in loss on ignition in relation to altitude (m OD), flooding duration (as a percentage of total time), reference water levels, approximate marsh floral zones and CONISS clusters (Section 5.1.4). Three distinct zones can be identified. The first of these, between 0.34 m OD and approximately 1.75 m OD, is characterised by relatively low and constant loss on ignition values (*c.* 15 %). This area represents the mudflat environment and is broadly indicative of altitudes between MSL and mean high water neap tide level (MHWNT, 1.45 m OD) where annual flooding duration is less than 25 %. Above c. 1.75 m OD, loss on ignition values increase from 15 % to approximately 50 % at MHWST level (2.65 m OD). Although a rising trend in loss on ignition values is evident, there is considerable scatter in the data; sediments of 50 % organic content are altitudinally juxtaposed with those of 17-18 % at c. 2.30 m OD, for example. This zone of increased loss on ignition broadly coincides with low and mid marsh environments and the scatter reflects the transitional nature of changes in marsh zone (Table 4.2). The overall rising loss on ignition trend within the saltmarsh zones of Greatham Creek conforms to the general model of northwest European marshes displayed in Figure 2.1. Pioneer, low and mid marsh environments are submerged for approximately 10 %, 5 % and 2.5 % of total time respectively.

At MHWST level, flooding duration decreases to less than 2 % of total time. Above 2.75 m OD, flooding rarely occurs (less than 0.7 % of total time). In the high marsh floral zone, loss on ignition values increase consistently and with less scatter than in the low/mid marsh zone to a maximum value of *c.* 85 %.

Flooding duration exerts a greater control on organic content (loss on ignition) than altitude (Figures 5.4 and 5.5). Linear regression analysis between loss on ignition and altitude yields an r^2 of 0.66. However, where the data are split into two visually observed 'zones' (a mudflat zone of low and constant loss on ignition; and a marsh zone, characterised by a linear rise in loss on ignition values), it can be seen that altitude exerts no control on loss on ignition values in the mudflat environment at Greatham Creek (r^2 = 0.01). This result is unsurprising given the absence of vegetative growth in mudflat environments. In contrast, the regression analysis on the marsh deposits reveals a strong altitudinal control on organic content (r^2 = 0.82). However, organic content is more likely to be determined by flooding duration (% of total time) (Figure 5.5); the two variables are related by a single logarithmic function (r^2 = 0.91 – a value higher than all of the r^2 values for each 'segment' of linear regression between loss on ignition and altitude shown in Figure 5.4). Again, this is logical since duration of flooding is amongst the dominant controls on the vertical zonation and productivity of vascular plant species (Silvestri *et al.*, 2005). Furthermore, decreased flooding frequency and duration also reduce opportunities for mineralogenic deposition, hence increasing the proportion of organic material in the sediment.

5.1.3 Particle size variation

Figure 5.6 displays variations in the relative abundance of the sand, silt and clay fractions (expressed as a percentage of the total minerogenic component) with respect to altitude, flooding duration, reference water levels, approximate marsh floral zones and CONISS clusters. At all altitudes, silt is the dominant particle size fraction. In mudflat environments, the silt fraction constitutes between 65 and 75 % of the clastic sediment. This value rises to 75 – 80 % in pioneer, low and mid marsh environments, before dropping to around 50 % at HAT level. This fall in silt content is accompanied by a corresponding increase above MHWST level in the sand fraction from less than 5 % to *c.* 45 %. Above MSL and below MHWST level, the sand fraction displays a generally inverse relationship with altitude. It is generally lower than 10 %, despite occasional peaks, such as at the transition from mudflat to pioneer marsh. Lower mudflat sediments are characterised by a sand content of 10 %. This value begins to decrease above *c.* 1.3 m OD to generally lower values (0 – 10 %) in the low and mid marsh sediments.

Clay content is more variable in the mudflat sediments, ranging between 20 and 35 %, but it is more stable in low and mid marsh environments (*c.* 20 %). Above MHWST level, clay content decreases from 25 % to less than 5 % at HAT level.

Differences in particle size distributions are generally associated with kinetic energy variations. As tidal water levels flood surfaces at higher altitudes, water depths and velocities decrease and so too do kinetic energy and turbulence levels (Möller *et al.*, 1999). When kinetic energy drops below a critical threshold for a particular particle size and density (the settling velocity), sedimentation will occur. This phenomenon can be inferred from Figure 5.6; sand content gradually decreases with increased altitude and the associated shallower flood-water depths and lower kinetic energy. However, the reversal of this trend in high marsh environments towards HAT level may result from the effects of high energy storm events that transport coarser sediments to altitudes which would not normally be submerged by high energy, sand-laden flood waters. The infrequency and low energy nature of the subsequent astronomical spring tides that do flood the highest parts of the upper intertidal zone means that these (relatively) coarse sediments are unlikely to be re-suspended and removed. Additional coarser-grained inputs may result from aeolian activity.

The silt fraction does not display any such gradual variation in relative abundance with respect to altitude. Rather, silt content seems to show a 'zonal' relationship with marsh vegetation zones with a higher abundance of the silt fraction in saltmarsh sediments in relation to that of the unvegetated mudflat. This relationship exists because vegetation provides an important control on flow within the vegetation canopy by increasing surface roughness and frictional drag. This decreases mean flow velocity and inhibits the production of turbulent eddies (Christiansen *et al.*, 2000; Leonard and Luther, 1995), leading to a dissipation of tidal wave energy (Möller *et al.*, 1999) and promoting particle settling.

Beierle *et al.* (2002) advocated the use of surface plots of particle size distributions. This method of visualisation requires data to be ordered as x, y and z values (particle diameter, altitude and volume percent, respectively). It represents looking 'downwards' on a series of particle size curves. Surface plots present the raw particle size data in their entirety, allowing the shapes of complex, polymodal particle size distributions to be observed. This also removes the emphasis on the arbitrary division between sand, silt and clay fractions since particle size is a continuous measure (Krumbein, 1934).

Figure 5.7 displays a surface plot of altitude against particle diameter (% of sample detected by each 'channel'/'bin' of the laser granulometer). It illustrates that mudflat, pioneer, low and mid marsh sediments have very similar particle size distributions, with minimum values around 0.1– 0.2 μm (clay) and maximum values of *c.* 150 μm (fine sand). However, the nature of the particle size distributions between these end values is different. In the mudflat sediments, a modal particle diameter of *c.* 30 – 40 μm (very coarse silt) is evident. In the pioneer, low and mid marsh sediments, the modal particle size shifts to a finer value of *c.* 10 μm (medium silt). At the transition from mid to high marsh, the sequence gradually fines further to a modal particle size of *c.* 6 μm (fine silt). Above an altitude of c. 2.70 m OD, the particle size distribution dramatically changes. The degree of sorting increases, indicated by the decrease in the range of particle sizes present in sediments at these altitudes (minimum value c. 2-3 μm – very fine silt; maximum value c. 150 μm – fine sand). Modal grain size coarsens to 90 μm (very fine sand). The surface plot (Figure 5.7) displays a clear zonation of particle size distributions in relation to altitude. Three distinct zones can be identified: a mudflat zone, a pioneer, low and mid marsh zone and a high marsh zone.

5.1.4 Unconstrained cluster analysis

Due to the continuous, transitional nature of sedimentological variation in the upper intertidal zone at Greatham Creek, splitting the sedimentological transect into homogenous material groups for the purposes of geotechnical testing on the basis of visual observation alone would be an arbitrary task. In order to numerically assign such groups, unconstrained cluster analysis was performed on the contemporary sedimentological data using the Constrained Incremental Sum of Squares (CONISS) program (Grimm, 1987; within TILIA, version 2.0, b5; Grimm, 1993). Cluster analysis, based on unweighted Euclidean distance and using no transformation or standardisation of the sedimentological percentage data, objectively groups the contemporary samples into more-or-less homogenous lithological zones ('clusters') (Horton and Edwards, 2006). Samples in different clusters tend to have maximum dissimilarity. Since the cluster analysis is 'unconstrained', clusters do not have to consist of altitudinally contiguous samples.

Since particle size is a continuous measure, multiple cluster analyses were performed using different divisions of grain size fractions using the modified Udden-Wentworth size scale (Blott and Pye, 2001). Loss on ignition remained constant during each cluster analysis. The output of the cluster analysis is an hierarchical dendrogram (Prentice, 1986). An example is displayed in Figure 5.8, based on eight grain size fractions and loss on ignition. Assigned clusters were found to be largely invariable using different grain size fractions. It is therefore evident that loss on ignition is the main cause of numerically-assigned material groups in the intertidal zone at Greatham Creek. The altitudinal boundaries between each cluster always overlapped.

From the dendrogram presented in Figure 5.8, 5 principal zones can be identified; these show a strong altitudinal zonation and are also displayed in Figures 5.3 and 5.6. Cluster 5 corresponds with the mudflat environment. Clusters 3 and 4 overlap considerably and represent the pioneer, low and mid marsh environments. Clusters 1 and 2 slightly overlap and are indicative of the high marsh environment. The overlapping nature of the clusters undoubtedly reflects the observed transitional boundaries between vascular plant floral zones noted in Section 4.2. These objectively determined clusters therefore reflect 'real-world' observations of marsh floral zone. Discrete lithological zones do not seem to exist within the intertidal zone, even on the basis of objective multivariate statistical analysis.

5.2 MATERIAL SELECTION

Due to the complications associated with predominantly organogenic facies, a decision was taken in Section 3.10 to focus only on mineralogenic sediments with organic contents of less than 50 %. All sediments at elevations below MHWST level have less than 50 % organic matter and hence meet the criterion to be considered in this study. However, in order to be tested in geotechnical laboratory apparatus (i.e. the oedometer and back-pressured shear box), sediments must be in a suitable condition; they must display cohesion to permit (a) undisturbed sampling from the main block sample and (b) trimming to fit perfectly within the oedometer cutting ring and back-pressured shear box cutting square. The highly organic mid- and high marsh samples consist of moist, poorly humified vascular plant remains with insufficient mineralogenic content to bind the biogenic sediment (Figure 5.9, a and b). It would not be possible to prepare a material for geotechnical tests in this unhumified, incohesive state without significant disturbance. Indeed, attempts to calculate a value of a structural parameter such as the voids ratio for these sediments would be meaningless. Although this sediment would undoubtedly undergo significant humification during burial that may lead to the production of a more 'cohesive' (and therefore 'testable') peat, the contemporary and recent sediments form only a thin veneer (less than *c.* 10 cm) covering the underlying fossil (i.e. more consolidated) clastic sediments and so an organic material in the 'required' state is not observed in the stratigraphy.

In light of the research focus and geotechnical sampling factors, high and mid marsh environments, and CONISS clusters 1, 2, and 3 are excluded from the material testing program. Similarly, since CONISS cluster 4 partially includes sediments unsuitable for use in this thesis, materials have been selected on the basis of the remaining vascular plant zones: mudflat (0.35 m OD – *c.* 1.77 m OD) and low marsh (*c.* 2.20 m OD to *c.* 2.31 m OD). Pioneer marsh was not tested in this investigation since these sediments can only be found in isolated, shallow tussock-like protrusions from the mudflat surface. Pioneer marsh sediments were dry, cracked and often friable, prohibiting undisturbed samples from being obtained. However, this does not invalidate the research undertaken in this study, since pioneer marsh forms only a narrow altitudinal zone at Greatham Creek. Furthermore, pioneer marsh sediments are essentially recently colonised mudflat sediments; with such low levels of *in situ* organic growth, the material properties of the mudflat sediment are likely to be largely applicable to the pioneer marsh zone.

In an effort to maintain as constant a lithology as possible for use in the geotechnical testing program, mudflat and low marsh samples were always obtained from the same altitudes. The mid-points of the altitudinal ranges in which mudflat and low marsh sediments are found were chosen as sampling altitudes since these are furthest removed from the transitional areas at the edges of each zone and are therefore taken to be the most representative of each sediment type. These altitudes are 1.06 m OD for the mudflat (flooding duration of *c.* 35 %) and 2.26 m OD (flooding duration of *c.* 6 %) for the low marsh. Photographs of each sampling site are displayed in Figures 4.4 and 4.5. Although this narrow altitudinal range may not reflect the full geotechnical environmental gradient, limiting the lithologies and sampling altitudes is necessary at this early stage of research into autocompaction to allow experiments to be undertaken to a satisfactory level of detail. Some of the implications of limiting the range of materials studied are discussed further in Chapter 8.

5.3 Physical Properties of Materials Selected for Geotechnical Testing

Figures 5.10 and 5.11 show photographs of the mudflat and low marsh materials respectively. A description of these sediments using the universal classification scheme proposed by Troels-Smith (1955) is provided in Table 5.1. These samples were analysed to determine the physical and geotechnical properties of the near-surface virgin material (Table 5.2). To gain insight into the variability in lithology and physical properties, multiple determinations were undertaken on each material type. Exact numbers of duplicates for each test are also displayed in Table 5.2. For the low marsh sediments, the top 0.5 cm was removed and disregarded, since this represented the live root mat that is unsuitable for geotechnical testing.

The materials are very similar in terms of their granulometric characteristics, with similar sand, silt and clay contents (Table 5.2). Silt content is essentially equal in the two samples (approximately 72 %). The mudflat sediments possess a higher sand content (mean = 15.21 %) than the low marsh sediments (mean = 12.17 %). There is also a lower clay content in the mudflat (mean = 12.44 %, compared to 15.81 % in the low marsh). This granulometric similarity is further illustrated in Figures 5.12 and 5.13. These ternary plots that compare the relative abundance of sand to the ratio of silt to clay contents indicate that both materials can be classified as sand-rich silts, though the lower sand content in some low marsh samples results in their classification as pure silts.

Table 5.1 Description of the selected upper intertidal sediments.

Sediment type	Troels-Smith (1955) analysis	Description
Mudflat	1011- Ag3Ga1As+	Light brown silt/clay with frequent bioturbation burrows and occasional iron staining. Very wet sediment with no air pockets visible in the soil structure.
Low Marsh	20210 Ag2 As1Th1[1] Ga+Sh+Lf+	Mid-brown organic silt with some clay and partly humified organic matter. Frequent *in situ* rootlets. Occasional iron staining/streaking. Obvious 'open' structure and visible air pockets present.

Figures 5.14 and 5.15 illustrate the particle size distributions of each sample from the two materials. Again, a definite degree of similarity can be seen between the two lithologies. Figure 5.14 shows that the samples are generally granulometrically similar, although the low marsh samples are finer - the cumulative percentage profiles plot to the 'left' of those for the mudflat. The low marsh material shows a gap-graded section at particle diameters of 30 - 40 μm (very coarse silt), indicating an absence of particles of these sizes within the samples. Both materials have a similar range of particle sizes present (maximum particle diameters of *c.* 200 – 300 μm) and both display polymodal particle size distributions (Figure 5.15). The mudflat samples generally have a primary modal particle diameter of 30 μm (coarse silt) and secondary modal peaks at *c.* 8 – 9 μm (medium silt) and 70 – 80 μm (very fine sand). The low marsh material is characterised by a primary modal peak of *c.* 8 – 9 μm (medium silt), with additional peaks at 20 – 30 μm (coarse silt) and between 60 and 200 μm (very coarse silt – fine sand). Despite an element of variation within in each material, these modes are clear and consistent in each sample analysed (Figure 5.15).

Figure 5.16 illustrates that there is generally very little variation in the clastic component both within and between each material type. The boxplots displayed show that the sand, silt and clay contents for both materials have similar median values (again, particularly for the silt fraction), and the small and similar ranges and inter-quartile ranges (and low standard deviations – Table 5.2) show that there is relatively little variation in and between the particle size characteristics of each material.

In contrast to the granulometric similarity of the samples, organic content (as revealed by loss on ignition data) differs greatly between materials (Table 5.2). The low marsh material has a generally higher mean value (24.68 %) than the mudflat (16.83 %). The low marsh also has smaller variability in this mean value, as revealed by Figure 5.17. The higher value in the low marsh reflects *in situ* vegetation growth; its low variability may reflect a tight altitudinal control on organic productivity. The loss on ignition values for the mudflat are higher than may be expected given the absence of organic growth in these environments. However, surface runoff from the saltmarsh transports considerable amounts of organic detritus to mudflats, increasing the (detrital) organic content here. Also, the high faunal activity in mudflat environments may lead to increased amounts of faecal matter, thus increasing loss on ignition values (G. Sills, *pers. comm.*). The variability of these two processes may account for the increased variability in loss on ignition values for the mudflat samples.

Specific gravities for the two materials had very limited ranges (Table 5.2) and so the mean values were used in geotechnical analyses. These are 2.48 for the low marsh and 2.63 for the mudflat. These values for specific gravity may seem lower than those of many 'conventional' engineering materials and can be explained by higher than 'normal' organic contents (Hobbs, 1986; Paul and Barras, 1999; Skempton and Petley, 1970).

Low marsh sediments have considerably higher natural moisture contents than the mudflat sediments (Figure 5.18). Low marsh materials have a mean natural moisture content of 163.80 % (standard deviation = 21.81 %), compared to a mean natural moisture content of 78.45 % (standard deviation = 10.54 %) in the mudflat sediments. This corroborates the visual observation (Table 5.1) that the low marsh sediments display an obvious open structure that is 'available' to be filled with greater amounts of water. Low marsh sediments also display a greater range of natural moisture contents.

Mudflat sediments display greater degrees of saturation than low marsh sediments (Table 5.2; Figure 5.19). Some mudflat samples were fully (100 %) saturated, whereas the maximum recorded degree of saturation in the low marsh sediments was 96.96 %. The minimum observed degree of saturation in mudflat sediments was 91.12 %, compared to a minimum of 85.16 % in the low marsh. In general, the degree of saturation in the low marsh (mean = 92.57 %) was lower than that in the mudflat sediments (mean = 96.27 %). Once again, these findings are consistent with the relative position of the low marsh and mudflat materials within the intertidal frame. The low marsh materials are flooded less frequently and for shorter durations than the mudflat sediments. This increases exposure to subaerial conditions, hence promoting desiccation and desaturation. Furthermore, the presence of vascular plants in the low marsh sediment has the effect of aerating the soil and enhancing the withdrawal of moisture from soil insterstices *via* evapotranspiration (Powrie, 2004).

The results of the Atterberg limit tests (Table 5.2) have been plotted on a plasticity chart (plasticity index plotted against liquid limit) (Figure 5.20). This allows the samples to be classified in terms of the British Soil Classification System (BSCS) (BSI, 1981). The A line provides an arbitrary division between silts ('M'; these plot below the A line) and clays ('C'; these plot above the A line). Five divisions of plasticity are recognised by the BSCS: low (L), intermediate (I), high (H), very high (V) and extremely high (E). From Figure 5.20, it can be seen that the mudflat material is classified as 'MI' – a silt of intermediate plasticity. The low marsh material is classified as 'ME' – a silt of extremely high plasticity.

The surface structural characteristics of the sediments are represented by the values for initial voids ratio, e_i (Table 5.2). Bulk density data are also presented to allow comparison with other studies. Initial voids ratios vary considerably between the two materials. The observed open structure of the low marsh material is again reflected in the high values of initial voids ratio. The mean value of initial voids ratio in the low marsh is 4.38 – more than twice the mean value of the initial voids ratio of the mudflat sediments (2.16). The low marsh also displays a greater range of initial voids ratios (Table 5.2), from a minimum of 3.69 to a maximum of 5.56. These values show a skewed frequency distribution; despite a close distribution of the interquartile range around the median value, there are occasional extreme values at the upper end of the range (Figure 5.21). The mudflat sediments show a lesser degree of structural variability; values of initial voids ratio are generally closer to the median value (Figure 5.21). The minimum observed initial voids ratio in mudflat sediments was 1.71 and the maximum was 2.68 (Table 5.2).

Table 5.2 Physical properties of contemporary low marsh and mudflat samples at Greatham Creek.

	Low Marsh					Mudflat				
	Mean	Min.	Max.	S.D.[1]	n*	Mean	Min.	Max.	S.D.[1]	n*
Loss on ignition (%)	24.68	23.45	27.05	1.20	36	16.83	14.11	19.37	1.57	30
Sand (%)	12.17	7.71	15.75	2.54	36	15.21	12.47	18.70	1.75	30
Silt (%)	72.01	68.58	74.78	1.66	36	72.35	69.66	74.62	1.27	30
Clay (%)	15.81	14.14	17.60	1.20	36	12.44	10.11	14.79	1.52	30
Specific gravity, G_s	2.48	2.48	2.49	0.00	3	2.63	2.61	2.65	0.02	3
Natural moisture content, w (%)	163.8	137.05	216.34	21.81	36	78.45	61.02	92.88	10.54	30
Degree of saturation (%)	92.57	85.16	96.95	3.70	12	96.27	91.12	100	2.58	10
Liquid limit (%)	96.71	96.59	96.83	0.17	2	48.99	48.79	49.19	0.28	2
Plastic limit (%)	56.31	56.06	56.56	0.00	2	28.94	28.64	29.25	0.43	2
Plasticity index	40.40	-	-	-	1	20.05	-	-	-	1
Initial voids ratio, e (H_s method)	4.38	3.69	5.56	0.49	12	2.16	1.71	2.68	0.33	10
Bulk density, ρ_b, at natural moisture content (g cm^{-3})	1.22	1.16	1.27	0.03	12	1.48	1.42	1.62	0.06	10
Bulk density, ρ_b, of saturated material (g cm^{-3})	1.26	1.20	1.31	0.04	12	1.47	1.42	1.62	0.09	10
Dry density, ρ_d (g cm^{-3})	1.22	1.16	1.29	0.04	12	1.47	1.41	1.60	0.07	10

[1] S.D. = standard deviation.
* n = number of samples upon which descriptive statistics are based.

5.4 INITIAL STRUCTURAL VARIABILITY

The higher loss on ignition values in the low marsh sediment (mean loss on ignition value of 24.58 %, compared to 16.82 % in the mudflat sediments analysed) reflect the presence of vascular plants which are known to create well-aerated, highly porous soil structures (Delaune *et al.*, 1994) that may be more prone to compression. In addition, the lower flow velocities on the saltmarsh (Christiansen *et al.*, 2000; Leonard and Luther, 1995; Möller *et al.*, 1999) during tidal flooding leads to slow deposition and an open random fabric (Burland, 1990). Sediment trapping by vegetation and the subsequent flaking of silt-rich crusts (Allen, 2000) may also assist in the creation of an initial open soil structure. In contrast, the absence of vegetation on the mudflat surface precludes the formation of an organogenic openly structured fabric. Initially denser structures are created by more rapid deposition rates from a denser suspension. This results in a more compact, oriented soil fabric (Burland, 1990).

Despite the general (altitudinal) control of organic content on initial structure, there is no within-material relationship between organic content and initial structure in the low marsh sediments; loss on ignition was not significantly correlated with the initial voids ratio (r = 0.674, n.s.). In the mudflat sediments, however, a significant correlation does exist between loss on ignition and the initial voids ratio (r = 0.793, p = 0.03) (Figure 5.22), suggesting that slight variations in detritral or algal organic content result in noticeable variations in structure. Within the low marsh sediments, no significant correlations existed between the initial voids ratio and sand (r = 0.269, n.s.), silt (r = -0.375, n.s.) and clay (r = -0.002, n.s.) contents. Unsurprisingly, given the granulometric similarity of the samples, there were also no significant correlations between initial voids ratios and sand (r = -0.408, n.s.), silt (r = -0.52, n.s.) and clay (r = 0.575, n.s.) contents in the mudflat samples analysed.

Within-sample variations in initial structure are likely to result from minor variations in depositional conditions. Spatial and temporal variations in water chemistry, flow velocity and density of sediment in tidal waters (Been, 1980; Been and Sills, 1981; Burland, 1990; Lintern, 2003; Sills, 1998) are likely to result in differences in floc size and soil structure. In addition, in the low marsh environment, variations in plant species assemblage and canopy height may result in variations in flow velocities, creating differences in settling style and sedimentation rate, creating rapid spatial differences in soil structure.

Whilst there is a general structural signature associated with each depositional environment (low marsh and mudflat) caused both directly and indirectly by vascular plant growth, random variations in depositional conditions cause within-sample variability in initial structure at the depositional surface.

5.5 Monitoring of hydrological parameters

5.5.1 Rationale

The field monitoring program was undertaken to identify and quantify the magnitude (i.e. depth of tidal water) and periodicity of tidal flooding in the intertidal zone and the depth of the local groundwater table beneath the low marsh surface through time. These variables were monitored *via* the local tide gauge and a piezometer respectively in order to gain insights into spatial and temporal variations in effective stress that result from tidal- and ground-water changes. The data obtained can then be used to:

1. assist in the explanation of potential structural values that may be observed in the soil, such as the preconsolidation (yield) stress;
2. sensibly modify conventional oedometer testing procedures to more accurately reflect intertidal field conditions;
3. design the dynamic loading testing scenarios in the back-pressured shear box to determine the effects of cyclic loading on the mechanical autocompaction behaviour of cohesive intertidal sediments. The repeated loading, unloading and reloading of these sediments may result in fundamentally different stress-strain behaviour.

It was necessary to calculate effective stresses in the ground for each of these applications of the hydrological data. Before this can be done, variations in the hydrological parameters are presented and described.

5.5.2 Tidal gauge data

The altitudes of the tidal waters obtained from the tide gauge at Greatham Creek, organised as monthly time-series, are presented in Figure 5.23. Altitudes of the mudflat and low marsh surfaces are displayed on the graphs. The full tidal range was not recorded due to logistical problems (Section 4.2.3).

The tidal regime at Greatham Creek is semidiurnal (i.e. tidal cycles of approximately twelve hours, resulting in two high water and two low waters each day). The tides are generally mixed, displaying conspicuous diurnal inequality between successive high low tides; this effect is particularly noticeable during the rising segments of the spring cycles during April 2004 and November 2004 and between 15th and 25th July 2005. Occasionally, the tides show semidiurnal equality, as occurs during the rising segment of the second spring cycle of February 2004, for example. The tides also show a neap-spring tidal cyclicity, though this more pronounced during some months than others.

In terms of flooding frequency, the mudflat surface (1.06 m OD) is flooded on the majority of tides. Only during neap phases in January, February, March and September 2004 does the tide not reach sufficient altitude to flood this surface. The low marsh surface (2.26 m OD) is generally flooded on high tides during spring tide phases, although on occasion this may not occur due to a 'depressed' spring high tide phase; this phenomenon occurs during May and July 2004.

Figure 5.24 presents monthly boxplots of the altitudes of tidal waters to ascertain the extent of monthly variations in tidal flooding characteristics. There is relatively little monthly variation in the form of the boxplots. The medians, interquartile ranges and 90th percentiles all plot at similar altitudes. Local mean sea level remains largely constant throughout the monitoring period. This is to be expected, since the tidal cycle is controlled by the gravitational effects of the moon and the sun in relation to the earth – factors that operate with largely constant periodicity. Minor fluctuations in the form of the boxplots and the upper 10th percentiles are caused by local/regional meteorological factors, such as increased or decreased precipitation that leads to increased or decreased surface runoff and fluvial input into the estuarine system, atmospheric pressure variations that depress or raise tidal water levels and the effects of wind driven storm surges. An extreme storm event occurred during January 2005, where tidal waters exceeded HAT (3.25 m OD), flooding altitudes of 3.70 m OD (Figure 5.23). This storm event was also recorded at Teesside Weather Station (http://homepage.ntlworld.com/d.e.martin); indeed, the highest recorded windspeed occurred during January 2005.

5.5.3 Groundwater data

Variations in the altitude of the groundwater level and in relation to the low marsh surface altitude (2.26 m OD) are displayed in Figure 5.23, organised as monthly time-series. The altitude of the low marsh surface displayed on each of the graphs also gives an indication of the depth to which groundwater drops beneath the low marsh surface during the monitoring period. No data were recorded during January 2005.

By comparison with the tidal time-series data, it is evident that groundwater variations are controlled to an extent by fluctuations in the altitude of tidal water. This relationship is most pronounced when the altitude of the tidal water rises and equals the altitude of groundwater during high spring tides. At times, the relationship is close to parity in terms of the timing of rises and falls in water levels. However, it is noticeable that the maximum amplitude of flooding recorded by the tide gauge does not equal that recorded by the piezometer when both water levels are above the surface of the low marsh. This may be due to a distortion of the principal tidal wave (as recorded by the tide gauge) by the upper intertidal bathymetry and geomorphology (e.g. tidal creeks, stepped saltmarsh topography etc.). Van der Molen (1997) noted that such factors lead to a lowering of the flood maximum with increasing distance from an estuary inlet and direct oceanic control. This may provide an explanation for the observed variation in amplitude of the flood peak.

As tidal waters recede during spring tide phases, so too do groundwater levels. However, they only fall to within a certain height above the low marsh surface due to the higher storage capacity of the saltmarsh sediment aquifer and the low permeability of these sediments, which together inhibit total drainage. During spring tide phases, the level of the groundwater during low tides remains largely constant throughout the sampling period, at approximately 2 m OD (*c.* 0.25 m beneath the low marsh surface). However, during neap tides when tidal waters do not recharge the upper section of the stratigraphic column semidiurnally, groundwater levels generally operate independently of the tidal forcing. During neap tides, despite smaller fluctuations in groundwater levels in response to tidal forcing, the groundwater level begins to reflect the prevailing meteorological conditions. During the wetter winter months (December – February 2004, for example), groundwater levels remain relatively high during neap phases, presumably due to a continuous supply of meteoric waters. In contrast, following lowering of tidal waters at the end of a spring cycle during warmer and drier summer months, the low marsh groundwater level continues to fall until it is recharged by rising spring tidal waters. This trend is evident between April and July 2004.

Seasonal (monthly) trends in groundwater variation are further highlighted in Figure 5.25. During the winter months, groundwater depths are generally within 0.40 m of the low marsh surface. As the weather became warmer and precipitation decreased in quantity and frequency, groundwater levels fell during drier months to a maximum depth 0.92 m beneath the ground surface on May 30th 2004. During these periods, the groundwater fluctuated over a larger range of depths than during wetter conditions.

5.6 THE INTERTIDAL EFFECTIVE STRESS ENVIRONMENT

Pore pressure variations in hydrostatic and hydrodynamic conditions were outlined in Sections 3.9.6 and 3.9.7. However, Ursino *et al.* (2004) added further complexity to analyses of pressure head distributions within saltmarsh environments. They presented a finite element model of saturated-unsaturated subsurface flow in a schematic saltmarsh. An important outcome of their numerical experiments was that pressure head distributions within saltmarsh environments are complex and do not vary solely in response to (tidally-driven) groundwater variations. The model shows that saltmarsh subsurface flow depends on the distance from the nearest tidal creek and that the subsurface water movement near tidal creeks is both vertical and horizontal, while further from creeks it is primarily vertical. This significantly complicates analysis of seepage pressures calculated from hydraulic gradients (Punmia, 1994). Furthermore, the study shows that if the hydraulic conductivity of the soil is relatively low (e.g. 10^{-6} m/s), a permanently unsaturated zone is present below the surface even after the tide has flooded the marsh. In such situations, hydraulic conductivity and gradient calculations differ substantially from saturated conditions, particularly when modified in a complex manner by evapotranspiration, which increases the extent and persistence of the aerated layer and induces a strong positive feedback, increasing soil oxygen availability and thereby creating conditions suitable for further ecological succession. Essentially, subsurface flow in the marsh leads to complex water table dynamics, even when the tidal forcing has a simple sinusoidal form.

Since the subsurface total pressure head distribution is extremely complex, conventional civil engineering analyses of pressure distributions are highly unlikely to be valid in intertidal (particularly saltmarsh) environments. Indeed, in order to develop an accurate understanding of pressure variation in saltmarsh stratigraphies, a synthesis of geotechnical (requiring installation of further instrumentation, such as piezometers monitoring pore pressures at depth and tensiometers for measuring capillary suction pressures) and biogeomorphological data (vascular plant zonation and dynamics, distance from tidal creeks, hydraulic conductivity, topography etc.) and analytical modelling techniques would be required.

In light of this complexity, it is necessary to simplify the observed effective stress variations in order to gain initial insights into the effects of dynamic loading on cohesive intertidal sediments. Therefore, first-order approximations of the magnitudes and periodicities of pore water pressures (and hence effective stresses) for use in this investigation are calculated from the tidal gauge and groundwater data using hydrostatic principles (Sections 3.9.5 and 3.9.6).

From the tidal and groundwater data, effective stresses acting on the surface materials in their virgin state can be calculated. Assuming hydraulic disconnectivity at the level of the sediment surface–tidal water interface, both mudflat and low marsh sediments are likely to undergo dynamic, cyclic loading by tidal waters. From Figure 5.24, it can be seen that the highest tidal depths were experienced during January 2005. The tidal loads (effective stresses) associated with the tidal flooding during this month are displayed in Figure 5.26 and Figure 5.27. The lower elevation of the mudflat sampling station in the intertidal zone means that it is flooded more frequently by greater depths of water. Surface mudflat materials therefore experienced greater effective stresses ranging from 3 to 25 kPa during January 2005. In contrast, the low marsh sediment surface is loaded less frequently (during spring tides only) and for shorter periods of time by shallower depths of water. Therefore, the low marsh surface sediments experience lower amplitude, more infrequent effective stress cycles within a range of 0 to 13 kPa during January 2005.

Before variations in effective stress induced by falls in groundwater could be ascertained, it was necessary to determine whether the sediments that underlie the low marsh (and indeed mudflat) surface are likely to remain saturated by capillarity during falls in groundwater. In order to do so, 1 metre sediment cores were obtained using a Russian chamber corer. The cores were sampled at 5 cm depth intervals and analysed to determine their particle size distributions from which values of D_{10} were obtained using GRADISTAT (Blott and Pye, 2001). The calculated air entry values at depths are presented in Figure 5.28. Also plotted are the capillary suction pressures that would be experienced at specific depths if the groundwater were to drop to 1 metre beneath the ground surface (hydrostatic variation). Since this capillary pressure never exceeds the air entry value of both the low marsh and the mudflat sediments, it is reasonable to assume that the sediments are prone to effective stress variations as a result of capillarity.

Figure 5.25 illustrates that the greatest depth of groundwater occurred during May 2004. Figure 5.29 illustrates the influence of these groundwater falls on effective stresses at the low marsh sediment surface. During the first spring tide phase of the month (3^{rd} – 10^{th}

May), effective stress acting on the surface sediment is cyclic/semidiurnal with an amplitude of approximately 3 kPa. As the groundwater level falls as neap tides operate (24^{th} May – 31^{st} May, for example), effective stress at the surface increases steadily, with minor fluctuations (amplitude of *c.* 0.5 – 1 kPa) superimposed on a generally positive linear trend. Surface effective stress reaches its peak (*c.* 9 kPa) when the groundwater level falls to its minimum (*c.* 0.9 m beneath the surface). Even during neap tide phases when groundwater is generally less dynamic (smaller amplitudes in effective stress variation), effective stress is never constant. Indeed, assuming the groundwater level beneath the mudflat surface operates with a similar periodicity (if not similar amplitude), a combination of tidal loading, capillary suction and subaerial exposure ensure that intertidal materials are rarely, if ever, subjected to a constant value of effective stress.

5.7 IMPLICATIONS FOR AUTOCOMPACTION BEHAVIOUR

The data presented in this chapter have characterised the contemporary geotechnical environment in terms of (a) the types of materials that form in the intertidal environment; and (b) the magnitude, frequency and periodicity of effective stresses applied to these materials. The implications of these data can be examined with reference to the first four of the research hypotheses defined in Section 3.10:

1. Mineralogenic intertidal sediments show no variability in structure and/or compression behaviour.

All low marsh samples were obtained from constant altitude (2.26 m OD) and displayed largely homogenous sedimentological properties. The mudflat samples, again collected from constant altitude (1.06 m OD), also displayed near-uniform lithology. However, considerable structural variability was observed in each sediment type. Voids ratios for the low marsh ranged from 3.69 to 5.56, and from 1.71 to 2.16 for the mudflat. Hypothesis 1 can therefore be partly rejected. The implications of this finding are that the single values of initial voids ratio used in Terzaghi's compression law are unlikely to adequately describe the range of structures observed at the depositional surface. Hence, the basic model will require modification to fully describe the range of initial voids ratios. Geotechnical testing is required to determine whether the observed structural variations translate into differences in one-dimensional compression behaviour, allowing a full rejection of the hypothesis.

2. Existing geotechnical laboratory methods sufficiently represent intertidal field conditions.

Two main findings are relevant to this hypothesis. Firstly all low marsh samples and all but one of the mudflat samples were not fully saturated (mean saturation values of 92.57 % and 96.27 % respectively). Secondly, the intertidal effective stress environment has been found to be highly dynamic due to the combined operation of tidal water loading, capillary suction and subaerial exposure.

Both of these findings contrast greatly with established geotechnical methods (as outlined in BSI, 1990) in which (a) samples undergo a 24 hour saturation phase prior to experimental loading and (b) load increments applied to the specimens remain constant for a 24 hour period.

Hypothesis 2 can therefore be emphatically rejected. The behavioural response of the sediments to field conditions is unknown. Therefore, in order to accurately simulate the effective stress conditions experienced by near-surface samples in the field, modifications must be made to conventional geotechnical testing methods (as outlined in BSI, 1990).

3. Near-surface mineralogenic intertidal sediments are normally consolidated.

Surface samples have been subjected to effective stresses in excess of the (by definition non-existent) sediment overburden. On the basis of groundwater data, the low marsh samples can be expected to have a preconsolidation stress of *c.* 9 kPa. Assuming hydraulic disconnectivity and sufficient time to allow dissipation of excess pore pressures, tidal loads can potentially create preconsolidation pressures of *c.* 13 kPa in the low marsh surface material and *c.* 25 kPa in the mudflat surface material. Furthermore, these preconsolidation stresses are minimum values that may have been substantially increased by, for example, desiccation and desaturation. Both sediment types are therefore likely to be overconsolidated, indicating that Terzaghi's compression law involving a backward projection of the normal compression line is a deficient description of mineralogenic intertidal compression behaviour. However, hypothesis 3 cannot be unequivocally rejected without confirmation from geotechnical testing.

4. Terzaghi's compression law and consolidation theory are applicable to mineralogenic intertidal sediments.

The material and environmental prerequisites for valid use of Terzaghi's compression law and consolidation equation were outlined in Section 3.6. The results presented in this chapter suggest, albeit tentatively, that these models are inapplicable to the two materials types selected in this chapter. Firstly, both the low marsh and mudflat materials have organic contents (mean values of 24.68 % and 16.83 % respectively) above the maximum threshold (5 %) for which Terzaghi's compression law and consolidation theory should strictly be applied. In the low marsh material in particular, the *in situ* organic growth may result in the soil particles being compressible. Hence, they may prone to considerable creep deformations. Both low marsh and mudflat materials are unsaturated to some degree (mean saturation values of 92.57 % and 96.27 % respectively). Again, this violates the assumptions of Terzaghi's laws and theories. Furthermore, the effective stresses experienced by near-surface sediments may result in primary consolidation that is not caused by overburden sedimentation and may result in overconsolidation. Without directly examining the results of geotechnical tests, it is not possible to reject hypothesis 4. Nevertheless, the results provided in this chapter provide strong evidence that neither Terzaghi's compression law nor consolidation equation are applicable to the dynamically loaded, semi-organic sediments of the intertidal zone at Greatham Creek.

5.8 Summary

Even before the results of the one-dimensional compression testing program are presented, analysis of the contemporary geotechnical intertidal environment and the physical properties of the materials that form there provides valuable and interesting insights into the nature of autocompaction processes in mineralogenic sediments.

Some of the fundamental assumptions of existing compression and consolidation theories employed for using in autocompaction analysis in intertidal sediments have, at the very least, been challenged if not unquestionably violated. In particular, high organic contents and unsaturated sediments may result in fundamentally different compression behaviour than is described using existing algorithms. Examination of the contemporary effective stress environment strongly suggests that materials are likely to be overconsolidated at least to some degree, even at the depositional surface. In conjunction with the confirmed structural variability of intertidal sediments, the validity of describing autocompaction behaviour in terms of a single compression index is further called into question. It is now necessary to determine how these materials of variable structure respond to identical loading scenarios to ascertain if there are any 'memory effects' of initial structural variation.

Detailed examination of the contemporary effective stress environment also indicates an undisputable degree of semidiurnal dynamism. This raises doubts about the usefulness of the oedometer test in reproducing *in situ* stress conditions and reinforces the need to undertake a detailed laboratory investigation into the compression behaviour of cohesive intertidal sediments.

Chapter 6 develops the material testing program developed on the basis of the idiosyncrasies of the dynamic intertidal environment and presents the results of these geotechnical laboratory tests.

CHAPTER 6: ONE-DIMENSIONAL COMPRESSION TESTING OF MINERALOGENIC INTERTIDAL SEDIMENTS

6.1. MATERIAL TESTING PROGRAM

6.1.1 Conventional oedometer testing

The material testing program for the low marsh and mudflat materials began with a series of oedometer tests that employed conventional, established methods (BSI, 1990). Table 6.1 displays the loading stages that are recommended in BS 1377, Part 5 (BSI, 1990) and Head (1988), where each new stage of compression for each loading increment involves an approximate doubling of the previous loading pressure (i.e. a load increment ratio, $\Delta\sigma/\sigma$, approximately equal to unity).

Table 6.1 Loading stages common to all incremental loading tests (following BS1377, Part 5 – BSI, 1990). Applied pressures on material specimens are calculated on the basis of a lever ratio of 11.04:1, a loading cap mass of 0.65 kg and a sample cross-sectional area of 441 mm^2.

Mass on hanger (kg)	Applied stress σ, (kPa)	$\Delta\sigma$	Load increment ratio, $\Delta\sigma/\sigma$
1	26	-	-
2	50	24	0.92
4	99	49	0.98
8	197	98	0.99
16	393	196	0.99
32	785	392	1.00
64	1570	785	1.00

Prior to the application of 26 kPa to individual specimens, load increment ratios did not always equal unity. A range of loading scenarios was employed in order to constrain the preconsolidation stress more accurately. An interesting outcome of this was that insights into the effect of variations in load increment ratios on stress-strain behaviour could be gained. The two low stress loading scenarios employed are displayed in Tables 6.2 and 6.3 to illustrate the differences in load increments, $\Delta\sigma$, and the load increment ratio, $\Delta\sigma/\sigma$.

Additional saturated tests involving extended (48 or 68 hours per load increment) or truncated (where loading durations equalled the time taken for excess pore pressures to dissipate, estimated from time-settlement plots) loading stages were undertaken to further investigate the effects of prolonging or excluding creep processes.

Table 6.2 Low stress loading scenario 1. Applied pressures on material specimens are calculated on the basis of a lever ratio of 11.04:1, a loading cap mass of 0.65 kg and a sample cross-sectional area of 441 mm^2. The 26 kPa loading stage is included to illustrate load increment ratio variation in relation to loading scenario 2 (Table 6.3).

Mass on hanger (kg)	Applied stress σ, (kPa)	$\Delta\sigma$	Load increment ratio, $\Delta\sigma/\sigma$
0.15	5	-	-
0.5	14	9	1.8
1	26	12	0.86

Table 6.3 Low stress loading scenario 2. Applied pressures on material specimens are calculated on the basis of a lever ratio of 11.04:1, a loading cap mass of 0.65 kg and a sample cross-sectional area of 441 mm^2. The 26 kPa loading stage is included to illustrate load increment ratio variation in relation to loading scenario 1 (Table 6.2).

Mass on hanger (kg)	Applied stress σ, (kPa)	$\Delta\sigma$	Load increment ratio, $\Delta\sigma/\sigma$
0.15	5	-	-
0.35	10	5	1
0.5	14	4	0.4
0.75	20	6	0.43
1	26	6	0.3

For each oedometer test, samples were loaded until the maximum compression limit of the oedometer apparatus was reached – i.e. when further downward movement of the load hanger was prevented by the fully lowered screw jack unit. This often occurred prior to the application of the selected maximum pressure application (1570 kPa) due to variations in material compressibility, particularly with the low marsh material. Care was taken to ensure that each load increment was allowed to act upon the sample for the complete loading duration (generally 24 hours) so that a loading stage was not ended prematurely. Behaviour during removal of loading (unloading) was determined by dividing the existing loading mass by three and rounding the value obtained up or down to the nearest 0.5 kg

(BSI, 1990). Reload behaviour was determined by re-applying the unloading sequence (BSI, 1990).

6.1.2 Modified one-dimensional compression testing

The field data presented Chapter 5 raised important questions regarding the ability of conventional oedometer tests to reproduce *in situ* stresses and saturation conditions and hence compression behaviour. Conventional oedometer methods, as outlined in BS 1377 (BSI, 1990), begin with a saturation phase. However, since materials in the field are generally unsaturated to at least some degree, the observed laboratory behaviour may differ from that which occurs *in situ*. Furthermore, load increments are generally applied for 24 hours. Following dissipation of excess pore pressures, the applied effective stress remains constant – a situation which does not reflect the highly dynamic effective stress conditions observed in the field. In addition to the conventional oedometer tests outlined above, further laboratory tests were undertaken to better reproduce the observed field conditions in order to gain insights into the effects of the geotechnical idiosyncrasies of the intertidal environment on autocompaction behaviour.

In order to determine the effect of increasing the effective stress on a sample at natural moisture content, low marsh and mudflat samples were loaded in an oedometer without the initial saturation phase. Samples were loaded to pressures up to and including 10 kPa for 24 hours per load increment; at this point, the samples were flooded and were allowed to saturate for 24 hours, during which any vertical displacement was recorded. Following saturation, loading continued in the conventional manner and included unloading and reloading phases. 10 kPa was chosen as the 'saturation pressure' on the basis of the maximum stress calculated at the low marsh surface as a result of groundwater fall (*c.* 9 kPa during May 2004, Figure 5.29).

The effects of dynamic effective stresses on one-dimensional compression behaviour were investigated using the back-pressured shear box apparatus. Before any dynamic loading of intertidal materials was undertaken, it was important to determine that the back-pressured shear box replicated the non-dynamic incremental loading behaviour observed in the oedometer for comparative purposes. Therefore, both low marsh and mudflat samples were compressed in the back-pressured shear box using the loading stage displayed in Tables 6.1 and 6.2. These tests are analysed with the data obtained from oedometer tests.

The first type of dynamic load test involved variations in effective stress acting on the surface materials. The actual amplitude, period and time between loading events was determined from hydrological field data. However, given the short-term nature of the laboratory tests (7 days including saturation) and the large (annual) datasets from which semidiurnal dynamic loading scenarios could be extracted, choosing a 'representative' testing scenario is a near-impossible task. Instead, it is important that the dynamic loading testing scenarios employed represent the correct range of magnitude of effective stresses operating at the low marsh and mudflat sampling altitudes. Furthermore, these magnitudes of effective stresses must be applied at the correct frequency and periodicity as those observed in the field.

Low stresses (< 3 kPa) were necessarily avoided due to accuracy constraints of the hydraulic pressure controllers. Pilot tests indicated that pressures can be controlled with an accuracy of ± 1 kPa. If negative effective stresses are created, tests stop prematurely. Therefore, to avoid such a situation, the minimum effective stress programmed into the test software for phases representing zero effective stress is 3 kPa. A critical outcome of this is that variations at very low effective stresses (between 0 and 3 kPa) could not be used in the dynamic loading tests.

In order to avoid inconveniently low and limiting effective stresses, the dynamic loading tests are based on the period of monitoring that recorded the highest tidal water level (January 2005 – Figure 5.24). Hence, the upper effective stress boundary conditions observed in the field were tested; it was decided that higher dynamic loads were necessary if variations in compression behaviour were to be observed at all. The period 9^{th} – 14^{th} January has been used as the basis of the testing scenario (Figures 5.26 and 5.27). In the absence of groundwater-driven effective stress data for the mudflat surface material, only tidal loading data are used (approximate effective stress amplitude range of 6 – 25 kPa). Furthermore, the range of effective stress amplitudes observed in the low marsh groundwater data (*c.* 1 – 10 kPa) and the semidiurnal frequency of operation are similar to those caused by tidal water loading. Although groundwater and tidal loading data suggest that effective stress is rarely constant at the depositional surface, periods of loading are purposefully interspersed with periods of constant effective stress. This is to determine the relative effects of constant and dynamic effective stress phases on compression.

The second type of test undertaken in the back-pressured shear box involved a combination of overburden and dynamic surcharge loading. The purpose of this test is to determine the effect of dynamic loading on rates and magnitudes of volumetric change at increasing depths of overburden. This test takes the general form of the oedometer loading increments displayed in Tables 6.1 and 6.2. However, superimposed on the incremental loading is a dynamic semidiurnal surcharge load. For simplicity, the amplitude and period of the semidiurnal load (applied in addition to the simulated overburden load) during each load increment are equal to those observed for the maximum tidal load recorded by the local tide gauge. For the mudflat, the amplitude of this dynamic load is 25 kPa and the period is 800 minutes. For the low marsh, the dynamic load amplitude is 13 kPa and the period is 554 minutes. For each overburden load increment (lasting 24 hours), two dynamic loading phases are spaced between three phases of constant effective stress.

The full material testing program is summarised in Table 6.4

6.1.3 Sample identification codes and physical properties

Sample identification codes for incremental loading tests reflect:

1. the material ('LM' for low marsh, 'MF' for mudflat);
2. the test number on each material;
3. whether the material was saturated ('S') or was at its natural moisture content ('N') at the beginning of the test;
4. the equipment used to undertake a particular test ('OED' for the oedometer, 'BPS' for the back-pressured shear box);
5. the duration of each load increment (24, 48 or 68 hours; 'NC' indicates creep was prevented).

Table 6.5 displays key physical properties of the samples used in the incremental loading tests. The initial voids ratio refers to the *in situ* field voids ratio – the voids ratio at the depositional surface. This must be distinguished from the initial voids ratio discussed by Sills (1998), which refers to the voids ratio at which effective stresses first develop within a soil sedimenting from a suspension. Although not mutually exclusive, the environmental conditions discussed in Chapter 5 suggest a degree of overconsolidation is likely and hence that these two reference voids ratios are likely to differ.

Table 6.4 Overview of the material testing program employed in this investigation.

Type of test	Number of tests undertaken:	
	Low marsh material	Mudflat material
Oedometer tests. Variable low stress loading stages. 24 hour load increment duration. Saturated. Load-unload-reload stages.	3	3
Oedometer tests with extended load increment durations: 48 hour per increment for mudflat; 68 hour per increment for low marsh. Saturated. Load-unload-reload stages.	1	1
Oedometer tests with truncated load increment durations. Load increment ended following dissipation of excess pore water pressures (creep prevented). Saturated. Loading stage only.	1	1
Unsaturated oedometer tests. 24 hour load increment duration. Load-unload-reload stages.	2	1
Back-pressured shear box test: incremental (oedometer) loading. Saturated. 24 hour load increment duration. Load-unload-reload stages.	1	1
Back-pressured shear box test: dynamic loading of surface materials based on tidal loading data. Saturated.	2	2
Back-pressured shear box test: dynamic loading superimposed on incremental loading scenario. Saturated.	1	1
Total	11	10

Table 6.6 displays some selected physical properties of samples used in one-dimensional cyclic compression test program. Samples identification codes for these tests reflect:

1. the material;
2. the test number;
3. the equipment used to undertake the test;
4. the type of test ('CYC' indicates the cyclic loading of near-surface materials; 'IL+CYC' indicates the combined incremental and cyclic loading tests).

All samples analysed in the back-pressured shear box cyclic loading tests were saturated (Section 4.5.3).

Table 6.5 Physical properties of samples used in one-dimensional, incremental loading compression test program.

Sample I.D.	Loss on ignition (%)	Sand (%)	Silt (%)	Clay (%)	Natural moisture content, w (%)	Initial voids ratio, e_i	Unload-reload phases
			Low marsh samples				
LM-1-S-OED-24[1]	23.82	14.12	71.74	14.14	147.57	4.30	✓
LM-2-S-OED-24[2]	23.63	12.62	71.65	15.73	155.61	3.98	✓
LM-3-S-OED-24[2*]	23.75	7.71	74.78	17.51	151.37	4.02	✓
LM-4-S-OED-68[1*]	24.66	12.91	71.71	15.38	137.05	3.69	✓
LM-5-S-OED-NC[1]	24.12	11.46	73.07	15.46	170.06	4.51	×
LM-6-S-BPS-24[1]	27.05	15.55	68.58	15.87	182.91	4.88	✓
LM-7-N-OED-24[2]	25.57	9.95	72.78	17.28	162.39	4.26	✓
LM-8-N-OED-24[2]	26.14	9.94	73.41	16.64	172.42	4.60	✓
			Mudflat samples				
MF-1-S-OED-24[2]	17.32	15.09	71.94	12.98	83.51	2.34	✓
MF-2-S-OED-24[2]	15.12	14.64	71.96	13.40	79.57	2.22	✓
MF-3-S-OED-24[2]	19.37	14.78	72.83	12.38	92.88	2.68	✓
MF-4-S-OED-48[1]	14.11	15.27	74.62	10.11	61.02	1.71	✓
MF-5-S-OED-NC[1]	16.22	13.22	71.99	14.79	80.48	2.32	×
MF-6-S-BPS-24[1]	16.56	16.90	72.49	10.61	69.47	1.86	✓
MF-7-N-OED-24[2]	16.78	15.91	72.82	11.27	88.83	2.41	✓

[1] – indicates sample is loaded using low stress loading scenario 1 (Table 6.2).
[2] – indicates samples is loaded using low stress loading scenario 2 (Table 6.3).
[1*] - indicates sample is loaded using low stress loading scenario 1 (Table 6.2) but without a 5 kPa loading stage.
[2*] - indicates sample is loaded using low stress loading scenario 2 (Table 6.3) but without a 20 kPa loading stage.

Table 6.6 Physical properties of samples used in one-dimensional dynamic loading compression test program.

Sample I.D.	Loss on ignition (%)	Sand (%)	Silt (%)	Clay (%)	Natural moisture content, w (%)	Initial voids ratio, e_i
			Low marsh samples			
LM-9-BPS-CYC	23.89	9.31	73.09	17.60	137.09	3.96
LM-10-BPS-CYC	24.05	15.75	69.81	14.44	160.62	4.37
LM-11-BPS-IL+CYC	26.04	13.92	71.18	14.90	172.12	4.48
			Mudflat samples			
MF-8-BPS-CYC	17.48	12.47	73.37	14.16	70.26	1.89
MF-9-BPS-CYC	18.32	18.70	69.66	11.64	69.24	1.75
MF-10-BPS-IL+CYC	17.73	15.16	71.83	13.01	89.27	2.41

6.2 ONE-DIMENSIONAL INCREMENTAL LOADING

6.2.1 Material compressibility and $e\log_{10}\sigma'$ analysis

$e\log_{10}\sigma'$ plots for low marsh samples are displayed individually in Figures 6.1 to 6.8 and collectively in Figure 6.9; for mudflat samples these are displayed individually in Figures 6.10 to 6.16 and collectively in Figure 6.17. The material properties obtained from these plots are presented in Table 6.7 and are discussed below. Preconsolidation stresses, σ'_c, the maximum vertical effective stress to which the soil has previously been subjected, were estimated using the widely-used graphical construction suggested by Casagrande (1936). This is based on an analysis of the intersection between the recompression line and the normal compression line (Figure 6.18). The preconsolidation stress also marks a significant change in the compressibility of the material; it is the vertical effective stress at which the compression stress path begins to move from the lower gradient recompression line to the steeper virgin compression line. It represents a critical point at which structural resistance to effective overburden stress fully breaks down (Burland, 1990).

6.2.2 Compression behaviour of low marsh material

Figures 6.1 to 6.9 illustrate that all low marsh samples have the same general form, displaying a consistent pattern of compression behaviour in $e\log_{10}\sigma'$ space. It is also immediately apparent that the material is overconsolidated - the preconsolidation stress is higher than the existing overburden (0 kPa, since all samples were collected from the depositional surface). Preconsolidation stresses for low marsh materials ranged from 23 kPa to 27 kPa (Table 6.7). Furthermore, as described by the overconsolidation ratio (mean value of 23.88), the material is heavily overconsolidated in relation to the (non-existent) overburden. The range of preconsolidation stresses reflects spatial variations in effective stress within the intertidal environment and is therefore unsurprising.

Table 6.7 Material properties of intertidal samples obtained from $e\log_{10}\sigma'$ plots.

Sample I.D.	C_r	C_c	C_s	σ'_c (kPa)	OCR	C_r^*	C_c^*	C_s^*
				Low marsh samples				
LM-1-S-OED-24	0.19	1.44	0.28	23	23	0.04	0.35	0.06
LM-2-S-OED-24	0.19	1.71	0.23	26	26	0.03	0.43	0.05
LM-3-S-OED-24	0.16	1.54	0.25	24	24	0.03	0.38	0.06
LM-4-S-OED-68	0.08	1.41	0.21	24	24	0.02	0.38	0.06
LM-5-S-OED-NC	0.25	2.12	-	27	27	0.03	0.44	-
LM-6-S-BPS-24	0.52	2.27	0.25	23	23	0.05	0.42	0.06
LM-7-N-OED-24	0.17	1.92	0.26	24	24	0.03	0.44	0.06
LM-8-N-OED-24	0.26	1.88	0.26	20	20	0.05	0.41	0.06
Mean	0.23	1.89	0.25	23.88	23.88	0.04	0.41	0.06
Standard deviation	0.12	0.43	0.02	2.10	2.10	0.01	0.03	0.01
				Mudflat samples				
MF-1-S-OED-24	0.19	0.57	0.08	8	8	0.08	0.24	0.03
MF-2-S-OED-24	0.15	0.62	0.08	8	8	0.06	0.27	0.03
MF-3-S-OED-24	0.14	0.73	0.10	10	10	0.05	0.28	0.03
MF-4-S-OED-48	0.06	0.50	0.05	12	12	0.04	0.29	0.04
MF-5-S-OED-NC	0.04	0.63	-	14	14	0.02	0.27	-
MF-6-S-BPS-24	0.02	0.52	0.05	10	10	0.02	0.30	0.05
MF-7-N-OED-24	0.25	0.64	0.09	14	14	0.09	0.26	0.04
Mean	0.12	0.60	0.08	10.86	10.86	0.05	0.27	0.03
Standard deviation	0.08	0.08	0.02	2.54	2.54	0.03	0.02	0.01

KEY

C_r = Recompression index
C_c = Compression index
C_s = Unload-reload swell index
σ'_c = Preconsolidation stress

OCR = Overconsolidation ratio = $\dfrac{\sigma'_c}{\sigma'_o}$

C_r^* = Normalised recompression index
C_c^* = Normalised recompression index
C_s^* = Normalised unload-reload swell index

Prior to the preconsolidation stress, volumetric change (reduction in voids ratio) is small per unit stress applied. The slope of this section of the loading curve is described by the recompression index, C_r. Values of C_r are presented in Table 6.7 (mean = 0.23). At vertical effective stresses above the preconsolidation stress, compression behaviour follows the virgin/normal compression line, the gradient of which is described by the compression index, C_c. For the low marsh material, the mean value of the compression index is 1.89 – a figure that is significantly greater than C_r, illustrating the dramatic variation in compressibility at stresses above and below the preconsolidation stress. The normal compression line is essentially linear in $e\log_{10}\sigma'$ space. However, samples which were subjected to stresses greater than 393 kPa (LM-1-S-OED-24, LM-3-S-OED-24, LM-4-S-OED-24, LM-6-S-OED-24 – Figures 6.1, 6.3, 6.4 and 6.6 respectively) illustrate a slight decrease in the compression index at higher stresses, giving the normal compression line a log-curvilinear form.

Despite the general trends in compression behaviour described above, there is obvious variability in these trends. Figure 6.9 and Table 6.7 display differences in the gradient of C_r. This is best illustrated by considering the differences in C_r between LM-6-S-BPS-24 (C_r = 0.52) and LM-4-S-OED-68 (C_r = 0.08). These samples have the minimum and maximum values of C_r respectively and also represent the maximum and minimum values of e_i (4.88 and 3.69 respectively). Indeed, there is a strong, highly significant positive correlation between e_i and C_r of the low marsh samples (r = 0.883, p = 0.004) (Figure 6.19). Variability also exists for values of C_c, ranging from a minimum of 1.41 (LM-4-S-OED-68) to a maximum of 2.27 (LM-6-S-BPS-24). Again, there is a significant positive correlation between e_i and C_c (r = 0.823, p = 0.012) (Figure 6.20).

The unload-reload swell line appears strikingly similar in all samples tested for unload-reload behaviour (Figure 6.9). This is confirmed in Table 6.7, where C_s ranges from a minimum of 0.21 (LM-4-S-OED-68) to 0.28 (LM-1-S-OED-24), but with a mean of 0.25 and a standard deviation of 0.02. Furthermore, C_s does not significantly correlate with initial voids ratio (r = 0.604, n.s.), implying a reduction in structural variability at high stresses.

It is interesting to note that, in $e\log_{10}\sigma'$ space at least, there appears to be little difference in the compression behaviour between samples saturated at the start of the test and those saturated at a vertical effective stress of 10 kPa (samples LM-7-N-OED-24 and LM-8-N-OED-24; Figures 6.7 and 6.8). Comparison of the compression and swell indices of both sets of samples (i.e. natural moisture content and saturated) reveals very little difference, with the values of these parameters of samples LM-7-N-OED-24 and LM-8-N-OED-24 falling within the range of observed values of the remaining (initially saturated) samples. It is also noticeable that there is no observable swell upon saturation on the $e\log_{10}\sigma'$ plots of samples LM-7-N-OED-24 and LM-8-N-OED-24 (Figures 6.7 and 6.8), possibly because the samples were already close to full saturation (LM-7-N-OED-24 was 94.51 % saturated; LM-8-N-OED24 was 93.01 % saturated). These observations suggest that the degree of *in situ* unsaturation is not sufficiently significant to cause any observable change in the compression behaviour of low marsh materials.

A more profound influence on the nature of the low marsh compression behaviour may be attributable to the (prevention of) operation of creep processes. Sample LM-5-S-OED-NC, for which creep was prevented from occurring, plots above the samples subjected to creep (Figure 6.9). This is most clear in the virgin compression phase. It is also noticeable that LM-5-S-OED-NC has a higher preconsolidation stress than samples subjected to creep (σ'_c = 27 kPa, Table 6.7). LM-5-S-OED-NC has a less steep recompression line than LM-8-N-OED-24, which has a similar initial voids ratio (Figure 6.9). Indeed, the segment of the recompression line of LM-5-S-OED-NC between 1 and 5 kPa has an extremely low gradient (nearly flat), perhaps implying that the majority of compression in this early stage is caused by creep processes. The slope of the normal compression line of LM-5-S-OED-NC also appears to be steeper (C_c = 2.12) following the later transition (i.e. higher preconsolidation stress) between the recompression and normal compression lines.

The effects of an extended load increment duration (68 hours each) on sample LM-4-S-OED-68 do not seem to have noticeably altered the compression behaviour of the material (Figures 6.4 and 6.9). Rather, the compression behaviour seems to be commensurate with the low initial voids ratio (3.69) of the sample; it is the least compressible of all low marsh samples tested (C_r = 0.08, C_c = 1.41).

6.2.3 Compression behaviour of mudflat material

Figures 6.10 to 6.17 show that all of the mudflat samples show consistent trends in $e\log_{10}\sigma'$ compression behaviour. All samples are overconsolidated and preconsolidation stresses range from 8 to 14 kPa (mean = 10.86 kPa, standard deviation = 2.54 kPa).

Recompression stress paths (mean C_r = 0.12; standard deviation = 0.08) are again significantly less steep than the normal compression stress paths (mean C_c = 0.60; standard deviation = 0.08), signifying a large difference in compressibility at stresses above and below the preconsolidation stress. Normal compression lines are strongly log-linear. As in the low marsh samples, variability exists in the compression behaviour of the mudflat samples (Figure 6.17, Table 6.7). The highest observed value of the recompression index, C_r, (0.25) is from sample MF-7-N-OED-24. The lowest value of C_r was calculated for sample MF-6-S-BPS-24. In contrast to the low marsh samples, only a weak and statistically insignificant correlation exists between e_i and C_r (r = 0.598, n.s.). This may be due to the presence and variability of vertical bioturbation hollows in the sediment (Table 5.1, Figure 5.10), providing a slight distortion of the 'true' (i.e. un-bioturbated) compression behaviour at low stresses. However, bioturbation of contemporary mudflat sediments is near-ubiquitous and its effect must be included in compression models. The highest value of C_c (0.73) was obtained from sample MF-3-S-OED-24. This sample also had the highest initial voids ratio (2.68). The lowest value of C_c is 0.50 (MF-4-S-OED-48), which corresponds with the value of the lowest initial voids ratio (1.71). There is a very strong and highly significant positive correlation between values of initial voids ratio, e_i, and compression index, C_c (r = 0.936, p = 0.002) (Figure 6.21). Unload-reload behaviour is similar in all samples (Figure 6.17). The mean value of observed unload-reload swell indices (C_s) is 0.08, with a standard deviation 0.02. In contrast to the low marsh material, initial voids ratio strongly and highly significantly correlates well with the unload-reload swell index (r = 0.982, p < 0.001) (Figure 6.22).

MF-7-N-OED-24 displays possible differences resulting from compression at natural moisture content up to and including 10 kPa (Figures 6.16 and 6.17). Prior to 5 kPa, it has a similar recompression index to sample MF-1-S-OED-24, both of which have similar initial voids ratios (2.34 and 2.41 respectively). However, MF-7-N-OED-24 has a higher preconsolidation stress than the remaining saturated samples. Despite this, it is difficult to determine whether this is a result of inherent structural variability or initial unsaturation. Post-yield, MF-7-N-OED-24 has a compression index of 0.64, which is towards the higher end of the range of observed values. Again, it is not easy to establish if this is due to compression at natural moisture content or the structural conditions of the sample given the strong positive correlation between initial voids ratio and compression index noted above.

Prevention of creep processes in sample MF-5-S-OED-NC has a clear impact on the degree of compression in the overconsolidated section of the $e\log_{10}\sigma'$ plot (Figure 6.14). At low stresses, the recompression curve is essentially flat, indicating no reduction in voids ratio despite the application of effective stress. Since all other samples noticeably compress at these stresses, and since creep was permitted in these samples, it is reasonable to assume that creep processes are the primary cause of volumetric reduction in the previously compressed (overconsolidated) section of $e\log_{10}\sigma'$ plots of mudflat sediments. Following the application of effective stresses greater than the preconsolidation stress, the compression behaviour of MF-5-S-OED-NC does not appear to be extraordinary ($C_c = 0.63$; compare with values for the mudflat material in Table 6.7).

Samples MF-4-S-OED-48 and MF-6-S-BPS-24 have very similar initial voids ratios (1.71 and 1.86 respectively) (Figure 6.17). Since these samples were also subjected to the same loading stages, direct comparison of their compression behaviour is possible, providing insights into the effects of extended operation of creep processes. It is evident from Figure 6.17 that there is no variation in compression behaviour other than a downward shift in the location (graphical intercept) of the normal compression line of sample MF-4-S-OED-48. The compression indices of these samples are very similar (0.5 for MF-4-S-OED-48 and 0.52 for MF-6-BPS-24). Recompression indices differ to a greater degree, with the value for MF-4-S-OED-24 (0.06) slightly greater than that of MF-6-BPS-24 (0.02); this phenomenon results from similar initial voids ratio and an increased rate of volumetric reduction occurring in MF-4-S-OED-24 due to additional creep settlement, leading to diverging recompression lines.

6.3 Compression behaviour of mineralogenic intertidal sediments

6.3.1 Material compressibility

Although both intertidal materials display similar compression behaviour, variability in this behaviour exists both between and within materials.

Mean values of all compression indices for the low marsh ($C_r = 0.23$, $C_c = 1.89$) are higher than those of the mudflat ($C_r = 0.12$, $C_c = 0.60$), indicating higher compressibility (more than three times greater during virgin compression) of the low marsh sediments. In addition, the mean unload-reload swell index of low marsh sediments ($C_s = 0.25$) is greater than that of mudflat materials ($C_s = 0.08$). This indicates a higher degree of elasticity in the low marsh sediments upon removal of overburden loads. The general higher compressibility and elasticity of the low marsh materials are undoubtedly related to the higher levels of organic material in the soil in comparison with mudflat sediments. The more 'open' initial structure of the low marsh sediments is more prone to volumetric reduction in response to loading than the initially denser mudflat materials. Furthermore, organic materials are themselves compressible (Head, 1988; Hobbs, 1986), adding an increased creep settlement element to consolidation processes.

Just as no intra-material relationship was observed between loss on ignition and initial voids ratio, correlations between organic content and the recompression index (r = 0.694, n.s.), the compression index (r = 0.649, n.s.) or the unload-reload swell index (r = 0.122, n.s.) indices were neither particularly strong nor statistically significant for the low marsh. Similar results were obtained from the mudflat sediments; loss on ignition was not significantly correlated with the recompression index (r = 0.325, n.s.), the compression index (r = 0.709, n.s.) or the unload-reload swell index (r = 0.696, n.s.). It can therefore be concluded that organic content affects the general compression trend of each material, but that sample-specific behaviour is governed by the initial structures of individual specimens, which are presumed to result from variations in depositional conditions.

It is evident that the observed variations in initial structure significantly affect the future one-dimensional compression behaviour of the sediment *via* structural 'memory' effects. The strong, positive and statistically significant correlations between the initial voids ratio, e_i and both the recompression (C_r) and compression (C_c) indices in low marsh and mudflat materials indicate that sediments with more 'open' initial structures are more susceptible to compression both pre- and post-yield. This correlation results in a convergence of the normal compression lines and reduction in the range of voids ratios at higher stresses, a phenomenon previously noted by Skempton (1970) and Burland (1990). The point at which the compression stress paths converge was termed the 'Omega point' by Tovey and Paul (2002). The application of an overburden pressure to materials with initial open fabrics results in a 'destructuration' process (Leroueil *et al.*, 1979, cited in Burland, 1990) during post-preconsolidation stress virgin compression. Destructuration removes the influence of initial structure. This results in similar compression behaviour at high stresses as indicated by converging compression lines and reduced standard deviations of values of the unload-reload swell index, C_s (0.02 in both low marsh and mudflat sediments) compared to C_r (0.12 in the low marsh material and 0.08 in the mudflat material) and C_c (0.43 and 0.08 in the low marsh and mudflat sediments respectively) (Table 6.7); higher standard deviations indicate a larger range of compression behaviour. Initially denser samples of lower initial voids ratio are more stable, closer to the intrinsic (i.e. independent of depositional structures) state of the material (Burland, 1990) and undergo less compression.

Correlations between the initial voids ratio and the unload-reload swell index provide further interesting insights into memory effects of initial structural conditions. In the low marsh material, the poor and statistically insignificant correlation between the two variables suggests a loss of memory of initial conditions following structural breakdown at the preconsolidation stress. In contrast, the strong and significant relationship between the initial voids ratio and the unload-reload swell index in the mudflat samples indicates that memory of the initial structure exists, despite considerable compression and destructuration at higher stresses. This possibly indicates the structural instability of virgin low marsh materials that is caused by increased organic content. In comparison, the denser initial structures in the mudflat sediments are more stable.

Clear variation in the compressibility of the low marsh and mudflat samples remains when voids ratios are normalised to the initial voids ratio (e/e_i) to minimise its of influence on graphical form and to allow easier comparison of compression trends. A series of nested curves is produced (Figures 6.23 and 6.24) where samples are 'stacked' in order of their relative compressibility; curves that plot lower on the *y*-axis indicate a greater degree of compressibility. Normalised recompression, compression and swell indices are displayed in Table 6.7.

Samples retain the same pattern of relative compressibility observed in the conventional $e\log_{10}\sigma'$ plots. In the low marsh specimens (Figure 6.23), sample LM-6-S-BPS-24, the most compressible of the low marsh samples observed previously, plots beneath each of the other compression curves. Furthermore, LM-4-S-OED-68 plots above the other samples, confirming it is the least compressible sample. LM-5-S-OED-NC again takes longer to enter its virgin compression phase and shows a more rapid rate of post-yield compression.

The same trends can be seen in Figure 6.24; samples tend to retain relative compressibility, with less compressible samples (i.e. those with lower initial voids ratios, such as MF-4-S-OED-48 and MF-6-S-BPS-24, and sample MF-5-S-OED-NC, which was not subjected to creep settlement) plotting higher than those with greater initial voids ratios.

6.3.2 Overconsolidation and preconsolidation stresses

In Chapter 5, field data were presented that suggested that both low marsh and mudflat materials would be overconsolidated. The preconsolidation stresses estimated from field data were 9 – 13 kPa for the low marsh sediment (as a result of falls in groundwater and tidal loading) and up to 25 kPa for the mudflat sediment (inferred from tidal loading). One-dimensional compression testing of low marsh and mudflat materials obtained from the intertidal zone at Greatham Creek has confirmed that both sediments are overconsolidated, and indeed to different degrees. However, there are clear differences between the predicted and observed values of preconsolidation stress; a mean preconsolidation stress of 23.88 kPa was observed in the low marsh and 10.86 kPa in the mudflat. It is therefore evident that the hydrological variables monitored in this thesis (tidal- and groundwater change-induced loading) have a less important influence on preconsolidation stress than was anticipated. Indeed, it is reasonable to suggest that the differences between predicted and observed values of preconsolidation stress are attributable to the duration of flooding experienced at each sample site. Low marsh sediments were sampled from an altitude of 2.26 m OD, at which flooding duration is *c.* 6 % of total time. Mudflat sediments were obtained from 1.06 m OD, which is flooded for *c.* 35 % of total time. The reduced hydroperiod at the low marsh sample site results in dramatically increased duration of exposure to desiccating subaerial processes in comparison to the mudflat sampling altitude, which is flooded on the majority of tides. Desiccation results in a reduction in moisture content and increases effective stress within the soil (Marinho and Chandler, 1993; Tovey and Yim, 2002) without overburden sedimentation and leads to overconsolidation, as observed in the stratigraphies of southern Essex and the inner Thames by Greensmith and Tucker (1971a; 1971b; 1973; 1976) (Section 3.9.9), for example. Additional effective stresses can be caused by plant root suction on the vegetated saltmarsh surface during evapotranspiration (Mathur, 1999). The lower preconsolidation stress observed in mudflat sediments is perhaps more likely to be a result of capillary suction stresses induced by groundwater falls and a lesser degree of desiccation during periods of subaerial exposure.

Within-material variability in preconsolidation stress results from small-scale spatial and temporal differences in the effective stress acting at the depositional surface. Such differences are inevitable given the range of stress-inducing processes operating in intertidal areas.

6.3.3 Creep processes

Variations in the duration of loading increments allowed inferences to be drawn regarding the influence of creep on compression behaviour.

Differences between conventional, 24 hour load increment tests and those with extended or truncated creep phases are perhaps subtle and did not result in dramatic differences in the general form of the $e\log_{10}\sigma'$ plots. This may be due to the differences in load increment duration not being great enough to effect stark differences in compression behaviour. Despite this, however, slight variations were observed. The first of these involves an apparent variation in the position (intercept) of the normal compression line. Samples LM-5-S-OED-NC and MF-5-S-OED-NC generally plot higher (above) samples that are subjected to creep. This trend is also evident in the plots of normalised voids ratio against vertical effective stress (Figures 6.23 and 6.24), with 'consolidation only' samples (LM-5-S-OED-NC and MF-5-S-OED-NC) plotting above the remaining samples subjected to creep. This indicates a lesser degree of compression, and so are less likely to be attributable to memory effects associated with the initial voids ratio. Such a finding is unsurprising, given that these samples are subjected to less opportunity for compression. At the other end of the spectrum, increased load increment times result in a greater degree of compression and a downward shift in the location of the normal compression line. This is particularly well demonstrated by differences between the normal compression lines of samples MF-4-S-OED-48 and MF-6-BPS-24 (samples of near-identical initial voids ratios). These trends conform to conventional soil mechanics theory on the effects of creep on compression behaviour, as conceptualised by Bjerrum (1967) (Section 3.7, Figure 3.13).

A second trend that was observed in sample LM-5-S-OED-NC involves an increase in the preconsolidation stress and increase in the value of the compression index in relation to the other samples analysed. This finding is in contrast to experiments undertaken on clay lithologies in which increased creep was observed to lead to increased structural resistance to compression and a subsequent increase in the value of the preconsolidation stress (Burland, 1990; Das, 1998). However in these tests on clay, normal compression lines were significantly steeper, as observed in sample LM-5-S-OED-NC.

The steepness of the normal compression line of sample LM-5-S-OED-NC indicates a convergence on the 'Omega point' (Tovey and Paul, 2002). Regardless of the settlement mechanism (i.e. primary consolidation or creep), it seems that the same overall magnitude of compression is occurring, but the style and rate (in relation to vertical effective stress) of compression may differ significantly. However, it would be unwise to state that these apparent creep-induced differences in behaviour irrefutably apply to all low marsh materials subjected to loading with truncated creep phases, since the differences could well be a result of the natural variability of the material. Further investigation is clearly required.

Additional insights into the importance of creep processes are gained from the results of the time-settlement plots and cyclic loading tests presented in Sections 6.4 and 6.5.

6.3.4 Estimations of the compression index from liquid limit tests

Decompaction techniques based on Terzaghi's compression law (outlined in Section 3.4) require a value of the compression index, C_c, for the constituent algorithms. It is often the case that C_c is not obtained at the time of sample collection, as noted by Paul and Barras (1998) and Massey *et al.* (2006). In addition, obtaining samples in a pristine, undisturbed condition for reliable geotechnical testing is difficult, particularly when the samples of interest are at considerable depth, prohibiting block sampling from an excavation or trench. In such situations, it is necessary to estimate the compression index indirectly, *via* the correlation between liquid limit and compression index observed by Skempton (1944) (Equation 3.7). As previously noted by Paul and Barras (1998), Tovey and Paul (2002), and Massey *et al.* (2006), Skempton's (1944) correlation is strictly only applicable to fine-grained plastic soils. Indeed, the data presented in this chapter corroborate this fundamental requirement. From the liquid limits reported in Table 5.2, compression indices predicted from Equation 3.7 are 0.87 for the low marsh and 0.51 for the mudflat. These figures still reflect differences in compressibility (higher value for the low marsh). However, the predicted value for the low marsh grossly underestimates the directly observed (oedometric) mean (1.89) and even the minimum (1.41) values of the compression index displayed in Table 6.7. The removal of field structures during mixing prior to the liquid limit test seemingly results in a significantly lower estimate of the compression index in the low marsh material. Indirect (liquid limit) methods result in an estimate of the compression index that perhaps describe the virgin compressibility of the remoulded sediment, which is related to the mineralogy of the inorganic component and free from the contribution to structure of *in situ* organic growth.

The value of the compression index derived for from liquid limit tests on mudflat sediments overlaps with the lower end of the oedometer-derived values of the compression index displayed in Table 6.7 (minimum = 0.50, mean = 0.60). Perhaps further tests would reveal a greater range of compression indices obtained from the liquid limit correlation.

These results highlight the importance of structural variation created by *in situ* organic growth; the potential pitfalls of using orthodox soil mechanics theory without sufficient adaptation; and the fact that there is no substitute for high-quality geotechnical testing of materials to obtain an accurate description of compression behaviour.

6.3.5 Implications for autocompaction modelling

Basic one-dimensional compression tests have revealed significant differences with existing models of autocompaction which are based on conventional soil mechanics theory that was originally developed on clay lithologies.

Terzaghi's compression law (Section 3.3.1) states that the change in voids ratio (and hence volume) of a material is directly proportional to the common logarithm of the increase in effective stress acting on the material during virgin compression, as illustrated in Figures 3.6, 3.8 and 3.9. Section 3.6 discussed how this model is based on a number of assumptions. A critical prerequisite of this model is that the materials are not overconsolidated and therefore their compression behaviour must follow the normal compression line from structural density to zero porosity. Clearly, the stress environment of the intertidal zone deviates from that for which Terzaghi's compression law was developed, and both of the materials studied have been shown to be overconsolidated, albeit to different degrees. Hence, a backwards extrapolation of the normal compression line to the intercept is an incorrect representation of compression behaviour prior to the preconsolidation stress. By using Terzaghi's compression law to pre- or retro-dict the thicknesses of various strata (as done by Massey *et al.*, 2006; Paul and Barras, 1998; and Tovey and Paul, 2002), changes will be greatly overestimated due to the rapid change in voids ratio at low effective stresses predicted by this model. Instead, changes at low effective stresses are minimal because of overconsolidation.

Also implicit in Terzaghi's compression law is that individual materials are of constant lithology and compression behaviour. The materials studied here are arguably as lithologically homogeneous as are found naturally (Table 5.2). However, the results presented above illustrate that compression behaviour of a material cannot adequately be described by single values for rheological parameters. Both the swell and compression indices in fact describe an idealised 'best-fit' line for each sample tested, with the actual values of each respective index varying slightly according to the vertical effective stress applied to a sample. This is shown by the increase in gradient in the recompression line before the preconsolidation stress is reached and the slight flattening of the normal compression curves at higher effective stresses, as illustrated in Figures 6.9 and 6.17. As a result of this 'smoothing' of the recompression and normal compression lines, any variability in the values of C_r and C_c are masked and an inherent prediction error is introduced when using a single value of each of these parameters. In addition to this within-sample variability, a considerable degree of between-sample variability in initial structure and compression behaviour is evident that does not appear to be related to minor variations in lithology (organic, sand, silt and clay contents). Furthermore, the variations in structure and compression behaviour were observed in sediments obtained from a small sampling area of the intertidal zone at Greatham Creek. Rapid changes in structure and compression behaviour through space at the depositional surface will undoubtedly translate into rapid stratigraphic variations within a sedimentary sequence.

These points clearly raise questions regarding the wisdom of applying a single value of the compression index to a geotechnically variable, although lithologically constant, material. The implications become increasingly evident when considered in relation to existing methodological and modelling approaches. The current procedure begins by splitting a stratigraphic column into a number of layers, within which the geotechnical and lithological properties are assumed to be constant (Massey *et al.*, 2006; Paul and Barras, 1998; Pizzuto and Schwendt, 1997; Tovey and Paul, 2002). The ideal situation would be to obtain one-dimensional compression tests and voids ratio measurements at the highest possible resolution; 2 cm, for instance, to allow an oedometer test to be undertaken on the material. Such an approach is idealistic given the time that would be required to test at this resolution throughout even a short (e.g. 1 metre) core. Instead, it has been common practice to obtain a value of the compression index for a given layer (Massey *et al.*, 2006, use a depth resolution of 0.1 to 0.2 m to obtain values of the compression index) based on a single oedometer or liquid limit test and apply this value to the entire unit. Due to the observed variation the compression behaviour in surface sediments, this approach is inadequate, since rapid downcore variation in compressibility will take place in apparently homogenous lithologies. This variation will not be captured by a single value of the compression index and so pre- and retro-diction of sediment volume becomes increasingly prone to error.

Further problems arise in determining previous compression behaviour and volumetric change because of the observed destructuration in intertidal sediments. Crossing and convergence of normal compression lines and the removal of depositional structures decreases the accuracy of any estimates of volumetric change based upon their backwards projection. This technique will fail to account for initial structural variations that are subsequently lost upon increased exposure to vertical effective stress.

All things considered, it seems that Terzaghi's compression law, involving a backwards extrapolation of the normal compression line to 1 kPa, provides a poor representation of the one-dimensional compressive behaviour of intertidal sediments. A combination of low sedimentation rates, high durations of subaerial exposure and falls in groundwater result an immediate (surface) stress history and rapid overconsolidation of materials following sedimentation. It is thus difficult to imagine a situation in the upper intertidal zone where sediments remain normally consolidated in near-surface sediments. Terzaghi's compression law therefore seems to describe 'impossible states' when applied to intertidal sediments. Furthermore, a significant degree of structural and behavioural variation was observed in the sediments. As a result, single values of compression indices to describe one-dimensional volume changes do not fully encapsulate the range of behaviour that has been observed to occur in surface sediments of near-identical lithology.

Clearly, this basic model has to be re-evaluated in the context of the intertidal environment and the materials under investigation; Terzaghi's compression law must be adapted for realistic application to the volumetric evolution of intertidal sediments. Firstly, the model must be modified to describe the overconsolidation observed at the depositional surface. Secondly, it is necessary to develop a statistical model that is capable of describing inherent variations in structure and behaviour that are independent of lithological variability.

However, formulating such a model is not possible at this stage of the study; the $e\log_{10}\sigma'$ plots presented above and the associated findings can be seen to be quite arbitrary because only short (on a geomorphological timescale at least) load increment durations have been considered. The extended and truncated load increment duration tests suggested that creep may have an important influence on the form of an $e\log_{10}\sigma'$ plot. It is therefore firstly necessary to consider the effects of time-dependent factors on the material behaviour. In addition, further insights into the idiosyncratic behaviour in intertidal sediments are provided by the dynamic loading tests.

6.4 TIME-SETTLEMENT BEHAVIOUR OF MINERALOGENIC INTERTIDAL SEDIMENTS

Time-settlement analysis is critical to evaluations of compression behaviour. It is from plots of logarithmic time against vertical displacement (where negative displacement indicates compression/volumetric reduction and positive displacement indicates swelling/volumetric increase) that the relative importance of the three causes of settlement can be established (Figures 3.10 and 3.11). Furthermore, it is from these graphs that rates of primary consolidation and creep can be ascertained; these can then be scaled and/or extrapolated to determine the time-rate of field settlement as a result of each process (Sections 3.5 and 3.7; Figure 3.12). Plots of time against vertical displacement for each load increment and decrement are shown in Figures 6.25 to 6.53 for the low marsh and Figures 6.54 to 6.71 for the mudflat. Both logarithmic time and square-root time are used, since it is easier to discern behavioural trends and differences in their rate of operation if both timescales are used in synergy.

Before the time-vertical displacement plots are presented, it is important to note that the abscissa corresponding to zero time cannot be plotted on a logarithmic scale. In order to overcome this, compression was only permitted to commence after the Ds7 software had begun logging. However, where this practice did not occur due to error, compression lines plotted on a logarithmic scale depart significantly from samples for which the initial reading was taken at a time greater than 0 minutes (compare, for example, samples LM-3-S-OED-24 with LM-7-N-OED-24 in Figure 6.26). Although this phenomenon prevents accurate determination of the extent of initial compression, it obviously makes no difference to the rate of subsequent primary consolidation and creep processes. Accordingly, little significance is assigned to deviations in graphical form at early stages in individual tests (i.e. less than *c.* 0.5 minutes) that arise from 'logging' error.

6.4.1 Time-vertical displacement analysis of low marsh material

The lowest effective stress applied to the low marsh sediments was 5 kPa. The time-vertical displacement plots for this loading stage are displayed in Figure 6.25. The low marsh samples display considerable variation in both rates and magnitudes of settlement reflecting initial structural characteristics and compressibility. It is difficult to distinguish between various settlement processes because breaks in curvature are not easy to identify; curves bear little resemblance to conventional Terzaghi plots (Figures 3.10 and 3.11). However, the majority of samples do display a very rapid (near-instantaneous) initial compression or rapid primary consolidation phase. Since the material is overconsolidated, samples have previously been exposed to this level of effective stress. The structure is therefore already likely to be in equilibrium with the applied effective stress; any excess pore water pressures that develop following the application of load are unlikely to be greatly in excess of pre-loading values. Furthermore, these excess pore water pressures will dissipate rapidly due to the high permeability of these sediments, as suggested by high voids ratios. As a result, distinguishing between initial compression and primary consolidation phases is difficult early in the loading stage due to the rapid rate of operation. Whatever the cause, settlement rates quickly drop to a lower value after this initial rapid settlement and begin to follow (what is likely to be) a creep settlement curve from *c.* 100 minutes. This rate is similar in each of the samples with the exception of LM-8-N-OED-24, which has a steeper secondary compression section. Indeed, LM-8-N-OED-24 has the greatest total settlement, with a greater contribution from the initial rapid settlement phase noted above. This may be due to this sample being initially unsaturated (degree of saturation = 93.01 %) when the load was first applied, leading to compression of gas pockets. However, sample LM-7-N-OED-24 was also unsaturated during this loading phase and shows less compression despite having a similar degree of saturation (94.51 %). It seems more probable that variations in the magnitude of total settlement during the 5 kPa loading phase result from initial structural variations.

Low marsh samples that were subjected to a vertical effective stress of 10 kPa (Figure 6.26) all show a similar graphical form. However, these forms do not resemble orthodox Terzaghi curves. As is more obvious in the square-root time plot (Figure 6.26 b), a rapid initial settlement and/or primary consolidation phase flattens to a largely constant creep segment. Differences in the total magnitude of settlement reflect initial structural characteristics to a certain extent (compare with Table 6.5).

The 14 kPa loading stage (Figure 6.27) is the first to show the effects of variation in load increment ratio on compression processes. Samples with lower load increment ratios (LM-2-S-OED-24, LM-3-S-OED-24, LM-7-N-OED-24 and LM-8-N-OED-24) plot higher, indicating less overall settlement. In contrast, samples LM-1-S-OED-24, LM-5-S-OED-NC and LM-6-S-BPS-24, which were not subjected to the 10 kPa loading phase, undergo a greater degree of settlement for this (14 kPa) loading stage. These observed differences in total settlement of these samples are to be expected. Samples with the lower load increment ratios had previously undergone a proportion of the total settlement experienced by those which had higher load increment ratios and so required less structural adjustment and volumetric reduction. Interestingly, samples LM-1-S-OED-24, LM-5-S-OED-NC and LM-6-S-BPS-24 have a greater rate of creep than samples with an intermediate loading phase. This is likely to result from an increase in structural resistance to compression in samples LM-2-S-OED-24, LM-3-S-OED-24, LM-7-N-OED-24 and LM-8-N-OED-24 that developed during fabric rearrangements which themselves occurred during the creep phase of the previous (10 kPa) loading increment. Also of interest in this loading stage is that samples LM-1-S-OED-24, LM-5-S-OED-NC and LM-6-S-BPS-24, which have comparatively higher load increment ratios, show square root time–vertical displacement plots that are beginning to resemble the theoretical curves produced by Terzaghi's consolidation equation. The greater load increment ratio has resulted in greater excess pore water pressures, an increase in the duration of time required for these excess pressures to dissipate to equilibrium values and so a more obvious straight-line portion of the curve.

14 kPa is the first loading stage applied to sample LM-6-S-OED-68. Given the trends outlined in the paragraph above, it would be logical to assume that this sample would undergo the maximum amount of settlement of all the samples subjected to this vertical effective stress. However, the amount of settlement is considerably less than that of samples that were loaded by the initial 5 kPa increment. As was observed in Section 6.2, this sample has the most dense, and therefore stable, initial structure (e_i = 3.69). Unsurprisingly, therefore, a lower amount of compression is observed during the 14 kPa loading stage, despite no previous laboratory load application. Very little volumetric adjustment is needed for the structure of this sample to be in equilibrium with the overburden stress.

Loading of samples LM-2-S-OED-24, LM-7-N-OED-24 AND LM-8-N-OED-24 at 20 kPa again represents a low load increment ratio (0.43; Table 6.3). As a result, there is no obvious primary consolidation phase; a creep phase follows a pronounced initial compression phase (Figure 6.28). The magnitude of total settlement is directly related to the initial structure of the material. LM-2-S-OED-24 had the lowest initial voids ratio (3.98) of the three samples subjected to this vertical effective stress and had undergone the least compression for this load increment (*c.* 0.03 cm). Conversely, LM-8-N-OED-24 had the highest initial voids ratio of the three, and accordingly has compressed the most.

As was noted for the 14 kPa loading stage, differences in load increment ratio have resulted in variations in the form of the time-settlement plots of samples subjected to a vertical effective stress of 26 kPa (Figure 6.29). Samples LM-2-S-OED-24, LM-7-N-OED-24 AND LM-8-N-OED-24 underwent loading at this stage with a load increment ratio of 0.3, compared to a load increment ratio of 0.86 for samples LM-1-S-OED-24, LM-3-S-OED-24, LM-5-S-OED-NC and LM-6-S-BPS-24 (Tables 6.2 and 6.3). Once again, samples LM-2-S-OED-24, LM-7-N-OED-24 AND LM-8-N-OED-24 do not seem to display a primary consolidation phase, entering a creep phase after a minimal amount of initial compression/primary consolidation. In contrast, samples LM-1-S-OED-24, LM-3-S-OED-24, LM-5-S-OED-NC and LM-6-S-BPS-24 show a rapid and pronounced primary consolidation phase which, without a break in the curvature, gradually undergoes a transition into a creep phase. It is interesting to note that the rate of creep that is occurring is similar in all samples despite variation in the magnitude of primary consolidation caused by differences in the load increment ratio.

The time-settlement response to an application of a vertical effective stress of 50 kPa is generally very similar in all samples (Figure 6.30). Once again, however, the plots display only a partial similarity to those generated by Terzaghi's consolidation equation. The logarithmic time–vertical displacement plot (Figure 6.30 a) shows no point of inflection (reversal of curvature) and the square root time–settlement plot (Figure 6.30 b) is continuously curved rather than showing distinct primary consolidation and creep components. The main differences in the magnitude of settlement are a result of variations in the contributions of primary consolidation; sample LM-8-N-OED-24 undergoes the most primary consolidation and LM-4-S-OED-68 the least. Indeed, the degree of primary consolidation settlement seems to be partially related to initial structural variation, with initially denser samples undergoing less volumetric reduction. However, this relationship is a general one and not all samples adhere to this. It is interesting to note that rates of creep settlement are largely uniform when primary consolidation has ceased. Similar patterns can be observed during the 99 kPa loading stage (Figure 6.31) (i.e. clear differences from the form of orthodox Terzaghi plots, variations in primary consolidation settlement affecting the overall degree of compression, a possible structural control on the magnitude total settlement and similar post-consolidation creep rates).

It is only during the 197 kPa loading stage (Figure 6.32) that samples begin to resemble the orthodox Terzaghi plots. In this stage and during the 393 kPa, 785 kPa and 1570 kPa loading stages (Figures 6.33, 6.34 and 6.35 respectively), reversals in the curvature occur on the logarithmic time–vertical displacement plots, although they are less pronounced in the 197 kPa loading stage (Figure 6.32 a). Similarly, the square root time–vertical displacement plots show more discrete primary consolidation and creep phases; in the 197 kPa loading phase (Figure 6.32 b), the transition between the two processes is difficult to define since the curves are still continuously curved. However, the curved nature of the plots becomes decreasingly evident at the higher effective stresses (Figures 6.33 b, 6.34 b and 6.35 b). Indeed, from effective stresses of 393 kPa and upwards, the time-settlement plots distinctly resemble conventional Terzaghi plots.

Also evident during exposure to effective stresses of 197 kPa and above is the increased convergence and similarity of the time-settlement behaviour of different samples. In contrast to loading stages below 197 kPa, there is a generally uniformity of response in terms of the style, rate and magnitude of settlement. This finding corroborates inferences drawn from the $e\log_{10}\sigma'$ plots regarding the destructuration of the sediments and an observable decrease in the influence of initial structure and density.

There are two notable exceptions to the increased homogeneity of behaviour at effective stress including and greater than 197 kPa. The first is sample LM-7-N-OED-24 during the 197 kPa loading stage (Figure 6.32). This sample displays a much greater degree of compression that does not conform to the behaviour of the remainder of the low marsh samples, although the general form is similar, particularly when considered in the square root time – vertical displacement plot (Figure 6.32 b). This variation in behaviour does not appear to be linked to any of the physical properties of the material (Tables 5.2 and 6.5) and a explanation for this different behaviour is difficult to find and so is attributed to natural variability.

The second exception to the general 'clustering' of samples in loading stages at vertical effective stresses of 99 kPa, 197 kPa and 393 kPa is the behaviour of sample LM-5-S-OED-NC (truncated creep phase). In each of these loading stages (Figures 6.31, 6.32 and 6.33 respectively), LM-5-S-OED-NC consistently undergoes the greatest rate and magnitude of settlement as a result of primary consolidation. Indeed, during loading by 99 kPa and 393 kPa, the sample is the most compressible despite having a reduced time for creep settlement to act.

A further consistent pattern of behaviour is that displayed by sample LM-4-S-OED-68. At vertical effective stresses at 50 kPa and above, this sample is consistently the least compressible sample despite the extended creep phase. As was suggested in Section 6.2, initially denser samples (LM-4-S-OED-68) retain their compact depositional structures during subsequent loading and remain less susceptible to compression. It is also noteworthy that the rate of creep of sample LM-4-S-OED-68 is similar to samples with 24 hour load increment durations and shows no signs of acceleration or deceleration at any effective stress level.

Further evidence for the proposed destructuration of samples, loss of their initial depositional structures and increased uniformity in time-vertical displacement behaviour is provided by the time-settlement behaviour of the low marsh samples during unload and reload stages (Figures 6.36 to 6.43 for unload behaviour; Figures 6.44 – 6.53 for reload behaviour). Swelling is essentially the reverse of the consolidation process and describes the volumetric increase of soil samples due to the absorption of water within the voids when the applied stress is reduced (Head, 1988). The swelling process occurs due to the tension forces within the soil skeleton upon removal of load (Powrie, 2004).

The time-swelling behaviour following the removal of overburden loads is similar in form in all low marsh samples (Figures 6.36 to 6.43); each phase of swelling is analogous to each of the compression processes outlined in Section 3.7. A small initial swell phase is followed by conventional primary swelling (water absorption) phase. When all existing suction stresses have equilibrated, the soil will continue to increase in volume in a reversal of the creep process; the soil skeleton will begin to readjust to the new lower ambient effective stress, undergoing time-dependent structural changes to a looser state of packing. However, as was noted in the unload-reload behaviour in analysis of $e\log_{10}\sigma'$ above, the material is only partially elastic and will not recover the full strain experienced during virgin compression.

This most pronounced display of uniformity in unload-reload time-settlement behaviour can be observed in samples LM-7-N-OED-24 and LM-8-N-OED-24. Both of these underwent the same unload-reload stages and effective stresses. Despite differences in initial voids ratio and virgin time-settlement compression behaviour, these samples behave almost identically when unloaded at effective stresses of 132 kPa, 46 kPa and 16 kPa (Figures 6.38, 6.41 and 6.43 respectively). This homogeneity of behaviour can also be observed in the reload behaviour, at effective stresses of 46 kPa, 132 kPa and 393 kPa (Figures 6.44, 6.46 and 6.50 respectively). Such uniformity cannot be observed in the other samples. Since samples LM-7-N-OED-24 and LM-8-N-OED-24 were initially unsaturated, perhaps a more significant degree of destructuration occurred when these samples were loaded prior to 10 kPa, resulting more homogeneous behaviour. However, such a conclusion is speculative and cannot be declared indisputable until further work is carried out.

Samples LM-3-S-OED-24 AND LM-6-S-BPS-24 underwent the same unload-reload stages: unload stages of 259 kPa, 87 kPa and 30 kPa (Figures 6.36, 6.39 and 6.42 respectively) and 87 kPa, 259 kPa and 758 kPa (Figures 6.45, 6.48 and 6.52). These samples also display highly uniform swelling characteristics indicative of a destructured soil fabric. The exception to this is the 30 kPa loading stage in which LM-3-S-OED-24 underwent less immediate swell upon removal of the load than LM-6-S-BPS-24.

The importance of the (un-/re-) load increment ratio is highlighted again by the unload-reload time-vertical displacement behaviour of samples LM-1-S-OED-24, LM-2-S-OED-24 and LM-4-S-OED-68. LM-2-S-OED-24 is the only sample for which the standard unload-reload loading stage method, involving the division of existing loading mass by three (outlined in Section 6.1.1), was not applied. Instead, this sample (subjected to a maximum vertical effective stress of 295 kPa) underwent the same unload-reload sequence as samples LM-2-S-OED-24 and LM-4-S-OED-68 (subjected to a maximum vertical effective stress of 1570 kPa). Clearly, the unload increment ratios differ in this situations. The effect of this difference in unloading sequence is illustrated in Figure 6.37 when unloaded to 172 kPa, samples LM-2-S-OED-24 and LM-4-S-OED-68 underwent a much greater degree of all swell processes. In contrast, the decrease in effective stress following unloading from 197 kPa to 172 kPa was much less in sample LM-2-S-OED-24, resulting in less swelling and noticeably different swelling behaviour.

6.4.2 Time-vertical displacement analysis of mudflat material

Loading by a vertical effective stress of 5 kPa (Figure 6.54) results in a large range of compression behaviour linked to initial structural variations. By comparison with Table 6.5, it is evident that samples with higher initial voids ratios are compressing to a greater extent, and indeed more rapidly, than those with denser depositional structures. For example, sample MF-4-S-OED-48 has the most dense initial structure (e_i = 1.71) and consequently experienced relatively little compression in comparison to, for example, sample MF-1-S-OED-24. This sample had an initial voids ratio of 2.34 and experienced the second greatest amount of overall settlement. Sample MF-7-N-OED-24 underwent the most compression during this loading stage, possibly due to its initial unsaturation and a resultantly increased capacity for compression *via* a reduction in the volume of pore air. Sample MF-6-S-BPS-24 cannot strictly be compared with the oedometer samples during this load increment since it underwent saturation at a vertical effective stress of *c.* 3 kPa. This would have undoubtedly led to some settlement and consequently less compression during subsequent loading at an effective stress of 5 kPa. Interestingly, and in stark contrast to the time-settlement behaviour of the low marsh materials at low stresses, the majority of mudflat samples show orthodox Terzaghi behaviour even at this low effective stress and in the overconsolidated range of the material (the mean preconsolidation stress for the mudflat was estimated at *c.* 11 kPa). An exception to this is sample LM-5-S-OED-NC; due to the shortening of the duration of the load increment, the 'expected' Terzaghi form is perhaps not obtained. Nonetheless, this finding means that Terzaghi's consolidation theory is applicable even during the overconsolidated section of the loading phase.

Samples subjected to a vertical effective stress of 10 kPa (Figure 6.55) do not show fully orthodox Terzaghi behaviour, particularly on plots of vertical displacement against logarithmic time; an inflection point in the straight line portion of the curve is not evident. When plotted in terms of square root time, however, greater similarity to conventional plots is evident, although the primary consolidation and creep segments are transitional rather than distinct due to the continuously curved nature of the lines.

At a vertical effective stress of 14 kPa (Figure 6.56), two types of time-settlement behaviour are evident. The difference in the form of the time–settlement plots can be attributed to differences in load increment ratio. Samples subjected to 14 kPa without the intermediate 10 kPa loading stage (MF-4-S-OED-48, MF-5-S-OED-NC and MF-6-S-BPS-24) have load increment ratios of 1.8 (Table 6.2). These samples show conventional Terzaghi behaviour. In contrast, samples MF-1-S-OED-24, MF-2-S-OED-24, MF-3-S-OED-24 and MF-7-S-OED-24 undergo less compression during 24 hours due to the previous compression that took place during the intermediate 10 kPa loading stage. Additionally, the recorded time-settlement behaviour of these is not consistent with standard Terzaghi time-settlement plots due to the lower load increment ratios ($\Delta\sigma/\sigma$ = 0.4). Despite variations in form and magnitude of settlement, post-consolidation phase creep rates are similar, with the possible exception of MF-1-S-OED-24.

The slightly erratic time–settlement behaviour displayed by MF-6-S-BPS-24 during the 14 kPa loading stage (Figure 6.56) was possibly due to the potentially variable hydraulic control on effective stress in the back-pressured shear box apparatus. Little attention is therefore given to this irregularity.

With the exception of the 26 kPa loading stage in which differences in load increment ratios result in variability in the graphical form of time-settlement behaviour, time-settlement plots from loading stages from 20 kPa to 1570 kPa (Figures 6.57 to 6.64) all show similar behaviour that conforms to conventional Terzaghi theory. In addition, there is a reduction in the variations of rate and magnitude of time-settlement behaviour, corroborating notions of post-yield destructuration and a consequent increase in behavioural homogeneity. For individual load increments, rates of creep are very similar in each sample (i.e. little inter-sample variability for a given loading stage). Moreover, the rate of creep observed for each loading stage for sample LM-5-S-OED-48 (extended load duration) does not display acceleration or deceleration of the pre-24 hour rate.

Sample MF-5-S-OED-NC behaves in a conspicuously different manner to the remaining mudflat samples, particularly at post-yield effective stresses (i.e. 14 kPa – 1570 kPa). This sample undergoes the greatest rate and magnitude of primary consolidation. Indeed, with the exception of the 50 kPa loading phase, the sample experiences the greatest amount of settlement in spite of the truncated creep phase.

Analysis of unload-reload time-vertical displacement behaviour for the mudflat samples substantiates findings from the analysis in $e\log_{10}\sigma'$ space in Sections 6.2.3 and 6.3.2; namely that the mudflat materials do not undergo the same degree of destructuration as the low marsh sediments and retain some 'memory' of their depositional structures. Figures 6.65 to 6.71 display the time-vertical displacement behaviour of samples during unloading and reloading. A greater range of behaviour than was observed in the low marsh sediments during unload and reload stages can be seen. However, the vertical displacement over time during unloading is consistent in form, with an extremely rapid primary swell phase that quickly switches to a creep phase that is operating at a very low rate, though this rate is consistent between samples for each unloading stage.

Reload behaviour shows an almost exact reversal of the unload behaviour. This is unsurprising given the well-documented elastic behaviour of the unload-reload hysteresis loop (e.g. Powrie, 2004). Reload behaviour conforms to conventional Terzaghi theory, particularly in plots of vertical displacement against square root time (Figures 6.69 to 6.71). The erratic nature of the lines for the majority of samples may be attributable to a possible elasticity in the samples, but is more likely to result from unwanted defects in the mechanical operation of the oedometer, such as an increase in side-friction on the oedometer cutting ring.

6.4.3 Time-vertical displacement behaviour of intertidal sediments

From the results presented in Sections 6.4.1 and 6.4.2 it is evident that variations in time-vertical displacement behaviour exist as a result of a number of material- and stress-related factors.

As discussed in Section 6.3, structural variations exist in the sediments that result in increased compressibility in samples that are initially less dense (higher initial voids ratios). This phenomenon is modified by the stress history of a particular sample, with decreased variability in the rates and magnitudes of compression occurring at vertical effective stresses greater than the preconsolidation stress. This reflects the destructuration process referred to previously that results in a decrease in the influence of depositional structures and an increased homogeneity of time-vertical displacement behaviour. As was apparent in the $e\log_{10}\sigma'$, destructuration occurs to a lesser extent in the mudflat material; at effective stresses greater than the preconsolidation stress and during both unload and reload phases mudflat samples tend to retain a greater degree of inter-sample variability. Low marsh samples tend to show a greater degree of post-yield uniformity in behaviour, reflecting a greater degree of destructuration. It is necessary to exercise some caution, however, when comparing trends between samples at higher stresses, since a greater number of low marsh than mudflat samples exceeded the compression limit of the oedometer apparatus. As a result, there are a greater number of mudflat samples that represent higher values of effective stress; the greater variation in behaviour of the mudflat samples may be an artefact of sample numbers and the size of the dataset.

Despite the observed variations in magnitude of settlement during loading, samples subjected to the same loading stages displayed similar intra-material behaviour. Low marsh samples rarely displayed conventional Terzaghi behaviour in either logarithmic time- or square root time–vertical displacement plots. This was certainly the case in the overconsolidated (i.e. < 26 kPa) phase of the materials. Within this stress range, the vertical displacement response was highly varied and unusual. At stresses greater than the preconsolidation stress, this unconventional time–vertical displacement behaviour continued, although the style of deformation was consistently displayed within different samples. The logarithmic time–vertical displacement plots showed no obvious inflection point at which the curvature changed from convex upwards to concave upwards. Furthermore, the square root time–vertical displacement plots were continuously curved, showing no distinct boundary between primary consolidation and creep processes and no clearly-defined straight-line portion. The continuously curved nature of the plots reflects the simultaneous operation of consolidation and creep processes and their complex interactive effect on structure. As water drains out of the pores due to the production of excess pore water pressures, the partially organic soil skeleton itself is being compressed and rearranged, modifying compressibility and permeability 'mid-stage'; this challenges one of the fundamental assumptions of Terzaghi's theory of consolidation outlined in Section 3.6. Such a violation in assumption and such significant departures in graphical form from Terzaghi's theoretical curves render the use of Terzaghi's consolidation theory inappropriate on the low marsh samples. Values of c_v, the coefficient of consolidation, cannot be obtained from these time-vertical displacement plots since they do not possess the required straight line portions and breaks in curvature required by the established Casagrande (logarithmic time) and Taylor (square root time) curve-fitting methods (Figures 6.72 and 6.73 respectively).

It is only at higher effective stresses (> 393 kPa) that the low marsh time–vertical displacement curves show conventional Terzaghi form. This may indicate a decrease in the influence of organic matter (*via* creep processes) in altering compressibility and permeability during the consolidation phase. However, although of interest in a geotechnical sense, such a finding is of little use to estimates of elevational adjustment in intertidal deposits since these high stresses are above those typically experienced in the majority of Holocene coastal stratigraphic sequences.

In contrast to the low marsh, mudflat samples consistently showed evidence of Terzaghi time-vertical displacement behaviour, even at effective stresses below the preconsolidation stress. The mudflat material conforms to the assumptions of Terzaghi's consolidation law and undergoes time–vertical displacement behaviour that can be analysed using conventional curve-fitting methods, allowing values of c_v to be obtained.

There are, however, exceptions to this finding. The influence of the load increment ratio is seemingly critical. In the mudflat samples, where conventional Terzaghi behaviour is generally evident, this only occurs when the load increment ratio is greater than, or close to, unity. Smaller load increment ratios result in fundamentally different graphical representations of time-vertical displacement behaviour which do not resemble conventional Terzaghi plots and so prevent successful curve-fitting and estimation of c_v. The fact that the load increment ratio has such a profound effect on the shape of the time-vertical displacement curve has previously been noted by Leonards and Girault (1961, cited in Hobbs, 1986). In a study on the Mexico City clay, they observed that, as a general rule, load increment ratios exceeding unity tend to produce curves of the type marked I in Figure 6.74 which they appropriately name Terzaghi, or Type I, curves. Leonards and Girault (1961) reiterated that these are characterised by a well-defined reversal of the slope in the logarithmic time-vertical displacement curves at the start of the distinctive creep 'tail', by which point the pore pressure has generally reached an infinitesimal value. In contrast, lower values of the load increment ratio tend to produce curves marked III in Figure 6.74, in which there is no reversal of the slope at or near the end of the primary consolidation stage. Such curves were observed in samples loaded using low stress loading scenario 2 (Table 6.3) (i.e. load increment ratios less than 1). No standard curve-fitting procedures have been developed for such curves; without direct measurements of pore pressure from the sample (which are not obtained in conventional oedometer testing), attempts to determine the exact time at which primary consolidation ceases would be speculative and misleading.

The behaviour of samples LM-5-S-OED-NC AND MF-5-S-OED-NC is of importance; the time-settlement plots of these samples, which were subjected to minimal creep, further display that there is no 'typical' behaviour for a material at a specific effective stress. Both samples undergo the greatest magnitude of post-yield primary consolidation settlement at the fastest rate due to a reduction in the stiffening of the material that is normally effected by the operation of secondary compression (creep). As was noted in Section 6.3.4, the volume of these materials seems continue to change by the same overall magnitude regardless of the causal process. This is essentially a stress history issue; the future behaviour of the material is partially controlled by previous exposure to effective stress and the degree of which structural resistance to compression has developed. In essence, the time-vertical displacement of a material is highly dependent on the duration of effective stress application and not solely on the effective stress level itself. Creep has an important influence on material behaviour during virgin compression.

6.4.4 *The contribution of creep to settlement*

To assess the relative contribution of initial compression, primary consolidation and creep processes to total settlement for a given load increment, compression ratios were calculated. These are calculated by dividing the observed settlement caused by each individual process by the total settlement observed for a single load increment, where the sum of the three compression ratios equals 100 %.

In 'conventional' materials, the settlement caused by each process can be determined reasonably accurately using the Casagrande and/or Taylor constructions (Figures 6.72 and 6.73). Indeed, these methods were used on time-vertical displacement plots obtained from the mudflat samples. Since these constructions could not be used on the time-vertical displacement plots obtained from the low marsh samples at stress lower than 393 kPa, the relative contributions of each of the settlement processes was estimated from logarithmic time-vertical displacement plots using the following methods, as outlined by Head (1988). The initial compression contribution is taken as all settlement that occurred prior to the first settlement reading being recorded (Figure 6.75), unless it is possible to directly observe initial compression phase (see, for example, Figure 6.29, sample LM-5-S-OED-24). The amount of settlement at the end of primary consolidation was calculated when a backwards extension of the creep segment of the curve (which shows a linear relationship with logarithmic time) first begins to diverge from the experimental line (Figure 6.75). This process is applicable to both concave and convex upwards plots.

Graphical representations of contributions to total settlement for each sample are displayed in Figures 6.76 to 6.82 for the low marsh material and Figures 6.82 to 6.87 for the mudflat material. Samples LM-5-S-OED-NC AND MF-5-S-OED-N are not included in this analysis due to the prevention of the operation of creep processes in these samples.

In the low marsh samples, the maximum contributions of initial compression to total settlement tend to occur at vertical effective stresses at which the material is overconsolidated (i.e. 5 – 26 kPa), reaching as much as *c.* 35 % in sample LM-7-N-OED-24 (Figure 6.81) at 14 kPa and even 40 % in LM-3-S-OED-24 (5 kPa) (Figure 6.78). Primary consolidation is nearly always the main cause of volumetric reduction in low marsh sediments, contributing on average to between 50 and 70 % of total settlement. There are occasional exceptions to this, such as during the 99 kPa loading stage of sample LM-2-S-OED-24, in which primary consolidation only contributes *c.* 25 % to total settlement, as opposed to the *c.* 65 % contribution from creep processes. It is the contribution of creep that is the of most interest, since it is often assumed to be negligible in the volumetric reduction of sediments (Section 3.7), despite the fact that the ongoing nature of this process has been shown to dramatically influence the nature and form of $e\log_{10}\sigma'$ plots (Bjerrum, 1967; Crawford and Morrison, 1996). Consistent trends in the contribution of creep to total settlement at particular effective stresses are difficult to observe. Indeed, there is no evidence for a consistent increase or decrease in importance of the process as overburden stress increases. The vertical effective stress at which creep is the most significant contributor to settlement are not constant in different samples, occurring at 5 kPa in LM-1-S-OED-24 (Figure 6.76), 99 kPa in LM-2-S-OED-24 (Figure 6.77) and 50 kPa in LM-4-S-OED-68 (Figure 6.79), for example. However, peak creep contributions do tend to occur at vertical effective stresses of 50 – 99 kPa.

In the mudflat samples, the contribution of initial compression is generally less than that observed in the low marsg (typically 5 – 10 % for the majority of loads), suggesting that greater the organic content of a material and the more open the initial structure, the greater the degree of initial settlement. In some samples, such as MF-1-S-OED-24 (Figure 6.83), MF-2-S-OED-24 (Figure 6.84) and MF-3-S-OED-24 (Figure 6.85), the relative contribution of initial compression total settlement increases. Once again, the main process that causes settlement over 24 hours is primary consolidation. It generally accounts for between 60 and 70 % of total settlement, particularly in the normally consolidated loading stages. Creep tends to contribute more to overall settlement at effective stresses lower than the preconsolidation stress (*c.* 11 kPa); this could be due to the overconsolidated state and the smaller load increment ratio at 10 kPa resulting in more rapid consolidation and hence a longer time for creep to exert an influence. Moreover, an increased creep component at lower stresses (5 – 20 kPa) may result from creep-like settlements produced by the closure of bioturbation burrows. However, consistent trends in the relative importance of creep are once again difficult to observe. Although the vertical effective stress at which creep is the most significant influence to total settlement is not constant, the largest contributions tend to be made during loading by either 10 kPa, 26 kPa or 393 kPa.

Despite the lack of consistent trends in terms of the relative contributions of various settlement processes in both the low marsh and mudflat materials, it is evident that creep cannot be ignored as a settlement process because it contributes significantly to the overall volumetric reduction of sediments, even during a 24 hour loading duration. Given the importance of creep to settlement over this relatively short time period, it is interesting to speculate on how the nature of $e\log_{10}\sigma'$ plots may be affected if creep processes were allowed to continue for longer durations. Before this can be done, C_α, the coefficient of secondary compression (Section 3.7) must be calculated for each sample at each loading (and also reloading) stage from plots of vertical displacement against logarithmic time, as outlined in Figure 3.12. The results of this have been plotted as graphs of C_α against vertical effective stress and are displayed in Figures 6.89 (low marsh) and 6.90 (mudflat). The mean value of C_α for both the unload and reload phases is presented (blue line); ± 1 standard error margins are also displayed to illustrate the observed variations in the value of C_α at each loading stage/effective stress. For comparative purposes, typical values of C_α are displayed in Table 6.8.

Values of C_α for the low marsh sediments are low (less than 0.005) below 20 kPa – values of effective stress that broadly correspond with the overconsolidated phase of the material. These values gradually rapidly rise to peaks values of *c.* 0.028 - 0.03 at effective stresses of 50 kPa and 99 kPa. These stresses at which peak values of C_α were recorded correspond with those for which the creep generally contributes the most to total settlement (Figures 6.76 – 6.82). After 99 kPa, values of C_α begin to decrease to a value of 0.01 at 1570 kPa. The greatest variability in creep rates occurred around the highest values of C_α. Values of C_α obtained from the reloading stages (0.001 – 0.009) are less than those of the virgin compression stages, indicating that stress history has an important influence on rates of creep. Values of C_α for the initial overconsolidated loading stages (i.e. < *c.* 25 kPa overburden) are similar to those from obtained from the reloading phase. In addition, variations in the observed values of C_α for the reload stages are generally less than those of the virgin loading stages, adding further weight to the argument that the sediments undergo a destructuration process that results in behavioural homogeneity.

Table 6.8 Typical values of C_α, the coefficient of secondary compression (Source: Lambe and Whitman, 1979).

Soil type	C_α
Overconsolidated clays (OCR > 2)	Less than 0.001
Normally consolidated clays	0.005 – 0.02
Very plastic clays	0.03 or higher
Organic clays	0.03 or higher

In general, values of C_α for the mudflat sediments are significantly less than those obtained from the low marsh sediments, as would be expected given the higher organic content in the latter (Hobbs, 1986). Moreover, values of C_α for different loading increments vary much less, ranging from a minimum of *c.* 0.004 (1570 kPa) to *c.* 0.0125 (26 kPa). It is interesting that rates of creep at higher effective stresses are lower than those observed during the initial overconsolidated stress range (< 11 kPa). The higher values of C_α at 5 – 20 kPa, and the greater variability in these values, are likely to be a result of the closure of bioturbation structures. Again, values of C_α, and variability therein, are lower for the reloading stages.

Lower values of C_α for reloading stages in both materials represent the fact that samples are overconsolidated and so have already been subjected to the applied stresses. Accordingly, their structures have already undergone plastic adjustment and reconfiguration to a more stable state. Hence, when reloaded to a previously experienced value of effective stress, the soil structure is already in structural equilibrium with the ambient stress and so does not experience rates of creep that were active upon virgin exposure to the specific level of effective stress.

The mean values of C_α for each stress increment during a test were used to predict the magnitude of settlement that would take place if the level of effective stress were to remain unchanged at a given load increment for the following durations: 1 day, 1 week, 1 month, 1 year, 10 years and 100 years. The estimated settlement was then converted to a voids ratio using the height of solids method and subtracted from the voids ratios observed in the laboratory at the end of the load duration (e.g. 24 hours) (after Nash *et al.*, 1992). The resultant series of curves are displayed in Figures 6.91 and 6.92. LM-1-S-OED-24 and MF-4-S-OED-48 were the samples upon which extrapolation was undertaken. Both plots show the dramatic influence that extended creep can have on the form of the $e\log_{10}\sigma'$ plots. Less variation is seen prior to the preconsolidation stress in both samples in accordance with the low values of C_α in the overconsolidated range. Post-yield variation, however, is dramatic, leading to an increase in values of the compression index with increased load increment durations. Extrapolated low marsh (Figure 6.91) normal compression lines diverge between 5 kPa and 200 kPa, before remaining essentially parallel until 1570 kPa. This is due to increases in the value of C_α up to 200 kPa, followed by decreases towards the higher stresses. In contrast, divergence of the projected normal compression lines is only evident between 5 kPa and 20 kPa in the mudflat sample (Figure

6.92). Again, this divergence coincides with peak values of C_α (maximum rates of creep) (Figure 6.90). At effective stresses above 20 kPa, normal compression lines are more parallel than those of the low marsh, reflecting less variation of C_α with effective stress (Hobbs, 1986). However, some divergence does exist, again leading to an increase in the hypothetical compression index. The low rates of creep in the overconsolidated section (i.e. prior to the preconsolidation stress) lead to very little change in the value of the recompression index.

The fact that the projected voids ratios in Figures 6.91 and 6.92 drop to values lower than zero (an impossible state) suggests that the 'real-world' logarithmic decay functions which describe creep are unlikely to be the equal to those observed during oedometer tests with load increment durations of 24 hours. The additional strain accumulated during the extrapolated creep phases would undoubtedly result in reduced rates and magnitudes of consolidation and creep strains in subsequent laboratory loading stages. It is unlikely that rates of creep observed in oedometer tests will continue indefinitely without showing signs of change (decrease). As a result, deviations from the idealised and arbitrary picture presented above could be considerable. Nonetheless, Figures 6.91 and 6.92 serve to illustrate how creep processes have a profound influence on the nature of $e\log_{10}\sigma'$ plots and hence any consequent pre- and retro-dictions of the volumetric behaviour of sediments.

6.5 Dynamic Loading of Mineralogenic Intertidal Sediments

6.5.1 Dynamic loading of surface materials

Graphs of vertical effective stress (kPa, left-hand ordinate axis, logarithmic scale) and vertical displacement (cm, right-hand ordinate axis) against time (in minutes since start of test, plotted on an arithmetic scale) obtained from dynamic loading tests on surface materials are shown in Figures 6.93 (low marsh) and 6.95 (mudflat). It is encouraging that effective stresses applied to the samples in the back-pressured shear box are highly accurate and reproducible even at low stresses, allowing direct comparison of material behaviour between different samples.

The two low marsh samples subjected to dynamic loading scenarios similar to those experienced by sediments at the depositional surface illustrate both similarities and differences in behavioural response. General trends in settlement behaviour are evident on the loading limbs of cyclic load applications. Upon loading, and prior to the maximum peak in the cyclic load, both samples show volumetric reduction. Sample LM-9-BPS-CYC displays a greater degree of settlement (negative vertical displacement) during each cyclic application of effective stress than LM-10-BPS-CYC. However, this is where trends in behaviour diverge; following peak load and in the subsequent reduction in effective stress, sample LM-9-BPS-CYC shows a plastic response, with no swell. The exception to this is following the application of a *c.* 13 kPa load at approximately 2800 minutes, where the sample does swell following the removal of the load. In contrast, LM-10-BPS-CYC shows a greater elasticity of response following peak effective stresses during cyclic loading phases, recovering some, but not all of the previous reduction in volume.

Between cyclic loading phases, and after any swell has occurred, the volumetric behaviour of the sediments is static, as would be expected when effective stress application is zero. This illustrates that any change in volume associated with an effective stress application is rapid and is not ongoing in surface sediments.

During the lower cyclic stress levels applied following the maximum load application of 13 kPa (i.e. loading that takes place after *c.* 3000 minutes), samples show very little response; it seems that only stresses greater than the previous maximum effective stress applied in the back-pressured shear box apparatus influence the vertical displacement.

The unload-reload behaviour associated with this low stress dynamic loading has also been considered in $e\log_{10}\sigma'$ space (Figure 6.94). A greater degree of compressibility is shown by LM-10-BPS-CYC. This is to be expected given the initial voids ratios of the samples (e_i of LM-9-BPS-CYC = 3.96; e_i of LM-10-BPS-CYCS = 4.37 – Table 6.6). The greater plasticity in LM-9-BPS-CYC is again evident, with variations in effective stress resulting in no variation in voids ratio unless the previous maximum effective stress applied in the apparatus is exceeded. LM-10-BPS-CYC again displays greater elasticity in $e\log_{10}\sigma'$ space; it shows more conventional unload-reload hysteresis loops and hence variations in voids ratio in relation to effective stress application. Such variations in the degree of elasticity/plasticity are difficult to explain. The organic contents of the two samples are not considerably different (23.89 % in LM-9-BPS-CYC, 24.05 % in LM-10-BPS-CYC) and are therefore unlikely to cause such obvious variations in material behaviour.

Similar trends are shown by the two mudflat specimens (MF-8-BPS-CYC and MF-9-BPS-CYC) subjected to realistic, depositional surface cyclic loading scenarios (Figure 6.95).

Both samples show settlement upon application of each cyclic load. Following the peak effective stress of each cyclic load being reached, the rate of change in settlement rapidly falls and begins to reverse, entering a swelling phase. The reversal of direction in vertical displacement (i.e. settlement to swell) is more gradual in sample MF-9-BPS-CYC; in MF-8-BPS-CYC, there is no initial change in vertical displacement following the reversal of increasing to decreasing effective stress. Only when effective stress has fully reversed onto its falling limb does MF-8-BPS-CYC begin to swell.

During the intermediate, static stress phases, swelling often continues to occur, but at a much slower rate. In sample MF-8-BPS-CYC, this swelling occurs in a series of occasional swelling 'steps'. Since this phenomenon does not occur in any of the other samples, it is possible that it results from temporary 'stiffness' in the load cell, perhaps due to improper lubrication and increased side friction in the apparatus.

Analysis in $e\log_{10}\sigma'$ space (Figure 6.96) illustrates that considerable variability exists in the compressibility of the samples despite similar initial voids ratios (e_i of MF-8-BPS-CYC = 1.89; e_i of MF-9-BPS-CYCS = 1.75). These variations could be attributable to the variable presence of bioturbation hollows.

The unload-reload behaviour of these samples is visibly elastic prior to the yield stress; both MF-8-BPS-CYC and MF-9-BPS-CYC show conventional unload-reload hysteresis loops in overconsolidated phases of material. If the preconsolidation stress is exceeded, irrecoverable, plastic deformation occurs. Also notable is the considerable variation in voids ratio at a constant stress of 3 kPa; this reflects the continued swelling, either gradual or sudden, that occurs between cyclic loading phases.

6.5.2 Combined overburden and dynamic surcharge loading of intertidal materials

Graphs of vertical effective stress (kPa, left-hand ordinate axis, logarithmic scale) and vertical displacement (cm, right-hand ordinate axis) against time (in minutes since start of test, plotted on an arithmetic scale) obtained from the combined overburden and dynamic surcharge loading tests are shown in Figures 6.97 (low marsh) and 6.106 (mudflat).

A comparison of these figures suggests that the overall behavioural response of both the low marsh and mudflat materials is similar. At lower overburden stresses ($\leq$ 14 kPa), consistent trends in behaviour are evident in both materials. Differences in the magnitudes of settlement between materials reflect both variable material compressibility and differences in the surcharge load applied (13 kPa for low marsh sediments; 25 kPa for mudflat sediments).

Upon initial application of an overburden stress increment, settlement occurs in the same way that it would during a normal (i.e. non-dynamic) incremental loading test. When the surcharge load is applied, this rate of settlement dramatically increases. This is particularly evident in the first (3 kPa) loading stage of both low marsh and mudflat sediments (Figures 6.97 and 6.106). Such a striking increase in rate of settlement reflects the fact that the applied surcharge effective stresses are approaching (low marsh) or exceeding (mudflat) the preconsolidation stress. On the downward limbs of the first effective stress cyclic surcharges in the 3 kPa, 5 kPa and 14 kPa stages, both mudflat and low marsh sediments swell and then, during the intervening static phases, remain at the same volume until the second surcharge load is applied. The overall magnitude of settlement is less than that which occurred upon the first surcharge loading, reflecting plastic deformation during virgin loading. Again, during the decrease in the magnitude of the surcharge load, the materials both undergo swelling followed by phases of no volumetric change.

At vertical effective stresses $\leq$ 14 kPa, changes in overburden load are accompanied by no, or very little, change, in volume. This can be attributable to overconsolidation caused by the surcharge loading in the previous overburden loading stage. Further settlement is only caused by the additional dynamic surcharge loads, which again are interspersed with periods of constant effective stress and no volumetric change. Again, the second surcharge load is accompanied by an elastic response to the loading and subsequent unloading of the sediment.

Similar behaviour is also observed in both materials when subjected to stress of 26 kPa – 99 kPa. In these stages, the increase in effective overburden stress between loading stages begins to have a greater influence on compression, particularly in the low marsh test but also noticeably between 50 kPa and 99 kPa in the mudflat test. The nature of the response to dynamic surcharge loading is also slightly different to that observed at lower stresses of 3 kPa – 14 kPa. Although the surcharge loading results in a phase of increased settlement rate, the post-surcharge swell phase is no longer evident; deformation is increasingly plastic. This may reflect either an increase in the plasticity of the material or may result from the increased overburden pressures preventing swell, or a combination of the two.

The effect of the surcharge load gradually diminishes as the overburden effective stress increases. At stresses equal to and greater than 197 kPa, the surcharge load barely influences the compression curves, with no enhanced settlement upon surcharge loading and no swell upon removal of the surcharge load.

The trends outlined above are more evident when each overburden/surcharge loading stage is considered in plots of vertical displacement against logarithmic and square root time, in the same way as the conventional incremental loading tests were analysed in Section 6.4. Indeed, for comparative purposes, the results of the conventional incremental loading tests are plotted with those of the combined overburden and cyclic loading tests in Figures 6.98 to 6.105 (low marsh) and Figures 6.107 to 6.115 (mudflat), using both logarithmic and square root time. The times at which the surcharge loads are applied and removed are also shown on the figures for clarity. Since the conventional incremental loading tests did not have a 3 kPa loading stage, graphical analysis begins from an effective stress 5 kPa and upwards.

5 kPa and 14 kPa loading stages in both materials (Figures 6.98, 6.99, 6.107 and 6.108) are characterised by initial periods of no settlement or only a small amount of settlement (resulting from initial compression) respectively, contrasting greatly with the primary consolidation observed in the conventional (no surcharge) incremental loading tests. This is due to the overconsolidation caused by the surcharge loading that took place in the previous loading stages. Upon application of the first surcharge load, consolidation-induced settlement takes place rapidly; again, this overconsolidates the material. Swelling occurs as the surcharge load is progressively removed. When the vertical effective stress has returned to the constant overburden value, only a small amount of swell can be seen that is quickly followed by no vertical displacement. This trend stands out sharply against the creep that is occurring in the conventional tests at the same time period. A greater relative magnitude (in relation to the volume lost during the application of the surcharge) of swelling can be seen in the mudflat material (Figures 6.107 and 6.108).

In the 26 kPa loading stage of the mudflat sample (MF-10-BPS-IL+CYC) (Figure 6.109), there is very little settlement prior to the application of the first dynamic surcharge load. When applied, the surcharge results in an acceleration in the rate of settlement, followed by a swell phase and an ultimate rate of change of zero following both surcharge loads. In contrast, the initial (pre-surcharge) compressive response in sample LM-11-BPS-IL+CYC begins to resemble that experienced by the samples loaded with small load increment ratios (LM-2-S-OED-24, LM-7-N-OED-24 and LM-8-N-OED-24) (Figure 6.100). This is due to the smaller surcharge stress applied to the low marsh sediments (13 kPa) in comparison that applied to the mudflat (25 kPa). When the first surcharge loading commences in the 26 kPa loading stage, rapid consolidation takes place in the low marsh samples (Figure 6.100). However, upon removal of the load, swell is now dramatically reduced and the behavioural response is plastic and static at a time when incrementally loaded samples continue to display significant creep settlement. The second surcharge application results in further rapid consolidation; this may be due to excess pore water pressures not having fully dissipated during the first surcharge loading.

In the low marsh sediments at overburden effective stresses of 50 kPa – 197 kPa (Figures 6.101 to 6.103), an increase in similarity can be observed between the primary consolidation phase of sample LM-11-BPS-IL+CYC and all remaining conventional increment loading tests. At overburden stresses of 50 kPa and 99 kPa (Figures 6.101 and 6.102), post-surcharge creep rates remain effectively at zero. However, as the overburden stress increases beyond 197 kPa, post-surcharge creep rates begin to increase until they approach those of the conventional incremental loading tests at a stress of 393 kPa (Figure 6.104). As overburden stress increases, the rate and magnitude of settlement upon application of the surcharge load decreases, and this phenomenon has disappeared at 393 kPa (Figure 6.104).

Trends in the behaviour of the mudflat sample (MF-10-BPS-IL+CYC) at effective stresses greater than 50 kPa are similar to those observed in the low marsh, but not identical. The observed differences in trends are likely to result from the higher value of the applied dynamic surcharge stress. At stresses of 50 kPa to 197 kPa, the pre-surcharge magnitudes and rates of settlement are still considerably less than those observed in the conventional incremental loading tests (Figures 6.110 to 6.112). However, as overburden stress increases (≥ 393 kPa), the pre-surcharge behaviour of MF-10-BPS-IL+CYC begins to resemble that of the conventional incremental loading tests and is largely identical at 1570 kPa (Figure 6.115). A dramatic increase in the rate and magnitude of primary consolidation during the first surcharge loading continues to be evident at stresses of 50 kPa to 785 kPa (Figures 6.110 to 6.114), although this effect gradually decreases and is barely noticeable at an overburden stress of 1570 kPa (Figure 6.115). A slight swell after the maximum surcharge stresses can still be observed at 50 kPa, followed by phases of zero vertical displacement. From 99 kPa onwards, these effects also decrease in prominence, until there is no post-surcharge swell and little variation in the creep rate compared to the conventional incremental loading samples (Figures 6.111 to 6.115).

The observed variation in behaviour can be explained in terms of overconsolidation and the surcharge ratio, which is the ratio of the maximum surcharge stress to the existing 'background' overburden stress (σ_s/σ). The surcharge ratios for each loading stage are shown in Table 6.9 for the low marsh and 6.10 for the mudflat.

The effect of the temporary surcharge load is to overconsolidate a material. At low stresses analogous to the conditions experienced in near-surface sediments, the surcharge ratio is high, with values as high as 8.33 in the 3 kPa stage in the mudflat (Table 6.9 and 6.10). Consequently, the degree of overconsolidation effected in these materials is also high. During surcharge loading, the volume of the material is reduced to a denser state. Upon removal of the surcharge load, both materials show plastic behaviour, remaining close to this dense state; although they do undergo some elastic swell, the virgin compression experienced is not fully recoverable. Since the materials are now overconsolidated, the increased density of each material has caused increased structural stability; following removal of the surcharge load the samples are already in volumetric and structural equilibrium with the ambient stress conditions. As a result, ongoing creep settlement does not occur. This suggests that creep is more likely to occur in normally consolidated sediments with open fabrics.

As overburden stresses increase, surcharge ratios decrease to values less than 1. The additional dynamic (surcharge) effective stress becomes less and less influential on material behaviour. Since the relative stress variation between the constant and dynamic stress states is less pronounced at higher stresses, indicated by the decreasing surcharge ratios in Tables 6.9 and 6.10, the degree of overconsolidation becomes less. The use of a logarithmic scale for effective stress readings in Figures 6.97 and 6.106 emphasises the diminishing effect of the dynamic surcharge relative to existing overburden stresses. Variations in structure before and after the application of the load diminish in magnitude. Creep rates at higher stresses are therefore expectedly similar pre- and post-surcharge.

Table 6.9 Effective overburden stresses, surcharge stresses and associated surcharge ratios applicable to the combined overburden and cyclic loading test undertaken on sample LM-11-BPS-IL+CYC.

Overburden effective stress, σ (kPa)	Surcharge stress, σ_s	Surcharge ratio, σ_s/σ
3	13	4.33
5	13	2.6
14	13	0.92
26	13	0.50
50	13	0.26
99	13	0.13
197	13	0.07
393	13	0.03
785	13	0.02

Table 6.10 Effective overburden stress, surcharge stresses and associated surcharge ratios applicable to the combined overburden and cyclic loading tests undertaken on sample MF-10-BPS-IL+CYC.

Overburden effective stress, σ (kPa)	Surcharge stress, σ_s	Surcharge ratio, σ_s/σ
3	25	8.33
5	25	5.00
14	25	1.79
26	25	0.96
50	25	0.50
99	25	0.25
197	25	0.13
393	25	0.06
785	25	0.03
1570	25	0.02

In essence, the higher the surcharge ratio, the greater the structural stability upon removal of this dynamic load. This consequently increases the overconsolidation ratio and lowers the creep rate. As the surcharge ratio decreases, the application of the surcharge stress has much less of an influence because the stress conditions are not dramatically different from those experienced in conventional incremental loading tests. Furthermore, at higher effective overburden stresses, permeability of the materials is lower, reducing the probability that the increase surcharge total stress is transferred to an effective stress in the time-frame involved.

The degree to which dynamic surcharge loading can reduce the rate of creep, as described by C_α, is displayed in Figures 6.116 and 6.117. These figures compare the values of C_α observed in the incremental loading tests (as displayed previously in Figures 6.89 and 6.90) with those calculated from the post-surcharge phases in samples LM-11-BPS-IL+CYC and MF-10-BPS-IL+CYC. The reduction in creep in these samples is striking, with values of C_α remaining at 0 until 100 kPa in the low marsh and 393 kPa in the mudflat. These stresses coincide with surcharge ratios of 0.13 (low marsh) and 0.06 (mudflat) (Tables 6.9 and 6.10). Whether or not these reflect critical values of the surcharge ratio at which dynamic surcharge loading diminishes in importance requires further investigation. Following the re-occurrence of creep at the respective values of overburden stress and surcharge load, values of C_α being to rapidly increase. In the mudflat specimen (MF-10-BPS-IL+CYC), this increase continues until it is effectively equal to the rate observed in virgin compression in the oedometer. This reflects the insignificance of the surcharge load at an effective stress of 1570 kPa. In the low marsh sample (LM-11-BPS-IL+CYC), post-surcharge values of C_α never reach those obtained from conventional incremental loading tests, but they do show a similar trend – i.e. an increase in C_α followed by a decrease towards higher stresses. The value of C_α obtained from sample LM-11-BPS-IL+CYC is typically approximately a third to a half that displayed by the incremental loading tests at effective stresses greater than 197 kPa. This confirms the persistent, yet waning, influence of the dynamic surcharge loading on the low marsh material.

In Figures 6.116 and 6.117, the values of C_α displayed by the oedometer reloading data series plot lower than those of the post-surcharge values obtained from the back-pressured shear box. This reflects the degree of overconsolidation in each sample and is to be expected given the nature of the tests. The surcharge ratio in the overburden/dynamic loading tests is low at these stresses (> 100 kPa), leading to a lesser degree of overconsolidation. The maximum total stress is only momentarily applied, preventing full dissipation of excess pore water pressures in this small timeframe in decreasingly permeable sediments. Full transferral to effective stress is unlikely. In contrast, the overconsolidation ratio in the oedometer reloading stages is considerably higher, resulting in a denser fabric that is in structural equilibrium at reduced stresses, leading to lower creep rates.

6.6 TIME-DEPENDENT DEFORMATION OF INTERTIDAL MATERIALS

The analysis of time-vertical displacement behaviour in Section 6.4 has revealed that the compression behaviour of the sampled low marsh and mudflat sediments displays considerable time-dependency. In particular, rates of creep in these sediments are sufficiently high to cause considerable contribution to total settlement (Section 6.4.4, Figures 6.77 to 6.88) and to variations in the form of $e\log_{10}\sigma'$ plots (Section 6.4.4, Figures 6.91 and 6.92). In view of the importance of creep processes to overall compression in conventional incremental loading tests, it seems that a time variable has to be introduced into $e\log_{10}\sigma'$ analysis. However, as the following discussion highlights, this is perhaps not only difficult, if not impossible, to do in intertidal sediments but, as suggested by the findings from the dynamic surcharge tests in the back-pressured shear box, may not necessarily be required.

6.6.1 Estimating rates and magnitudes of soil compression: a civil engineering perspective

An orthodox engineering analysis of laboratory data to assess and predict field behaviour following construction on clay soils would typically involve basic oedometer testing in which load increments closely reflect the loading stages associated with phases of building and constructions (Powrie, 2004). From each load increment, for which load increment ratios exceed 1, values of c_v would be obtained using the Casagrande or Taylor methods. According to Terzaghi's theory, the time necessary to obtain a given degree of primary consolidation can be evaluated by:

$$t = \frac{T_v H^i}{c_v} \tag{6.1}$$

where:

T_v = the theoretical time factor
t = the time necessary to obtain that degree of consolidation
H = the length of the longest drainage path
c_v = the coefficient of consolidation.

The exponent i equals 2 in conventional Terzaghi theory. Using this formula, the time at the end of the primary consolidation in the field can then be evaluated from the laboratory observations (Das, 1998). Secondly, values of C_α can be calculated, scaled to the height of a stratigraphic layer and extrapolated to determine the magnitude of creep settlement (Berry and Poskitt, 1972), if indeed it makes a significant contribution to settlement.

6.6.2 Estimating rates and magnitudes of soil compression: an intertidal geomorphological perspective

At effective stresses similar to those observed in coastal stratigraphic sequences, it is difficult to obtained values of c_v due to the significant deviations of the observed time-vertical displacement curves from the conventional curves predicted by Terzaghi's consolidation equation. Such deviations, such as the continuously curved nature of square-root time-settlement plots and an absence of a reversal in curvature in logarithmic time-settlement plots, arise from the overconsolidation of the materials, small (< 1) load increment ratios and the simultaneous operation of consolidation and creep processes. With the lack of alternative curve-fitting techniques and without monitoring of excess pore pressure reductions, including a time element of primary consolidation into autocompaction models is prohibited.

The validity and predictive capacity of Terzaghi's consolidation theory is also open to question. Firstly, it is has frequently been noted that the two existing curve-fitting methods typically give different results (Das, 1998; Lambe and Whitman, 1979). Secondly, values of c_v for a given load increment often vary considerably for lithologically homogeneous samples due to inherent variability in material properties (Lambe and Whitman, 1979). These factors make it difficult to select a reliable value of c_v for use in a field application. As a consequence, Lambe and Whitman (1979) state that analysis using Terzaghi's linear theory of consolidation can at best provide only an order-of-magnitude estimate of the time required for consolidation to cease.

Comparisons between field data and laboratory tests have also highlighted a further general problem of the use of Terzaghi's consolidation theory. Such experimental comparisons have questioned the accuracy of the scaling exponent i in Equation 6.1. Conventional wisdom assumes that $i = 2$. However, in soils with organic components, it has been illustrated that $1 \leq i \leq 2$. Samson and La Rochelle (1972) present values of between 1.6 and 2.0 in peat deposits. Lefebvre *et al.* (1984) report values of 1.1 to 1.5 in a fibrous peat. Such values mean that drainage is faster than predicted by the one-dimensional theory of consolidation and the required laboratory values. This is attributable firstly to macrostructural factors that are not sufficiently represented in small, thin oedometer samples; and secondly to the effects of lateral drainage in the field that are not reproduced in laboratory tests with only vertical drainage (Lambe and Whitman, 1979). In the intertidal materials considered, values of *i* are unknown and may be lower than 2 even in the mudflat material, where bioturbation structures may increase secondary permeability and increase the rate of primary consolidation and settlement.

Differences between laboratory and field creep behaviour have also been noted. Lefebvre *et al.* (1984) showed that the secondary compression observed in the field is about twice that was observed in the laboratory. Secondly, it is not uncommon for values of the coefficient of secondary compression to increase with time, particularly in organic soils, as observed by Barden and Berry (1968, in Berry, 1983). Quite clearly, estimates of creep based on short-term laboratory compression experiments may bear no resemblance to *in situ* rates and can lead to erroneous predictions of settlement.

Terzaghi's consolidation theory was developed for analysis of the time-settlement behaviour a single argillaceous stratum that is over- and under-lain by free-draining sand layers. Again, intertidal stratigraphies deviate from this idealisation as they generally consist of a number of layers, each comprising different lithologies and geotechnical properties. Since decompaction approaches typically spilt even a single stratigraphic unit into thinner layers increased uniformity of geotechnical properties, analysis using Terzaghi theory is further complicated due to the interaction of the different strata in terms of permeability, pressure gradients and compressibility. Lambe and Whitman (1979) discuss how even a relatively simple two strata system results in quite complicated analysis in which the relatively more compressible, rapidly draining layer alters the pore pressure conditions, and hence consolidation behaviour, in the over- or under-lying stratum. Such interactions are not explicitly treated by Terzaghi's theory and analysis of such a system, let alone a multi-strata, highly dynamic intertidal stratigraphy, would require additional more complex analytical modelling in a finite or discrete element package, for example.

On the balance of evidence the application of Terzaghi's consolidation theory would be misleading, even in the mudflat sediments where values of c_v can be estimated from time-settlement curves.

6.6.3 Low rates and magnitudes of overburden sedimentation

Despite a great deal being learned in this study about the time-dependent compression behaviour of the sampled low marsh and mudflat materials, further difficulties arise in the application of this knowledge due to differences in the magnitude and rate of effective stress applications in the field and laboratory.

Overburden stress is applied to stratigraphic sequences in the upper intertidal zone at low rates and in very low stress increments. This slow, continuous and time-variable application of total stress contrasts greatly with stepped, incremental loading scenarios employed in the laboratory. *In situ* sedimentation will result in very small load increment ratios and only small increases in excess pore water pressure. It has been shown experimentally both in this study and elsewhere (e.g. Leonards and Girault, 1961) that load increment ratios less than 1 result in time-settlement curves that substantially differ from orthodox Terzaghi curves (Figure 6.74). In such situations, values of c_v cannot be calculated and so it becomes even more difficult to apply Terzaghi's consolidation theory to these intertidal sediments, even in the mudflat sediments that display conventional time-settlement behaviour when load increment ratios are greater than or equal to 1. Therefore, loading samples in the laboratory at loading stages and stresses that equal those experienced in the field would not only be extremely time-consuming, but would also yield data that is of no practical use if c_v cannot be estimated. Furthermore, when considered in the light of *in situ* rates of overburden load increase, analysis of creep behaviour becomes difficult to predict on the basis of conventional oedometer test data, since values of C_α are only strictly applicable to the load increments from which they were estimated.

Without realistic estimates of the parameters that describe time-dependency (c_v and C_α), it would be unwise to attempt to include a time factor in analysis of settlement rates and magnitude. The situation is even more complicated in the low marsh, where primary consolidation and creep occur simultaneously, as indicated by continuously-curved square root time-settlement plots with no obvious change in gradient. Needless to say, however, simply stating that time-dependency is difficult to include in models of the autocompaction behaviour of mineralogenic intertidal sediments on the basis of conventional incremental loading test data is insufficient and does not resolve the issue. It is perhaps fortunate, therefore, that the dynamic loading of intertidal sediments may offer a solution to this issue.

6.6.4 Dynamic loading and overconsolidation

The dynamic surcharge loading of intertidal materials was shown to overconsolidate the materials following removal of these surcharge loads. This overconsolidation increased the structural stability of the samples at stresses lower than the maximum previous effective stress experienced by the samples and hence reduced the creep rate to zero at effective stress levels generally experienced in intertidal stratigraphies.

This post-surcharge reduction in creep rate and magnitude is not a new finding. Hobbs (1986), discussed how organic soils typically undergo a similar swelling ('rebound') process to that observed in the low marsh and mudflat samples in this study following removal of the dynamic surcharge. Following application of the surcharge at point A (Figure 6.118) and subsequent removal at point B, Hobbs (1986) observed an immediate swell (line BC, Figure 6.118) followed by a long-term secondary swell (line CD, Figure 6.118). In the samples loaded dynamically in this study, this long-term swelling phase rapidly flattens to a condition of zero vertical displacement due to the existing overburden preventing swell. After the long-term swelling phase, Hobbs (1986) notes a further change in behaviour; the swell curve eventually rejoins the secondary compression curve that theoretically would have taken place had the surcharge load not been applied. This occurs at the intersection of the long-term swelling curve and the theoretical secondary compression line (Point D on Figure 6.118).

The re-assertion of secondary compression is not observed in the cyclic loading tests in the back-pressured shear box due to the subsequent application of dynamic loads and step increases in load increments. Indeed, given the constant variation in effective stress observed in Chapter 5, it is highly unlikely that a continuation of the theoretical creep settlement rate will occur in the dynamic intertidal environment. The dynamic loading tests undertaken on samples LM-9-BPS-CYC, LM-10-BPS-CYC, MF-8-BPS-CYC and MF-9-BPS-CYC showed that samples behaved either plastically (LM-9-BPS-CYC and LM-10-BPS-CYC) or elastically (MF-8-BPS-CYC and MF-9-BPS-CYC); no creep was observed even between stages of effective stress application. Indeed, it is the combination of dynamic surcharge loading, the consequent post-surcharge overconsolidation and reduced rate and magnitude of creep, and continuous cyclic loading of materials that provides a basic mechanism for the prevention of creep in intertidal materials. It is thus not unreasonable to suggest that settlement of intertidal materials is dictated largely by consolidation processes. Particularly in the mudflat sediments, in which creep processes have been shown to occur only following dissipation of excess pore water pressures, the dynamism of effective stress variation means that creep processes are never given the opportunity to operate. For the first time, a scientific, empirical rationale exists for the exclusion of creep processes in models of autocompaction in mineralogenic intertidal materials.

6.7 SUMMARY: IMPLICATIONS FOR AUTOCOMPACTION BEHAVIOUR

This chapter has examined the geotechnical (compression) properties of mineralogenic intertidal sediments in considerable detail. The novel adaptation of conventional soil mechanics techniques to better reflect the dynamics of effective stress variation in the upper intertidal zone has revealed some unique and significant insights into the operation of compression processes in these sediments. The implications of these data for the development of accurate models of autocompaction in mineralogenic intertidal sediments can be considered with reference to four of the research hypotheses defined in Section 3.10:

1. Mineralogenic intertidal sediments show no variability in structure and/or compression behaviour.

Variability in compression behaviour was observed both within and between materials. The low marsh specimens displayed a range of values of the recompression (C_r, 0.08 – 0.52) and compression (C_c, 1.41 – 2.27) indices. The mudflat samples showed values of C_r between 0.02 and 0.05; values of C_c ranged from 0.50 to 0.73.

It was noted in Chapter 5 that variations in structure (voids ratio) occur in lithologically homogenous sediments obtained from constant altitudes. The results presented in Section 6.2 have demonstrated that these initial structural variations directly translate into variations in compression behaviour. Strong, positive and statistically significant correlations were found between the initial voids ratio, e_i, and both C_r and C_c in both the low marsh and mudflat materials. This suggests that sediments with more 'open' structures at the depositional surface are more prone to compression both pre- and post-yield.

Variations in behaviour and structure were found to vary at stresses less and greater than the yield stress. Loss of initial structure *via* a destructuration process following yield increases structural and behavioural homogeneity, leading to a convergence of compression lines at high stresses and an increase in similarity of values of the compression index.

Hypothesis 1 is therefore rejected. The compression behaviour of a mineralogenic material cannot adequately be described by single values for the initial voids ratio and the rheological parameters, C_r and C_c. It is therefore necessary to develop a statistical model to describe inherent variations in structure and behaviour that occur in lithologically uniform materials. The destructuration that takes place following yield leads to an increase in structural and behavioural homogeneity; in order to satisfactorily describe this process, the statistical error term should be different at stresses greater and less than the yield stress.

3. Near-surface mineralogenic intertidal sediments are normally consolidated.

The mineralogenic intertidal materials tested in this chapter were shown to be overconsolidated. Values of the preconsolidation stress, σ'_c, in the low marsh samples ranged from 20 to 27 kPa; these values are greater than the 9 – 13 kPa predicted on the basis of effective stress variations caused by tidal loading and groundwater variation in Chapter 5. In the mudflat specimens, observed values of σ'_c ranged from 8 -14 kPa; these values contrasted with the 25 kPa predicted from tidal loading. It was concluded that hydrological variables play a less important role in controlling the degree of overconsolidation in mineralogenic intertidal sediments than was originally considered. The observed preconsolidation stresses were deemed to result from desiccation during subaerial exposure and, in the low marsh samples, plant root suction stresses.

It was noted, however, that the transition from the low gradient, overconsolidated recompression line to the steeper gradient normal compression line begins to occur at a stress lower than the preconsolidation stress, which is approximated using a graphical construction. The earlier change from the recompression to the normal compression line is particularly pronounced in the dynamic loading tests on surface materials (Section 6.5.1).

On the basis of this data, hypothesis 3 is rejected. A backwards projection of the normal compression line does not accurately describe the low stress compression behaviour of the intertidal sediments studied. Consequently, a 'conditional' regression model is required in which values of the gradient and intercept are conditional on whether a value of effective stress is less or greater than the initial yield (not preconsolidation) stress.

4. Terzaghi's compression law and consolidation theory are applicable to mineralogenic intertidal sediments.

Variations in structure and compression behaviour were observed in both the low marsh and mudflat materials. These mineralogenic materials are also overconsolidated. Hence, Terzaghi's compression law, involving a backwards projection of the normal compression line to 1 kPa and the use of single values of initial voids ratios and compression indices, is a deficient rheological model.

Terzaghi's consolidation theory cannot be applied to low marsh materials due to the simultaneous operation of consolidation and creep processes. This leads to deviations from conventional Terzaghi plots of time-settlement curves, prohibiting the successful use of established curve-fitting procedures. In addition, low values of the load increment ratio lead to similar deviations in graphical form in both the low marsh and mudflat materials. Such low load increment ratios typify the low magnitudes of stress resulting from low annual sedimentation rates in intertidal environments. If a time-dependent element was to be used in autocompaction models, values of c_v are needed. Clearly, this is not possible and Terzaghi's consolidation theory does not adequately reflect the geomorphological problem of interest.

Hypothesis 4 is therefore rejected. Although modelling can still occur within the $e\log_{10}\sigma'$ framework, considerable modification to Terzaghi's compression law is required to account for the structural and behavioural variability and overconsolidation (as highlighted above).

5. Primary consolidation is the principal settlement process in mineralogenic intertidal sediments; creep is unimportant.

Whilst creep makes a significant contribution to total settlement in conventional incremental loading compression tests, dynamic surcharge loading overconsolidates the sediments and significantly reduces the rate of creep. Creep is prevented from recurring following removal of the surcharge load due to constant variations in effective stress in the tidally-driven dynamic intertidal environment. This was particularly evident in the dynamic loading of surface materials in which the cyclic loading result in a clear elastic response and creep deformation at constant effective stress was not observed.

It is therefore not possible to reject hypothesis 5. In combination with the inapplicability of Terzaghi's consolidation equation to the intertidal environment and intertidal materials, the prevention of operation of time-dependent creep processes suggests that a time variable is not required in models autocompaction of mineralogenic intertidal sediment.

6.8 Conclusions

Previous studies undertaken into autocompaction in different stress and diagenetic environments (Audet, 1995; Been and Sills, 1981; Nygard *et al.*, 2004) displayed that modifications to Terzaghi's compression law and consolidation equation are required before their predictive capacity can be considered accurate and representative. Terzaghi's basic models have also previously been applied to intertidal sediments and environments (Massey *et al.*, 2006; Paul and Barras, 1998). These applications were made without adaptation of the models, yet no firm empirical rationale existed for their unmodified use. The data presented in this chapter have, for the first time, provided significant insights into the compression behaviour of mineralogenic intertidal sediments.

Many of the fundamental assumptions associated with Terzaghi's compression law and consolidation equation are challenged by the dynamic intertidal environment and the materials that form there. The findings of this chapter have clearly demonstrated both the dangers of applying the Terzaghi's models to intertidal areas without sufficient critical appraisal of their applicability and that detailed geotechnical testing is a critical prerequisite to the development of phenomenologically correct models of autocompaction.

On the basis of the findings of the geotechnical testing program, empirically-informed decisions can now be made regarding the modelling approach. Before this can be done, there is also a need to consider the influence of influences on compression behaviour that result from (bio-)chemical diagenetic processes. The results of the investigation into near-surface diagenetic phenomena are presented in Chapter 7.

Chapter 7: Diagenetic Processes in the Overconsolidated Vadose Zone

This chapter presents results of an investigation into diagenetic processes in shallow stratigraphies and their effect on *in situ* voids ratio and compression behaviour. Diagenetic processes are known to operate in the vadose zone (Section 3.9.10) and it is necessary to identify such processes and investigate their influence. Near-surface sediments are likely to be overconsolidated and thus it is possible that the influence of effective stress is secondary to that of other diagenetic processes. Understanding the relative importance of these processes is therefore critical to the modelling approach.

7.1 Diagenetic processes in shallow mineralogenic intertidal stratigraphies

The first step in determining the influence of diagenetic processes on *in situ* voids ratios involved their identification in shallow, vadose zone stratigraphies. This was achieved through examination of a series of overlapping cores collected from the beneath the low marsh and mudflat (sampling altitudes 2.26 m OD and 1.06 m OD) according to the methods described in Section 4.2.2.

From the low marsh site, cores LMX-1 and LMX-3 extend from the surface to a depth of 0.17 m. Core LMX-2 samples depths of 0.16 - 0.33 m and core LMX-4 covers the vertical range of 0.21 – 0.39 m (the maximum depth sampled). The stratigraphy of the sampled sedimentary profile is illustrated and described in Figure 7.1. The profile consists of 5 primary stratigraphic units. Working from the base of the core upwards, the lowest unit (L1, 0.39 – 0.15 m) is a dark brown organic silt. The darker, mottled colour may reflect either the presence of insoluble iron-sulphides and reducing conditions (controlled by the depth of the water table beneath the low marsh surface; Cundy and Croudace, 1995b), humification of organic matter or a combination of the two. Differentiating between humification and sulphide enrichment is difficult by visual observation alone, since both phenomena tend to result in a darkening of the sediment. The overlying unit, L2 (0.15 – 0.13 m) has a similar lithological make-up but is mid-brown in colour, suggesting a lesser degree of humification and/or the presence of iron oxyhydroxides (Zwolsman *et al.*, 1993). There is a return to the darker, humified state in unit L3 (0.13 – 0.12 m), before the sediment once again displays a lighter brown colour and an apparently lesser degree of humification within unit L4 (0.12 – 0.07 m). At 0.07 m, there is a fairly diffuse boundary into a more organic unit (L5 a); this is the contemporary low marsh material (an organic silt) that extends upwards to a depth of 0.005 m. Extending downwards into unit L5a is L5b - a loosely bound organic silt that is matted by the surface vegetation. This unit was unsuitable for geotechnical testing and was previously disregarded in the geotechnical testing section (Chapter 6). All sediments above 0.12 m display red, brown and sometimes grey mottling (i.e. no distinct redox zonation), indicative of a fluctuating water table (Zwolsman *et al.*, 1993).

The mudflat cores obtained (five in total) extend to a greater depth (*c.* 0.48 m) beneath the contemporary mudflat sampling altitude. Cores MFX-1 and MFX-4 extend from the surface to a depth of 0.18 m. Cores MFX-2 and MFX-3 cover the range of depths from 0.12 – 0.31 m. Core MFX-5 extends from 0.31 – 0.48 m. The stratigraphy is displayed in Figure 7.2 and is less complex that that of the low marsh cores, with two main units present. The lowest stratum (unit M1) is a dark black silt with some sand and clay. The black colour of the material indicates an highly reduced sediment and the presence of iron-sulphides. It is probable that this section of the stratigraphic column remains fully saturated by capillary action. Desiccation in this sediment is also likely to be rare given the high (semidiurnal) frequency of flooding by tidal waters and the reduced duration of subaerial exposure (Section 5.5). This stratum also has occasional, localised sections of randomly oriented, and hence detrital, organic material. The overlying unit (M2, 0.03 – 0.00 m) is of the same lithology. However, the sediment is light brown (oxidised) and is characterised by the presence of bioturbation burrows. It is unlikely to be coincidence that the depth of the oxidised zone and the depth to which these burrows occur is the same; indeed, faunal bioturbation structures are likely to be responsible for the aeration and oxidation of this zone.

The visual description of the sample cores suggests the operation of two non-mechanical diagenetic processes in the shallow intertidal stratigraphies. Firstly, redox remobilisation and zonation is evident in both cores, most obviously in the mudflat samples. Secondly, there is evidence that humification has differentially operated at specific depths in the low marsh cores. This process is absent in the mineralogenic mudflat cores.

7.2 Quantification of Variables

The next stage in determining the relative influence of different variables (effective stress, lithology, geochemical variations and degree of humification) on *in situ* voids ratio involves their quantitative measurement. For the majority of these variables, such quantification is straightforward. The relevant methods are well-established and outlined in Chapter 4. Humification, however, is difficult to quantify for the purposes of this investigation.

Analysis of peat humification has its origins in studies of ombrotrophic (i.e. deriving moisture and nutrients directly from rain water) peat bogs, which are used as potential archives of palaeoclimatic and palaeoenvironmental data (Caseldine *et al.*, 2000). This technique rests upon the assumption that plant decay (humification) is primarily determined by surface wetness and temperature at the time of peat deposition. The darkness and colour of the humic acids produced by the decomposition of organic matter are indicative of the degree of humification and hence decomposition and climate at the time of formation (Blackford and Chambers, 1993). Analyses have evolved from 'hands-on', and hence subjective, field determination methods. For example, the von Post scale of peat decomposition categorises different degrees of humification according to a ten point scale on the basis of soil texture and colour and the nature of the water squeezed from the peat. More recently, analysis typically involves examination of humic substances that are chemically extracted from the soil (Anderson, 1998; Borgmark, 2005; Borgmark and Schoning, 2006; Langdon and Barber, 2001). The most widely used of these more quantitative chemical methods is that proposed by Blackford and Chambers (1993). This involves the use of 8 % sodium hydroxide (NaOH) as the extractant, into which dried peat samples (typically 0.1 - 0.5 g) are placed and heated for an hour. Samples are then filtered and diluted to the required concentration, which depends on the mass of sample used in analysis. Measurement is then undertaken using UV/VIS spectrometry at a wavelength of 540 nm and expressed as percentage light transmission (or absorbance). Some researchers, however, convert the 'raw' transmission measurements to humification by relating them to humic acid standards (Caseldine *et al.*, 2000; Chambers *et al.*, 1997).

Humification analysis was attempted on the low marsh core, sampling at 0.01 m depth intervals and using the established Blackford and Chambers (1993) technique. Results were, however, inconclusive; downcore values of percentage light transmission remained largely constant at *c.* 92 – 93 %. This suggests that either there is no variation in humification throughout the core (which is in contrast to conclusions obtained from visual observation of the core), or that the technique is not suited to predominantly mineralogenic intertidal sediments. This may be because the Blackford and Chambers (1993) technique was developed for freshwater peats that are composed principally of decayed upland mosses, sedges and shrubs. The chemical and spectral properties of the humic acids of these peats are likely to differ from those obtained from intertidal halophytic vascular flora and so the choice of chemical extractant and wavelength is critical (Caseldine *et al.*, 2000). Indeed, Garcia *et al.* (1993) for example, illustrated that NaOH preferentially released high molecular weight humic fractions in *Sphagnum*-based but not *Carex*-based peats. Since even the descriptive von Post method was developed for 'pure' peats, assigning reliable, quantitative values of humification is extremely difficult in the low marsh sediments. In the absence of a suitable method, humification is not considered as a predictor variable. However, some implications of humification are readdressed later (Section 8.2.1).

For the remaining variables, their downcore variation is now discussed, starting with the dependent variable (voids ratio). Depth profiles of the independent (predictor) variables are described, along with any relationships that they have with voids ratio, as determined by visual observations and *via* Pearson's correlation tests (Tables 7.1 and 7.2). The foraminiferal biostratigraphy of the cores is also presented to determine the environment of formation of the deeper sections of the cores (saltmarsh or mudflat environments).

In terms of geochemistry, seven diagenetically sensitive substances have been selected on the basis of previous work into the redox remobilisation and zonation in near-surface intertidal sediments by Cundy and Croudace (1995b), Cundy and Croudace (1996), Thomson *et al.* (2002) and Cundy (*pers. comm.*). These are: silica (SiO_2), iron (III) oxide (Fe_2O_3), manganese monoxide (MnO), calcium oxide (CaO, a proxy for $CaCO_3$ – Thomson *et al.*, 2002), phosphorus pentoxide (P_2O_5), sulphur (S) and iodine (I). Geochemical analysis was undertaken at a sampling resolution of 0.02 m in both the low marsh and mudflat cores.

7.2.1 Voids ratio profile of the low marsh core

Downcore voids ratio profiles for the four low marsh cores, as derived from x-ray core scanning, are displayed in Figure 7.3. Figure 7.3 (a) displays the mean voids ratio value as well as a root squared error (±), which incorporates both the standard error of the mean and the systematic instrumental error (Section 4.3.4). It is clear that the associated errors are relatively small, other than the occasional larger error (at *c.* 0.3 m, for example) which is likely to represent a random, unavoidable surge in the electrical mains supply. Henceforth and for clarity, only the mean value of the upward and downward scans is considered (Figure 7.3 (b)).

Following x-ray scanning, cores LMX-1 and LMX-2 were extruded from the sampling tubes and sliced open to allow sampling for litho-, bio- and chemo-stratigraphic analysis. The remaining cores (LMX-3 and LMX-4) were carefully extruded and prepared for subsequent oedometer testing.

Cores LMX-1 and LMX-2 were also photographed. Figure 7.4 displays these photographs, overlain by the relevant voids ratio profiles. The visual stratigraphy is also displayed. It can be seen that each lithological unit has its own voids ratio signature. This confirms that voids ratios reflect lithology, indicating the reliability of the technique and its potential as a non-destructive method for sediment characterisation. However, there is considerable within-unit variation in voids ratio, presumably reflecting smaller-scale differences in depositional conditions, moisture content and stress history, as observed in the geotechnical samples at the depositional surface (Chapter 5).

In the thicker and deepest humified stratigraphic unit (L1; 0.39 – 0.15 m), voids ratios display an upward rising trend. At 0.39 m, values of *e* begin at *c.* 2.5 and rise to *c.* 4 at 0.15 m. Occasional fluctuations in voids ratio are evident throughout unit L1 that cannot be visually attributed to any variations in lithology. Voids ratios fall to *c.* 2.5 in unit L2 (clay-rich silt with organic matter, 0.15 – 0.13 m) before rising again to *c.* 4 in unit L3 (humified organic silt; 0.13 – 0.12 m). In the overlying unit (L4; clay-rich silt with some organic matter, 0.12 – 0.07 m) voids ratios decrease and fluctuate between 2 and 3.5.

At the boundary to the uppermost unit (L5a; organic silt, *c.* 0.07 – 0.00 m), values of e rise to fluctuate between 3.5 and 6.5. This rise is more pronounced in core LMX-1, perhaps indicating a sharper stratigraphic boundary. Such rapid lateral and stratigraphic variations in voids ratio corroborate the previous suggestion (Section 6.3.5) that rapid spatial variations in initial voids ratio at the depositional surface translate into lateral and stratigraphic voids ratio variations. In core LMX-3, there is an obvious decreasing trend in voids ratio, from *c.* 5.25 at 0.07 m to *c.* 3.5 at 0.025 m. This trend then reverses, again increasing to a value of *c.* 5.25 at the low marsh surface. A similar, although less pronounced, trend is evident within unit L5a in core LMX-1. Values of *e* are generally higher in this core sample, and reach their maximum (*c.* 6.3) at *c.* 0.04 m.

7.2.2 Effective stress profile of the low marsh core

The effective stress profile of the low marsh core is displayed in Figure 7.5. In general, effective stress increases linearly with depth. However, variations in this trend occur and these can be attributed to lithological, and hence density, variations. The most obvious of these variations in gradient occurs at a depth of *c.* 0.07 m, which corresponds with the transition from unit L5a to L4.

7.2.3 Biostratigraphy of the low marsh core

The low marsh cores were sampled for foraminifera at a minimum resolution of 0.02 m. Foraminiferal assemblages (as a percentage of the total counted at each sampling depth) and total counts are displayed in Figure 7.6 in relation to depth, voids ratio profiles and visual stratigraphy. Cluster analysis was also performed on the foraminiferal assemblages using CONISS (Section 5.1.4; Grimm, 1987; 1993). Stratigraphic constraint was used during analysis, meaning assigned clusters must consist of stratigraphically contiguous samples.

Foraminifera counts are generally greater than 100, allowing reliable palaeoenvironmental reconstruction. However, between 0.08 and 0.14 m, counts do drop below 100. It is general practice to exclude such counts from palaeoenvironmental analysis; however, the monospecific assemblage (*Jadammina macrescens*) at these depths increases statistical confidence.

Cluster analysis split the samples into four foraminiferal zones. Zone LF1 (0.39 – 0.32 m) consists almost entirely of *Jadammina macrescens* (80 – 100%) with occasional *Trochammina inflata* (maximum of 20 %) and *Haplophragmoides* spp. (< 10 %).

In zone LF2 (0.30 – 0.16 m), *Miliammina fusca* becomes increasingly dominant between 0.30 and 0.16 m depth, constituting 20 – 40 % of the total foraminifera counted; this is accompanied by a commensurate decrease in the relative abundance of *Jadammina macrescens*. *Trochammina inflata* (< 10 %) and *Haplophragmoides* spp. (< 10 %) are also found within this zone.

Zone LF3 (0.14 – 0.04 m) is characterised by low species diversity (i.e. low number of species), again with *Jadammina macrescens* being dominant. The assemblage is monospecific between 0.12 and 0.10 m depth, and very nearly so at 0.08 m depth where *Miliammina fusca* constitutes < 5 % of total foraminifera.

In zone LF4 (0.07 – 0.00 m), species diversity increases and the dominance of *Jadammina macrescens* accordingly decreases. Particularly in the uppermost 0.04 m, *Trochammina inflata* becomes a more significant species, and even becoming dominant at 0.02 m.

All foraminifera present are agglutinated species, indicating that the entire core section accumulated within a saltmarsh environment (Horton, 1997; Horton, 1999; Horton and Edwards, 2000). Furthermore, the low species diversity is also typical of saltmarsh foraminiferal assemblages (in comparison with calcareous mudflat assemblages).

As well as being related to the visual stratigraphic units, the foraminiferal assemblages are also related to the voids ratio profile. Most noticeably, the highest (near-surface) voids ratios at the top of the core correspond with assemblages dominated by *Trochammina inflata*. Additionally, the lowest voids ratios (0.07 – 0.12 m depth) correspond with the monospecific *Jadammina macrescens* foraminiferal assemblage. Since the distribution of foraminifera in intertidal environments is a direct function of altitude (a surrogate for flooding duration and frequency) (Horton *et al.*, 1999), this provides strong additional evidence that the voids ratio of a material is related to the position at which it formed within the intertidal frame.

7.2.4 Lithostratigraphy of the low marsh core

Lithostratigraphic analyses are based on particle size and loss on ignition measurements, sampling at 0.01 m resolution. These variables, plotted in relation to depth in the core, are displayed in Figure 7.7 with the voids ratio profiles and the visual stratigraphy of the low marsh cores.

It can be seen that particle size and organic content variation are directly related to the stratigraphic units. The uppermost stratum (L5a; 0.00 – 0.07 m) is characterised by the highest organic contents. Unit L4 (0.07 - 0.12 m) has a lower organic content (*c.* 15 %). Variations in particle size also correspond with stratigraphic layers; shifts in mean particle size and the range of particle size distribution are accompanied by visual changes in lithology.

The loss on ignition profile is strikingly similar to the voids ratio profile. The highest voids ratios (*c.* 4 – 6) correspond to the maximum organic contents (20 – 30 %) in unit L5a (0.00 – 0.07 m). In the less organic layers (particularly L4; 0.08 – 0.12 m), loss on ignition values fluctuate around 15 %. This corresponds to voids ratios of *c.* 2 – 3. This strongly suggests that organic content exerts a significant control on voids ratio. Indeed, the two variables are strongly correlated ($r = 0.85$, $p < 0.001$) (Table 7.1).

There is also a less pronounced relationship between particle size and voids ratio. The deepest stratigraphic unit (clay-rich silt with some well-humified organic matter, 0.17 – 0.38 m depth) has the finest particles (modal particle diameter of 8 – 9 μm; a medium silt (Blott and Pye, 2001)). Particle size is fairly constant within this stratigraphic unit, although it displays a subtle coarsening-upwards trend that may be linked to the observed variations in voids ratios with depth. In the overlying strata typified by lower organic contents and lower voids ratios (0.07 – 0.12 m depth), the grain size distributions coarsen to a range of approximately 8 – 800 μm, before fining again in the uppermost (surface) stratigraphic unit. Here, the model grain size is *c.* 10 – 30 μm and the overall range is in the region of 4 – 300 μm. Throughout the sequence, silt is inversely and weakly (r = -0.367), yet significantly (p = 0.03), correlated with voids ratio. Clay is positively correlated with voids ratio, although only weakly ($r = 0.353$, $p = 0.04$) (Table 7.1).

7.2.5 Chemostratigraphy of the low marsh core

The stratigraphic distribution of redox sensitive elements is displayed in Figure 7.8. Stratigraphically constrained cluster analysis was performed on the geochemical data using CONISS (Grimm, 1987; 1993) and the low marsh core displays definite chemostratigraphic zonation. Five chemozones were assigned; these broadly agree with the visual (stratigraphic) observations. For example, chemozones LC1and LC2 correspond with lithological unit L1. Similarly, chemozones LC3 and LC4 overlap with lithological units L2 – L4. Chemozones LC4 and LC5 occupy the same depth range as lithological unit L5a.

Geochemical variations in chemozones LC1 (0.34 – 0.24 m) and LC2 (0.22 – 0.16 m) are generally linear, with both increasing and decreasing trends being displayed. For example, SiO_2 displays a gradual yet consistent falling trend from the base of the core upwards, from *c.* 46 wt % to *c.* 42 wt % at 0.16 m. In chemozones LC1 and LC2, a

Table 7.1 Correlations between geotechnical, lithological and geochemical variables for the low marsh core.

		Voids ratio, e	$\log_{10}\sigma'$ (kPa)	Loss on ignition (%)	Sand (%)	Silt (%)	Clay (%)	SiO_2 (% wt)	Fe_2O_3 (% wt)	MnO (% wt)	CaO (% wt)	P_2O_5 (% wt)	S (ppm)
$\log_{10}\sigma'$ (kPa)	*r*	**-0.757**											
	p	**<0.001**											
Loss on ignition (%)	*r*	**0.853**	**-0.605**										
	p	**<0.001**	**<0.001**										
Sand (%)	*r*	-0.021	0.082	0.095									
	p	0.905	0.639	0.586									
Silt (%)	*r*	**-0.367**	**0.717**	-0.182	**-0.350**								
	p	**0.030**	**<0.001**	0.294	**0.039**								
Clay (%)	*r*	**0.353**	**-0.725**	0.087	**-0.515**	**-0.623**							
	p	**0.037**	**<0.001**	0.618	**0.002**	**<0.001**							
SiO_2 (% wt)	*r*	**-0.507**	0.230	-0.466	**0.516**	-0.232	-0.307						
	p	**0.032**	0.358	0.051	**0.028**	0.354	0.215						
Fe_2O_3 (% wt)	*r*	-0.044	0.044	-0.162	**-0.737**	0.327	0.442	**-0.659**					
	p	0.864	0.862	0.521	**<0.001**	0.185	0.066	**0.003**					
MnO (% wt)	*r*	**0.668**	-0.450	**0.611**	0.347	-0.415	0.039	-0.402	0.099				
	p	**0.002**	0.061	**0.007**	0.159	0.087	0.879	0.098	0.697				
CaO (% wt)	*r*	**-0.562**	**0.810**	-0.334	0.257	**0.496**	**-0.730**	0.421	-0.342	-0.316			
	p	**0.015**	**<0.001**	0.175	0.304	**0.036**	**0.001**	0.082	0.165	0.202			
P_2O_5 (% wt)	*r*	**0.890**	**-0.629**	**0.810**	0.111	-0.355	0.221	**-0.585**	0.090	**0.636**	**-0.497**		
	p	**<0.001**	**0.005**	**<0.001**	0.660	0.149	0.378	**0.011**	0.723	**0.005**	**0.036**		
S (ppm)	*r*	**0.687**	-0.253	**0.817**	0.379	0.006	-0.392	-0.457	-0.223	**0.486**	-0.115	**0.682**	
	p	**0.002**	0.311	**<0.001**	0.121	0.981	0.107	0.056	0.373	**0.041**	0.649	**0.002**	
I (ppm)	*r*	0.387	0.121	**0.521**	0.080	0.295	-0.360	**-0.666**	0.333	**0.553**	0.008	**0.551**	**0.718**
	p	0.112	0.633	**0.027**	0.752	0.234	0.142	**0.003**	0.177	**0.017**	0.976	**0.018**	**0.001**

r = Pearson's correlation coefficient. *p* = significance. Significant correlations are in bold type.

Table 7.2 Correlations between geotechnical, lithological and geochemical variables for the mudflat core.

		Voids ratio, *e*	$\log_{10}\sigma'$ (kPa)	Loss on ignition (%)	Sand (%)	Silt (%)	Clay (%)	SiO_2 (% wt)	Fe_2O_3 (% wt)	MnO (% wt)	CaO (% wt)	P_2O_5 (% wt)	S (ppm)
$\log_{10}\sigma'$ (kPa)	*r*	**-0.694**											
	p	**<0.001**											
Loss on ignition (%)	*r*	**0.440**	-0.270										
	p	**0.002**	0.060										
Sand (%)	*r*	-0.235	**0.657**	**0.293**									
	p	0.105	**<0.001**	**0.041**									
Silt (%)	*r*	**-0.668**	**0.608**	**-0.354**	-0.021								
	p	**<0.001**	**<0.001**	**0.013**	0.886								
Clay (%)	*r*	**0.503**	**-0.861**	-0.113	**-0.901**	**-0.414**							
	p	**<0.001**	**<0.001**	0.438	**<0.001**	**0.003**							
SiO_2 (% wt)	*r*	-0.050	0.196	**-0.509**	0.027	0.122	-0.075						
	p	0.811	0.349	**0.009**	0.897	0.560	0.721						
Fe_2O_3 (% wt)	*r*	**0.721**	**-0.751**	**0.667**	-0.223	**-0.628**	**0.464**	**-0.526**					
	p	**<0.001**	**<0.001**	**<0.001**	0.284	**0.001**	**0.019**	**0.007**					
MnO (% wt)	*r*	**0.531**	-0.321	-0.064	-0.129	**-0.429**	0.295	**0.561**	0.197				
	p	**0.006**	0.117	0.762	0.538	**0.032**	0.152	**0.003**	0.344				
CaO (% wt)	*r*	-0.213	0.373	**0.518**	0.396	0.129	**-0.430**	-0.256	-0.073	-0.220			
	p	0.307	0.066	**0.008**	0.050	0.538	**0.032**	0.217	0.729	0.292			
P_2O_5 (% wt)	*r*	**0.561**	**-0.843**	0.031	**-0.623**	**-0.513**	**0.802**	0.015	**0.584**	**0.486**	**-0.553**		
	p	**0.004**	**<0.001**	0.884	**0.001**	**0.009**	**<0.001**	0.945	**0.002**	**0.014**	**0.004**		
S (ppm)	*r*	0.294	-0.214	**0.789**	0.146	-0.303	-0.019	**-0.802**	**0.663**	-0.335	**0.514**	-0.110	
	p	0.153	0.305	**<0.001**	0.486	0.141	0.928	**<0.001**	**<0.001**	0.101	**0.009**	0.599	
I (ppm)	*r*	-0.237	**0.639**	0.340	**0.806**	0.063	**-0.797**	-0.068	-0.132	-0.011	0.459	**-0.465**	0.176
	p	0.255	**0.001**	0.096	**<0.001**	0.766	**<0.001**	0.748	0.529	0.958	0.021	**0.019**	0.400

r = Pearson's correlation coefficient. *p* = significance. Significant correlations are in bold type.

gradual upward falling trend in CaO concentration can be observed (maximum of *c.* 2.75 wt % at 0.36 m). An upward rising trend in P_2O_5 is evident within chemozones LC1 and LC2, although only between concentrations of 0.1 – 0.3 wt %.

S and I have increased concentrations in chemozones LC1 and LC2 relative to the overlying chemozone, LC3 (0.14 – 0.08 m). In LC1 and LC2, I shows no persistent increasing or decreasing trends in concentrations, remaining between *c.* 75 and 125 ppm. S begins to decrease following its peak (*c.* 5500 ppm) at 0.16 m. The darker colour of lithological unit L1 does indeed seem to be related to the increase in S, reflecting the bacterial reduction of sulphate and the presence of black iron sulphides (Spencer *et al.*, 2003). However, the partly mottled nature of this stratum indicates that a stable redox zonation has not developed at this depth as a result of the fluctuating (over weekly to annual timescales) water table. The dynamic water table causes short-term fluctuations in the redox boundary (Casey and Lasaga, 1987).

Chemozone LC4 displays enrichment in MnO, P_2O_5, S and I in comparison to the underlying geochemical unit, LC3, and the overlying LC5. LC4 also displays relative depletion of SiO_2. Other than depletion in zone LC5 (values of *c.* 3.5 wt %), Fe_2O_3 displays little variation with stratigraphy or chemozone. Concentrations fluctuate around these values until a depth of *c.* 0.28 m, before dropping to 5 -6 wt % within chemozone LC1.

CaO concentrations largely correspond with the visual stratigraphy, falling from *c.* 1.75 wt % at the surface (top of chemozone LC5) to *c.* 0.5 wt % at 0.10 m (LC3). This is followed by a more rapid increase in concentration to levels similar to those observed at the surface in the lithological unit L2 (0.13 - 0.15 m depth).

MnO concentrations are generally constant downcore, fluctuating between 0.0 and 0.1 wt %, other than a peak at 0.04 m (0.45 wt %) in chemozone LC4. Similarly, P_2O_5 concentrations range from just *c.* 0.1 to 0.6 % by weight, and I levels never exceed 225 ppm. Variations in the geochemical profile are instead dominated by SiO_2 (*c.* 35 -55 wt %) and to a lesser extent by variations in Fe_2O_3 (3 – 7 wt %), S (2000 – 6500 ppm) and CaO (0 – 3 wt %).

SiO_2 and CaO are inversely correlated with the voids ratio variations (Table 7.1). In contrast, P_2O_5 and S show direct relationships with voids ratio. The correlation between

P_2O_5 and voids ratio in particular is very strong ($r = 0.890$, $p < 0.001$). Interestingly, I is not significantly correlated with voids ratio ($r = 0.387$, $p = 0.112$) despite displaying similarities in the form of the respective depth profiles. In contrast, MnO correlates fairly strongly and significantly ($r = 0.668$, $p = 0.002$) with voids ratio; such a relationship is perhaps not evident in the depth profiles (Figure 7.8).

7.2.6 Voids ratio profile of the mudflat core

The voids ratio profiles of cores MFX 1 -5 are displayed together in Figure 7.9. As in the low marsh cores, the root squared error (as displayed in Figure 7.9 a) is small and so only the mean value of the upward and downward scans is considered (Figure 7.9 b).

After the x-ray scanning procedure, cores MFX-1 and MFX-2 were extruded, horizontally split and photographed. Cores MFX-3, MFX-4 and MFX-5 were extruded with care to prevent disturbance and prepared for oedometer testing. Cores were also sampled for litho-, bio- and chemo-stratigraphic analysis.

Figure 7.10 displays photographs of the cores, overlain by the voids ratio profiles at each depth and the visual stratigraphy. There is substantial lateral and vertical variation in voids ratios within lithological unit M1, both within and between cores. These variations do not appear to be linked to any changes in lithology. There is an offset between cores obtained from the same sample depths (i.e. between cores MFX-1 and MFX-4; and between MFX-2 and MFX-3) that is not attributable to variations in the electrical mains supply between different scans, since each core scanning session was individually calibrated. The voids ratio profiles are also characterised by occasional 'upward' and 'downward' peaks at, for example *c.* 0.05 m (MFX-1), *c.* 0.10 – 0.12 m (MFX-1 and MFX-4) and at *c.* 0.14 m (MFX-4) (Figure 7.9 b). The low standard errors associated with the voids ratio profiles displayed in Figure 7.9 (a) indicate that these peaks are indeed 'real' and are not an artefact of sampling error. However, given the visual uniformity of lithology downcore, they cannot be directly linked to variations in grain size or organic content and may reflect the range of depositional structures caused by variations in depositional and/or subaerial conditions. An interesting feature of core MFX-5 is the decrease in voids ratio from *c.* 1.9 to *c.* 1.7 between 0.36 and 0.40 m, and the subsequent increase back to *c.* 2 at 0.46 m. Again, this feature cannot be attributed to any obvious lithological variation.

In the uppermost (oxidised) stratum (M2; 0.03 – 0.00 m), voids ratios increase rapidly from *c.* 2.0 to 2.8, reflecting the more open structures associated with newly deposited sediment, variations in the presence of bioturbation structures and the higher occurrence of these structures towards the depositional surface.

7.2.7 Effective stress profile of the mudflat core

Despite minor variations in gradient that are attributable to small structural (density) fluctuations between cores, it is evident from Figure 7.11 that effective stress increases linearly downcore, reaching a maximum of *c.* 2.55 at the base of the core. This linear increase of effective stress with depth suggests downcore lithological and geotechnical homogeneity (overconsolidation) within of the mudflat core.

7.2.8 Biostratigraphy of the mudflat core

Samples were taken from the mudflat cores generally at a resolution of 0.04 m. Foraminiferal assemblages (as a percentage of the total counted at each sampling depth) and total counts are displayed in Figure 7.12 in relation to depth, voids ratio profile and visual stratigraphy. Species constituting less than 2 % of the total foraminifera counted are excluded from the diagrams for clarity.

The mudflat core is characterised by greater species diversity than the low marsh core. The majority of foraminifera (always greater than 90 %, and generally greater than 95 %) are calcareous, confirming that the core was formed in a tidal flat environment (Horton, 1997; Horton, 1999; Horton and Edwards, 2000). This is consistent with the lithological characteristics of the core. The remaining 5 -10 % of foraminifera (*Jadammina macrescens* and *Miliammina fusca*) are agglutinated and are likely to be allochthonous. Total counts are higher towards the base of the core, reaching 500 at 0.44 m. Counts drop below 200 above *c.* 0.15 m, and below 100 at *c.* 0.12 m. Environmental interpretations from these low (< 100) counts must be treated with caution.

Stratigraphically constrained cluster analysis using CONISS produced four foraminiferal zones. In all zones, foraminiferal assemblages are dominated by *Haynesina germanica*, which commonly constitutes > 50 % of the total. *Stainforthia fusiformis* is the second most abundant species and generally contributes between 20 and 30 % to the total foraminifera counted.

In the deepest zone (MF1; 0.44 – 0.36 m), the relative abundance of *Haynesina germanica* is at its lowest, falling below 50 %. There is a commensurate increase in the proportion of *Elphidium magellanicum* (*c.* 5 – 10 %) in this zone. The proportion of *Haynesina germanica* increases to *c.* 55 – 70 % in zone MF2.

Maximum counts of *Stainforthia fusiformis* relative to other species (> 25 %) are found in zone MF3 (0.22 – 0.08 m). There is a corresponding fall in the proportion of *Haynesina germanica* (*c.* 40 – 55 %). In zone MF4 (0.04 – 0.00 m), *Haynesina germanica* again increases in dominance to *c.* 75 %. This results in a fall in the proportion of *Stainforthia fusiformis* (< 20 % in MF4).

The remaining species in the core make only minor contributions. *Quinqueloculina* spp. and *Cibicides lobatulus* are the most significant of the minor species in zones MF3 and MF4.

There seems to be very little relationship between foraminiferal assemblage and the voids ratio profile. However, a subtle direct relationship may exist between variations in the abundance of *Haynesina germanica* and voids ratio, particularly at depths between 0.15 and 0.25 m. This may suggest an influence of relative elevation on the voids ratio profile, possibly caused by initial structural variations.

7.2.9 Lithostratigraphy of the mudflat core

Particle size and loss on ignition variation, sampled at 0.01 m resolution, are displayed in Figure 7.13 in relation to depth, the voids ratio profile and the visual stratigraphy of the mudflat cores.

Downcore particle size variations reveal three main units. The top unit, coinciding with the oxidised layer (0.00 – 0.03 m) is the coarsest section of the core. Particle size generally ranges from *c.* 8 – 600 μm, with a modal diameter of *c.* 50 – 80 μm (very fine sand). Between 0.03 and *c.* 0.18 m, the particle size distribution becomes finer, ranging from *c.* 4 – 300 μm, with a modal diameter of 8 – 10 μm (medium silt). At greater depths (> 0.18 m), particle size fines again, ranging from 0.9 to 100 μm, with a model grain size of 7 – 8 μm (fine/medium silt). This coarsening upwards sequence is indicative of increasing energy conditions through time, suggesting an increase in tidal water depths (positive sea level tendency). These subtle variations in particle size do not appear to be linked to the voids ratio profile.

The loss on ignition profile varies considerably, ranging from values typical of the contemporary mudflat (Section 5.1.2) (*c.* 12 %) at the surface of the core to a maximum of *c.* 19 % at 0.25 m depth. These values then fall to *c.* 13 % towards the base of the core (> 0.40 m) and remain largely constant. The higher loss on ignition values in the core are similar to those found in the contemporary low marsh sub-environment at Greatham Creek. From the visual description of the lithology and from the foraminifera present within the core (Section 7.3.5), it is evident that the material was not formed in a saltmarsh environment. Changes in loss on ignition values are likely to reflect variations in detrital organic matter in the core that has been transported onto the mudflat during times of high productivity, or maybe erosion, in the saltmarsh. Alternatively, the higher loss on ignition values may result from the combustion of chemical elements, such as sulphur or carbonate.

The considerable variations in particle size and loss on ignition are not reflected in the voids ratio profile, at least by visual observation of the data alone. The quantitative lithological data displayed in Figure 7.13 also suggest that the voids ratio profile of the core does not seem to be related to its lithology. Surprisingly, therefore, when the particle size distributions are split into sand, silt and clay fractions, voids ratio significantly correlates with silt (r = -0.668, p < 0.001) and clay (r = 0.503, p < 0.001) (Table 7.3). Variations in loss on ignition also correlate with those in voids ratio, albeit weakly (r = 0.440, p = 0.002).

7.2.10 Chemostratigraphy of the mudflat core

The depth distributions of SiO_2, Fe_2O_3, MnO, CaO, S and I are displayed in Figure 7.14 in relation to depth, the voids ratio profile and the visual stratigraphy of the mudflat core. The redox zonation of the core, as noted previously, is reflected in the geochemistry of the core. In particular, S shows depletion (*c.* 4500 – 7000 ppm) in the light brown oxic surface layer (chemozone MC3); 0.00 – 0.03 m) and enrichment (*c.* 10500 ppm) in the black anoxic zone (chemozones MC1 and MC2; > 0.03 m depth). This confirms previous speculations that sediment beneath the 0.03 m redox boundary remains permanently saturated, probably by capillary action, and is never exposed to oxygen. CaO also shows depletion in the oxic layer (chemozone MC3) and this may also reflect the stong redox zonation of the core. However, neither S (r = 0.294, p = 0.153) nor CaO (r = -0.213, p =

0.307) (Table 7.2) is strongly nor significantly correlated with voids ratio, suggesting the inferred redox mobilisation of these compounds has little effect on *in situ* voids ratio.

Other compounds display variation above and below the redox boundary; SiO_2, MnO and, to a lesser extent, P_2O_5 display enrichment in chemozone MC3. This is initially followed by sudden decreases in concentrations of this substance in the anoxic chemozone MC2. However, redox zonation cannot account for the fluctuations in the concentrations of these substances below the redox boundary within chemozone MC1, particularly the distributions of SiO_s and MnO. Indeed, SiO_2 concentrations do not appear to be related to the voids ratio profile, neither visually nor statistically (r = -0.050, p = 0.811), although the increase and subsequent decrease in SiO_2 between 0.35 and 0.48 m in MC1 is broadly coincident with the anomalous increase/decrease trend in the voids ratio profile at the same depths, noted previously. MnO also varies proportionally with the voids ratio profile at these depths. Indeed, MnO concentrations visually mimic the voids ratio variations downcore and the two variables correlate significantly, if not particularly strongly (r = 0.531, p = 0.006). The strongest correlation, however, is between Fe_2O_3 and voids ratio (r = 0.721, p < 0.001). The general decreasing trend in voids ratio downcore is also visible in the Fe_2O_2 profile, particularly in chemozone MC1.

7.3 MULTIPLE REGRESSION ANALYSIS OF VARIABLES AFFECTING IN SITU VOIDS RATIOS

On the basis of visual examination of the data and correlation coefficients (Tables 7.1 and 7.2), it is evident that relationships exist between the lithological and geochemical variables and the *in situ* voids ratios in the cores. However, observations and simple correlations do not reveal the relative contributions of different predictor variables on voids ratios. This can be undertaken by employing multiple regression techniques and considering the standardised beta coefficient, which is a measure of how strongly each predictor (independent) variable influences the dependent variable (voids ratio, *e*, in this case). It is measured in units of standard deviation; for example, a beta value of 2 indicates that a change of one standard deviation in the predictor variable will result in a change of 2 standard deviations in the dependent variable. Hence, the higher the standardised beta coefficient, the greater the impact of the predictor variable on the dependent variable. The standardised beta coefficient allows comparisons to be made regarding the strength of the relationship of each predictor variable to the dependent variable.

In the multiple regression analysis below, the adjusted *r*-square (r^2_{adj}) value is used, since this takes into account the number of independent variables in the model and the number of observations upon which the model is based. Conventional r^2 values tend to overestimate the predictive capacity of a model, and will always be increased by adding predictor variables to the model. r^2_{adj} is therefore considered to be a more useful measure of the relative 'success' of different regression models.

Mean values of the dependent variable, voids ratio (*e*), at 0.01 m resolution were taken to remove the effect of the observed rapid and often large variations therein. Similarly, mean values of effective stress were taken at a similar sampling interval to account for minor, inter-core variations in the effective stress profile.

7.3.1 Multiple regression analysis of the low marsh core

Initially, all independent variables (effective stress, and lithological and geochemical parameters) were simultaneously entered into a multiple regression model. An overall statistically significant model emerged ($F_{11, 6}$ = 17.897, p = 0.001) with a very high predictive capacity (r^2_{adj} = 0.92). Predictor variables are displayed in Table 7.3 (along with their significance), ranked according to their relative importance in predicting the voids ratio, *e*, as indicated by their respective standardised beta coefficients.

The most important predictor variable is S (ppm). This is followed by silt content. I (ppm), loss on ignition and sand content have similar relative predictive power. Interestingly, effective stress ($\log_{10}\sigma'$ (kPa)) ranks lowly, suggesting it has a relatively minor role in predicting voids ratios in the overconsolidated vadose zone. However, the validity of these rankings can be called into question because only two predictor variables (S and silt) are statistically significantly linearly related to voids ratio at the 0.05 significance level.

Regression based only on effective stress ($\log_{10}\sigma'$, kPa) still produces an overall statistically significant model ($F_{1, 33}$ = 44.40, p < 0.001). In addition, effective stress is a statistically significant predictor variable (t = -6.664, p < 0.001). However, the predictive capacity of the model is reduced (r^2_{adj} = 0.56).

If the lithological variables (loss on ignition, sand, silt and clay contents) are entered into the model with effective stress, the model remains significant ($F_{4, 30}$ = 34.57, p < 0.001) and increases substantially in predictive power (r^2_{adj} = 0.80). However, a regression model based only on lithology performs equally well ((r^2_{adj} = 0.79) whilst remaining significant ($F_{4, 30}$ = 34.57, p < 0.001). This suggests that effective stress is not a dominant control on *in situ* voids ratio and that material/lithological factors are playing a more important role.

When geochemical variables are used as predictors, an r^2_{adj} of 0.87 is obtained ($F_{7, 10}$ = 17.28, p < 0.001). This performance of the model does not improve when effective stress is included as a predictor variable (r^2_{adj} = 0.86; $F_{8, 9}$ = 13.81, p < 0.001).

Table 7.3 Multiple regression parameters and the significance of predictor variables obtained from multiple regression analysis of the low marsh sediment core. The dependent variable is voids ratio, *e*. Variables are ranked according to their relative importance in prediction of *e*, as indicated by standardised beta coefficients. Significant predictor variables are shown in bold.

	Unstandardised beta coefficient	Standardised beta coefficient	*t* statistic	*p*
Intercept (constant)	4.54	-	0.97	0.368
S (ppm)	**0.001**	**1.50**	**3.12**	**0.021**
Silt (%)	**-0.10**	**-0.81**	**-2.44**	**0.050**
I (ppm)	-0.01	-0.567	-0.57	0.352
Loss on ignition (%)	-0.12	-0.56	-0.56	0.067
Sand (%)	-0.06	-0.50	-0.50	0.251
CaO (wt %)	0.51	0.42	0.42	0.145
Fe_2O_3 (wt %)	0.48	0.41	0.41	0.275
P_2O_5 (wt %)	2.69	0.38	0.38	0.130
MnO (wt %)	2.234	0.26	0.26	0.131
$\log_{10}\sigma'$ (kPa)	-0.20	-0.09	-0.09	0.872
SiO_2 (wt %)	0.01	0.05	0.05	0.889

It is clear that the most powerful predictive model was created when all variables were used in the regression analysis (r^2_{adj} = 0.92). This suggests that all variables need to be considered in analysis of voids ratio variation in these overconsolidated vadose zone sediments. However, many of the predictor variables used were statistically insignificant, with the t statistics not exceeding the critical value at either the 0.01 or 0.05 significance levels. In this case, the t statistic is being used to test the hypothesis that each predictor variable is linearly related to the dependent variable. Furthermore, such a model, with a large number of predictor variables, may not provide optimum parsimony.

In order to overcome these problems, backward elimination multiple regression was employed. This stepwise regression procedure begins by regressing the dependent variable on all of the predictor variables. If any variables are statistically insignificant, the variable making the smallest contribution to predictive capacity (the variable with the lowest partial correlation with the dependent variable) is considered for removal. The partial correlation measures the strength of the association between the response variable and each of the predictor variables not currently in the equation after removing the effect of the variables currently in the equation. Predictors variables are only removed if the significance of the change to the F statistic exceeds 0.10 (the removal criterion). Otherwise, it remains in the model. This variable elimination procedure continues until there are no variables in the equation that exceed the removal criterion.

For the low marsh core, using the backward elimination multiple regression method, a highly significant model ($F_{9,8} = 18.264$, $p < 0.001$) with excellent predictive capacity (r^2_{adj} = 0.937) emerged. Indeed, this r^2_{adj} is greater than that obtained from all variables. The significant predictor variables are displayed in Table 7.4.

Perhaps most significantly, effective stress has been excluded from the model; its influence is not sufficiently significant to warrant its inclusion. According to the model, the most important predictor variable is S (ppm); its influence in predicting voids ratio is almost twice as important of that of the variable ranked second in order of importance in prediction (silt).

7.3.2 Multiple regression analysis of the mudflat core

Simultaneous entry of all dependent variables produced a significant model ($F_{11,13} = 7.383$, $p = 0.001$) with good predictive capacity ($r^2_{adj} = 0.75$). Predictor variables are displayed in Table 7.5 (along with their significance), ranked according to their relative importance in predicting the voids ratio, e, as indicated by their respective standardised beta coefficients. Loss on ignition is the most influential predictor variables, followed by CaO and MnO. As in the low marsh analysis, effective stress has a relatively minor role in predicting the voids ratio in the overconsolidated vadose zone. Additionally, only two predictor variables are statistically significant (loss on ignition and CaO), although MnO borders on statistical significance at the 0.05 significance level.

Using only effective stress ($\log_{10}\sigma'$, kPa) as a predictor variable, a significant model emerges ($F_{1, 47}$ = 43.63, p < 0.001). The predictive capacity of the model, however, is reduced (r^2_{adj} = 0.47). By adding the lithological variables into the model, statistical significance remains ($F_{4,44}$ = 17.60, p < 0.001) and the r^2_{adj} value increases to 0.58. However, when effective stress is not entered into the model and only the lithological predictor variables remain, the model performs no worse (r^2_{adj} = 0.58; $F_{3,45}$ = 22.814, p < 0.001).

Table 7.4 Multiple regression parameters and the significance of predictor variables obtained from backward elimination stepwise multiple regression analysis of the low marsh sediment core. The dependent variable is voids ratio, *e*. Variables are ranked according to their relative importance in prediction of *e*, as indicated by standardised beta coefficients. Two predictor variables, I (ppm) and Fe_2O_3, are statistically insignificant (shown in bold) but remain in the model due to their significance to overall model performance.

	Unstandardised beta coefficient	Standardised beta coefficient	*t* statistic	*p*
Intercept (constant)	4.87	-	2.10	0.690
S (ppm)	0.001	1.51	5.07	0.001
Silt (%)	-0.10	-0.81	-3.45	0.009
I (ppm)	**-0.01**	**-0.66**	**-2.97**	**0.18**
Loss on ignition (%)	-0.10	-0.54	-2.88	0.020
Sand (%)	-0.05	-0.45	-2.33	0.048
Fe_2O_3 (wt %)	**0.51**	**0.43**	**2.11**	**0.067**
P_2O_5 (wt %)	2.89	0.41	2.81	0.023
CaO (wt %)	0.47	0.39	2.48	0.038
MnO (wt %)	2.28	0.27	2.47	0.039

Similarly, there is little difference in the predictive capacity of regression models based upon geochemistry alone (r^2_{adj} = 0.63; $F_{7, 17}$ = 6.947, p = 0.001), and geochemistry and effective stress in combination (r^2_{adj} = 0.65; ($F_{8, 16}$ = 6.487, p = 0.001). In the latter model, effective stress possesses the highest standardised beta coefficient, indicating its importance as a predictor variable.

Since the most powerful model is produced by using all variables simultaneously, it is clear that effective stress is not the only control on *in situ* voids ratios and the predictive capacity is increased by incorporating lithological and diagenetic variables.

Using the more statistically rigorous backward elimination method, a highly significant model ($F_{5,19}$ = 18.264, p < 0.001) with good predictive capacity (r^2_{adj} = 0.78) emerged. Indeed, with fewer predictor variables, the predictive capacity of the model is greater than that of the simultaneous entry method. Predictor variables are displayed in Table 7.6.

Table 7.5 Multiple regression parameters and the significance of predictor variables obtained from multiple regression analysis of the mudflat sediment core. The dependent variable is voids ratio, *e*. Variables are ranked according to their relative importance in prediction of *e*, as indicated by standardised beta coefficients. Significant predictor variables are shown in bold.

	Unstandardised beta coefficient	Standardised beta coefficient	*t* statistic	*p*
Intercept (constant)	1.27	-	0.76	0.463
Loss on ignition (%)	**0.05**	**1.00**	**2.33**	**0.036**
CaO (wt %)	**-0.20**	**-0.79**	**-2.44**	**0.030**
MnO (wt %)	36.82	0.70	2.15	0.051
Fe_2O_3 (wt %)	-0.13	-0.54	-0.823	0.422
S (ppm)	2.77×10^{-5}	0.45	0.90	0.385
P_2O_5 (wt %)	-1.35	-0.43	-1.12	0.285
Sand (%)	-0.01	-0.38	-0.82	0.427
$\log_{10}\sigma'$ (kPa)	-0.05	-0.30	-0.61	0.552
Silt (%)	-0.004	-0.06	-0.18	0.859
I (ppm)	-0.001	-0.06	-0.20	0.847
SiO_2 (wt %)	0.002	0.02	0.08	0.938

Table 7.6 Multiple regression parameters and the significance of predictor variables obtained from backward elimination stepwise multiple regression analysis of the mudflat sediment core. The dependent variable is voids ratio, *e*. Variables are ranked according to their relative importance in prediction of *e*, as indicated by standardised beta coefficients. All predictor variables significantly contribute to the model.

	Unstandardised beta coefficient	Standardised beta coefficient	*t* statistic	*p*
Intercept (constant)	1.06	-	2.87	0.010
$\log_{10}\sigma'$ (kPa)	-0.10	-0.68	-3.52	0.002
Loss on ignition (%)	0.03	0.59	4.48	<0.001
P_2O_5 (wt %)	-1.64	-0.52	-2.35	0.030
MnO (wt %)	26.53	0.51	4.51	<0.001
CaO (wt %)	-0.11	-0.44	-2.92	0.009

With this more robust method of multiple regression, the most important predictor variable is effective stress. The remaining predictor variables suggest minor, additional controls on *in situ* voids ratio are exerted by the organic content and geochemical concentrations.

7.3.3 Factors affecting in situ voids ratios in the sample cores

If the multiple regression models above are valid, then variations in voids ratio in the low marsh core are primarily controlled by S (ppm), silt, I (ppm), loss on ignition, sand and, to a lesser extent, the remaining redox sensitive elements. In the mudflat core, where lithological variation is minimal, voids ratio is most strongly controlled by vertical effective stress, with additional contributions from loss on ignition, P_2O_5, MnO and CaO. This suggests that existing compression models are not only insufficient in terms of mechanical compression, but also may fail to predict voids ratio changes sufficiently well due to the importance of diagenetically sensitive elements. However, before models of autocompaction are developed on the basis of these findings, a critical and sensible assessment of the multiple regression models is required.

Two main issues arise which suggest that geochemical factors are not necessarily useful predictors of voids ratio and hence cast doubt upon the findings of the multiple regression analysis. Firstly, if the relative concentrations of the most important predictor variables in the low marsh core are considered, the integrity of their predictive power can be questioned. The trace elements S and I are present in extremely low concentrations (2000 – 6000 ppm and 50 – 525 ppm). Similarly, MnO (0.1 – 0.5 wt %), P_2O_5 (0.1 – 0.6 wt %) and CaO (1 – 2 wt %) make minimal contributions to the geochemical composition of the sediment. It therefore seems doubtful that such low concentrations of these substances, and the low downcore variations therein, are likely to exert such a dominant influence on voids ratio in relation to lithological variations. In the mudflat core, geochemical compounds and elements are also only present in what are deemed to be geotechnically insignificant quantities; P_2O_5 varies between 0.1 – 0.4 wt % and MnO between 0.054 – 0.060 %. Again, not only are these quantities small but it is also difficult to envisage the small variations therein significantly contributing to voids ratio changes. CaO may have more of an influence on *in situ* voids ratio, ranging between more substantial values of *c.* 4 and 7 %.

Secondly, by considering correlations between predictors variables (Tables 7.1 and 7.2), the argument that geochemical factors may not be useful predictors of *in situ* voids ratios gains strength. Strong relationships exist between lithological and geochemical variables. In the low marsh core, the two most important geochemical predictors variables, S (r = 0.817, $p < 0.001$) and I ($r = 0.521$, $p < 0.001$), both strongly correlate with loss on ignition, as do P_2O_5 ($r = 0.810$, $p < 0.001$) and MnO ($r = 0.611$, $p = 0.007$). CaO shows close association with the silt ($r = 0.496$, $p = 0.036$) and clay ($r = -0.730$, $p = 0.001$) fractions and Fe_2O_3 correlates with sand content ($r = -0.737$, $p < 0.001$).

In the mudflat core, relationships exist between the various lithological fractions and the three geochemical compounds selected using the backward elimination multiple regression method. MnO ($r = -0.429$, $p = 0.032$) and P_2O_5 ($r = -0.513$, $p = 0.009$) significantly correlate with the silt fraction. Indeed, P_2O_5 is associated with all grain size fractions (sand: $r = -0.623$, $p = 0.001$; clay: $r = 0.802$, $p < 0.001$). CaO is significantly correlated with loss on ignition ($r = 0.518$, $p = 0.008$) and the clay fraction ($r = -0.430$, p = 0.032).

Such strong correlations between geochemical and lithological variables suggest that the observed distribution of redox sensitive elements and compounds is not a function of redox remobilisation and zonation. Rather it may be a result of downcore compositional changes and/or due to preferential adsorption of redox sensitive substances onto different grain size fractions or organic substrates. The latter mechanism has been observed previously, and is driven by the bacterially-mediated remineralisation or organic carbon in decaying marsh vegetation. The available electron acceptors involved in the oxidation of organic carbon are successfully utilised by microbes in terms of decreasing thermodynamic advantage (O_2 before NO^{-}_3 before MnO_x before Fe_2O_3 before seawater SO^{2-}_4 before methanogeneisis; Froelich *et al.*, 1979). When such oxyhydroxides enter anoxic conditions (beneath the permanent water table) during burial, they undergo reductive dissolution (Thomson *et al.*, 2002), releasing any associated/complexed elements (Zwolsman *et al.*, 1993). As a consequence, dissolved diagenetically sensitive geochemical substances become mobile in pore water solution and are free to move vertically or advectively through the sediment to precipitate as or with hydrous Fe and Mn oxides in the oxic (or at least periodically oxic, mottled) zone. Generally, this occurs at the top of the sulphate reduction zone. However, where redox zonation is not clearly defined, as occurs at depths associated with a fluctuating water table and hence lacking a strong redox boundary, diagenetically mobile elements (particularly S and I) may instead be adsorbed onto organic compounds which provide reactive sorption sites (Cundy and Croudace, 1995a; A. Cundy, *pers. comm.*; A. Plater, *pers. comm*; Plater *et al.*, 1998). Such a phenomenon has also been documented in the clastic component by Zwolsman *et al.* (1993) and Cundy and Croudace (1995). In conjunction with the significant correlations between geochemical and lithological variables, this mechanism suggests that the geotechnical importance assigned to geochemical substances is in fact an artefact of sediment composition and preferential adsorption of redox sensitive elements onto various components of the lithology/substrate. Consequently, the predictive importance assigned to geochemical variables by the multiple regression modelling procedure can be attributed to their close association with lithological parameters.

On the basis of these arguments involving (insignificantly) low concentrations of elements and preferential adsorption onto various lithological fractions, it seems that the dominant factor in causing *in situ* voids ratio change within the overconsolidated low marsh vadose zone sediments is lithology (in the low marsh cores obtained) and effective stress (in the mudflat cores obtained). It is highly likely, therefore, that the multiple regression modelling is misleading and the observed trends in geochemistry are likely to be an artefact of preferential binding of redox sensitive elements onto lithological carrier phases, particularly the organic fraction. Geochemical factors, therefore, are judged to be passive and their distribution reflects their association with the 'true' (lithological) controls on voids ratio.

By disregarding geochemical variables, variations in voids ratio can now only be explained by variations in lithology and effective stress. Hence, in the low marsh cores, where lithology was highly variable, lithology is the most important predictor variable. This does not indicate that an effective stress modelling approach is insufficient. Rather, it is a reflection of the observed downcore variations in lithology and strongly highlights the need to develop individual compression models for each lithology in the low marsh core for successful pre- and retro-diction of volumetric change.

In the mudflat core, where lithology is constant, effective stress is the most important predictor variable for calculations of voids ratio. It is therefore likely that a stress-based compression model is sufficient and that geochemical parameters need not be incorporated into the autocompaction model.

7.4 THE EFFECTS OF DIAGENETIC POINT CONTACT CEMENTATION ON COMPRESSION BEHAVIOUR

Other than by filling void spaces following their precipitation, diagenetically sensitive substances can also act as point-contact cements (Hawkins, 1984; Tovey and Yim, 2002) that have the potential to change material compressibility and alter future mechanical behaviour from that observed in the virgin materials obtained from the depositional surface.

In order to ascertain the significance of such effects in compression behaviour, it is necessary to undertake oedometer tests on the diagenetically altered materials obtained from depth within the stratigraphic sequences analysed above. However, due to the rapid stratigraphic variation in lithology in the low marsh cores, it is only the lithologically homogenous mudflat cores that can reliably be used to determine any effects of diagenetic cements on compression behaviour. Therefore, oedometer samples were obtained from cores MFX-2, MFX-3 and MFX-5 at depths that showed noticeable relative enrichment or depletion in diagenetically sensitive substances (Figure 7.14). These sampling depths, along with sample IDs, are shown in Figure 7.15 and in Table 7.7, which also displays the lithological characteristics and initial structural parameters. Specific gravity measurements were made at 0.25 m and 0.49 m to ascertain whether any changes occurred downcore. Values were similar (mean = 2.63) and equalled those obtained from the surface samples (Table 5.2). $e\log_{10}\sigma'$ are displayed collectively in Figure 7.16. Material properties obtained from these plots are displayed in Table 7.8. Samples were tested for their loading behaviour only according to low stress scenario 1 (Table 6.2), followed by the incremental loading steps displayed in Table 6.1.

The downcore homogeneity of lithology has been noted previously but is confirmed by the data presented in Table 7.7. Values of initial voids ratio from the core samples (1.80 – 2.17; Table 7.7) fall within the same range as those measured on the surface samples (1.71 – 2.68; Table 6.5).

Table 7.7 Physical properties of oedometer samples obtained from cores MFX-4, MFX-5 and MFX-6.

Sample I.D.	Approximate sampling depth (m)	Loss on ignition (%)	Sand (%)	Silt (%)	Clay (%)	Natural moisture content, w (%)	Initial voids ratio, e_i
MFX-1-S-OED-24	0.17	18.65	7.1	75.5	17.4	77.00	2.16
MFX-2-S-OED-24	0.22	18.15	16.6	71.7	11.7	78.29	2.17
MFX-3-S-OED-24	0.30	17.06	13.0	75.5	11.4	76.38	2.09
MFX-4-S-OED-24	0.34	16.87	17.1	72.8	10.1	76.49	2.10
MFX-5-S-OED-24	0.40	13.50	10.3	78.9	10.9	68.93	1.80
MFX-6-S-OED-24	0.46	13.39	14.5	75.3	10.2	71.52	1.93

Table 7.8 Material properties of oedometer samples obtained from cores MFX-4, MFX-5 and MFX-6.

Sample I.D.	C_r	C_c	σ'_c (kPa)
MFX-1-S-OED-24	0.07	0.63	18
MFX-2-S-OED-24	0.09	0.62	17
MFX-3-S-OED-24	0.09	0.60	18
MFX-4-S-OED-68	0.09	0.62	17
MFX-5-S-OED-NC	0.09	0.53	16
MFX-6-S-BPS-24	0.09	0.57	18

The $e\log_{10}\sigma'$ plots in Figure 7.16 display compression trends that are consistent with those obtained from the surface samples in their virgin state. All samples are again overconsolidated. However, preconsolidation stresses range from 16 – 18 kPa, compared to 8 – 14 kPa in the surface samples. Also, the range of recompression indices is lower in and more uniform in the core samples (0.07 – 0.09; Table 7.8) than the surface samples (0.02 – 0.25), perhaps indicating a removal, and hence decreased importance, in variations in depositional structure and, consequently, of compressibility in the overconsolidated section of the stratigraphic column. These variations in preconsolidation stress and recompression index may reflect the influence of the diagenetic point contact cement manifesting itself as an increased structural resistance to compression and a higher stress transition to virgin compression (Gutierrez and Wangen, 2005; Nygard *et al.*, 2004). However, this trend may equally be indicative of the closure with depth of the bioturbation hollows and an associated increase in the structural integrity of the sediment.

Post-yield, however, the compression behaviour of the core samples closely matches that of the surface samples. This is revealed by a similar range of values for the compression index in both surface (0.50 – 0.73; Table 6.7) and stratigraphic (0.53 – 0.63; Table 7.8) samples. This suggests that any diagenetic modifications to compressibility are associated with the overconsolidated part of the stress range. Once structural breakdown occurs, any influence of a point contact cement is removed and there is no evidence of any memory of this low stress increase in compressive strength.

Figure 7.17 compares the samples obtained from the cores (red lines) with those sampled from the depositional surface (tested in Chapter 6; black lines). From Figure 7.17, it can be seen that the overall influence of shallow burial and any post-depositional modification of the sediment is minimal. Compression stress paths are highly similar. Although compression parameters vary slightly, particularly in terms of the recompression index and the preconsolidation stress, the one-dimensional compression stress paths of the MFX core samples are very similar to those of the surface samples.

7.5 SUMMARY OF PRINCIPAL FINDINGS

The aim of this chapter was to determine if and how diagenetic processes operating in the overconsolidated vadose zone had any influence on *in situ* voids ratios, particularly relative to vertical effective stress, and on the one-dimensional compression behaviour of undisturbed samples obtained from depth. Two main post-depositional processes were identified in the vadose zone stratigraphies. Firstly, visual evidence of diagenetic remobilisation of redox sensitive substances was observed in both low marsh and mudflat cores. Secondly, humification was identified in the low marsh cores. In the absence of a suitable quantitative technique for measuring the degree of humification in predominantly mineralogenic intertidal sediments, humification was not able to be considered any further in the statistical prediction of *in situ* voids ratios.

Variations in voids ratio with depth were compared to the depth profiles of lithological, geochemical and geotechnical variables. In the low marsh core, lithology varies rapidly stratigraphically. This is in contrast to the lithologically homogeneous mudflat core. Geochemically, redox zonation in the low marsh cores seemed to be minimal. In the mudflat core, however, there is a pronounced redox boundary that is reflected to a certain degree by the depth profiles of geochemical elements, particularly S, MnO and CaO.

Multiple regression analysis was undertaken to determine the relative importance of effective stress and lithological and geochemical factors on *in situ* voids ratios. Backward elimination stepwise regression suggested that geochemical factors were the dominant control on voids ratios in the low marsh core. Geochemical variables were of secondary importance to effective stress in the mudflat core, yet were still statistically of some importance. However, the overall importance of these variables was questioned and, ultimately, rejected since concentrations of these elements were deemed to be too low to affect voids ratio changes. The distribution of geochemical substances is likely to be an artefact of preferential adsorption onto lithological substrates (particularly the organic fraction).

One-dimensional (oedometer) compression tests undertaken on mudflat samples obtained from depths in the cores that displayed noticeable enrichment/depletion in redox sensitive elements demonstrated similar compression stress paths to tests performed on virgin surface samples. However, higher preconsolidation stresses were noted in the stratigraphic samples and this was attributed to the effect of weak diagenetic point contact cements and/or the closure upon burial of bioturbation structures.

Research hypothesis 6 was defined in Section 3.10 as follows:

6. Diagenetic changes are unimportant and do not significantly modify mechanical compression behaviour.

On the basis of the data presented in this chapter (particularly those obtained from the mudflat core), effective stress remains the dominant control on voids ratio in the overconsolidated vadose zone. Also, post-depositional geochemical modification of the sediment does not significantly alter the compression behaviour of recently-formed sediments obtained from the depositional surface. Hence, hypothesis 6 cannot be rejected. However, the effect of humification in saltmarsh materials on voids ratio and compression behaviour remains a key unknown.

Each of the six research hypotheses have now been addressed following investigations into mechanical and (bio-)chemical autocompaction processes in intertidal vadose zone stratigraphies. It is now possible to develop an empirically-informed predictive autocompaction model. This is undertaken in Chapter 8.

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CHAPTER 7: DIAGENETIC PROCESSES IN THE OVERCONSOLIDATED VADOSE ZONE

This chapter presents results of an investigation into diagenetic processes in shallow stratigraphies and their effect on *in situ* voids ratio and compression behaviour. Diagenetic processes are known to operate in the vadose zone (Section 3.9.10) and it is necessary to identify such processes and investigate their influence. Near-surface sediments are likely to be overconsolidated and thus it is possible that the influence of effective stress is secondary to that of other diagenetic processes. Understanding the relative importance of these processes is therefore critical to the modelling approach.

7.1 DIAGENETIC PROCESSES IN SHALLOW MINERALOGENIC INTERTIDAL STRATIGRAPHIES

The first step in determining the influence of diagenetic processes on *in situ* voids ratios involved their identification in shallow, vadose zone stratigraphies. This was achieved through examination of a series of overlapping cores collected from the beneath the low marsh and mudflat (sampling altitudes 2.26 m OD and 1.06 m OD) according to the methods described in Section 4.2.2.

From the low marsh site, cores LMX-1 and LMX-3 extend from the surface to a depth of 0.17 m. Core LMX-2 samples depths of 0.16 - 0.33 m and core LMX-4 covers the vertical range of 0.21 – 0.39 m (the maximum depth sampled). The stratigraphy of the sampled sedimentary profile is illustrated and described in Figure 7.1. The profile consists of 5 primary stratigraphic units. Working from the base of the core upwards, the lowest unit (L1, 0.39 – 0.15 m) is a dark brown organic silt. The darker, mottled colour may reflect either the presence of insoluble iron-sulphides and reducing conditions (controlled by the depth of the water table beneath the low marsh surface; Cundy and Croudace, 1995b), humification of organic matter or a combination of the two. Differentiating between humification and sulphide enrichment is difficult by visual observation alone, since both phenomena tend to result in a darkening of the sediment. The overlying unit, L2 (0.15 – 0.13 m) has a similar lithological make-up but is mid-brown in colour, suggesting a lesser degree of humification and/or the presence of iron oxyhydroxides (Zwolsman *et al.*, 1993). There is a return to the darker, humified state in unit L3 (0.13 – 0.12 m), before the sediment once again displays a lighter brown colour and an apparently lesser degree of humification within unit L4 (0.12 – 0.07 m). At 0.07 m, there is a fairly diffuse boundary into a more organic unit (L5 a); this is the contemporary low marsh material (an organic silt) that extends upwards to a depth of 0.005 m. Extending downwards into unit L5a is L5b - a loosely bound organic silt that is matted by the surface vegetation. This unit was unsuitable for geotechnical testing and was previously disregarded in the geotechnical testing section (Chapter 6). All sediments above 0.12 m display red, brown and sometimes grey mottling (i.e. no distinct redox zonation), indicative of a fluctuating water table (Zwolsman *et al.*, 1993).

The mudflat cores obtained (five in total) extend to a greater depth (*c.* 0.48 m) beneath the contemporary mudflat sampling altitude. Cores MFX-1 and MFX-4 extend from the surface to a depth of 0.18 m. Cores MFX-2 and MFX-3 cover the range of depths from 0.12 – 0.31 m. Core MFX-5 extends from 0.31 – 0.48 m. The stratigraphy is displayed in Figure 7.2 and is less complex that that of the low marsh cores, with two main units present. The lowest stratum (unit M1) is a dark black silt with some sand and clay. The black colour of the material indicates an highly reduced sediment and the presence of iron-sulphides. It is probable that this section of the stratigraphic column remains fully saturated by capillary action. Desiccation in this sediment is also likely to be rare given the high (semidiurnal) frequency of flooding by tidal waters and the reduced duration of subaerial exposure (Section 5.5). This stratum also has occasional, localised sections of randomly oriented, and hence detrital, organic material. The overlying unit (M2, 0.03 – 0.00 m) is of the same lithology. However, the sediment is light brown (oxidised) and is characterised by the presence of bioturbation burrows. It is unlikely to be coincidence that the depth of the oxidised zone and the depth to which these burrows occur is the same; indeed, faunal bioturbation structures are likely to be responsible for the aeration and oxidation of this zone.

The visual description of the sample cores suggests the operation of two non-mechanical diagenetic processes in the shallow intertidal stratigraphies. Firstly, redox remobilisation and zonation is evident in both cores, most obviously in the mudflat samples. Secondly, there is evidence that humification has differentially operated at specific depths in the low marsh cores. This process is absent in the mineralogenic mudflat cores.

7.2 QUANTIFICATION OF VARIABLES

The next stage in determining the relative influence of different variables (effective stress, lithology, geochemical variations and degree of humification) on *in situ* voids ratio involves their quantitative measurement. For the majority of these variables, such quantification is straightforward. The relevant methods are well-established and outlined in Chapter 4. Humification, however, is difficult to quantify for the purposes of this investigation.

Analysis of peat humification has its origins in studies of ombrotrophic (i.e. deriving moisture and nutrients directly from rain water) peat bogs, which are used as potential archives of palaeoclimatic and palaeoenvironmental data (Caseldine *et al.*, 2000). This technique rests upon the assumption that plant decay (humification) is primarily determined by surface wetness and temperature at the time of peat deposition. The darkness and colour of the humic acids produced by the decomposition of organic matter are indicative of the degree of humification and hence decomposition and climate at the time of formation (Blackford and Chambers, 1993). Analyses have evolved from 'hands-on', and hence subjective, field determination methods. For example, the von Post scale of peat decomposition categorises different degrees of humification according to a ten point scale on the basis of soil texture and colour and the nature of the water squeezed from the peat. More recently, analysis typically involves examination of humic substances that are chemically extracted from the soil (Anderson, 1998; Borgmark, 2005; Borgmark and Schoning, 2006; Langdon and Barber, 2001). The most widely used of these more quantitative chemical methods is that proposed by Blackford and Chambers (1993). This involves the use of 8 % sodium hydroxide (NaOH) as the extractant, into which dried peat samples (typically 0.1 - 0.5 g) are placed and heated for an hour. Samples are then filtered and diluted to the required concentration, which depends on the mass of sample used in analysis. Measurement is then undertaken using UV/VIS spectrometry at a wavelength of 540 nm and expressed as percentage light transmission (or absorbance). Some researchers, however, convert the 'raw' transmission measurements to humification by relating them to humic acid standards (Caseldine *et al.*, 2000; Chambers *et al.*, 1997).

Humification analysis was attempted on the low marsh core, sampling at 0.01 m depth intervals and using the established Blackford and Chambers (1993) technique. Results were, however, inconclusive; downcore values of percentage light transmission remained largely constant at *c.* 92 – 93 %. This suggests that either there is no variation in humification throughout the core (which is in contrast to conclusions obtained from visual observation of the core), or that the technique is not suited to predominantly mineralogenic intertidal sediments. This may be because the Blackford and Chambers (1993) technique was developed for freshwater peats that are composed principally of decayed upland mosses, sedges and shrubs. The chemical and spectral properties of the humic acids of these peats are likely to differ from those obtained from intertidal halophytic vascular flora and so the choice of chemical extractant and wavelength is critical (Caseldine *et al.*, 2000). Indeed, Garcia *et al.* (1993) for example, illustrated that NaOH preferentially released high molecular weight humic fractions in *Sphagnum*-based but not *Carex*-based peats. Since even the descriptive von Post method was developed for 'pure' peats, assigning reliable, quantitative values of humification is extremely difficult in the low marsh sediments. In the absence of a suitable method, humification is not considered as a predictor variable. However, some implications of humification are readdressed later (Section 8.2.1).

For the remaining variables, their downcore variation is now discussed, starting with the dependent variable (voids ratio). Depth profiles of the independent (predictor) variables are described, along with any relationships that they have with voids ratio, as determined by visual observations and *via* Pearson's correlation tests (Tables 7.1 and 7.2). The foraminiferal biostratigraphy of the cores is also presented to determine the environment of formation of the deeper sections of the cores (saltmarsh or mudflat environments).

In terms of geochemistry, seven diagenetically sensitive substances have been selected on the basis of previous work into the redox remobilisation and zonation in near-surface intertidal sediments by Cundy and Croudace (1995b), Cundy and Croudace (1996), Thomson *et al.* (2002) and Cundy (*pers. comm.*). These are: silica (SiO_2), iron (III) oxide (Fe_2O_3), manganese monoxide (MnO), calcium oxide (CaO, a proxy for $CaCO_3$ – Thomson *et al.*, 2002), phosphorus pentoxide (P_2O_5), sulphur (S) and iodine (I). Geochemical analysis was undertaken at a sampling resolution of 0.02 m in both the low marsh and mudflat cores.

7.2.1 Voids ratio profile of the low marsh core

Downcore voids ratio profiles for the four low marsh cores, as derived from x-ray core scanning, are displayed in Figure 7.3. Figure 7.3 (a) displays the mean voids ratio value as well as a root squared error (±), which incorporates both the standard error of the mean and the systematic instrumental error (Section 4.3.4). It is clear that the associated errors are relatively small, other than the occasional larger error (at *c.* 0.3 m, for example) which is likely to represent a random, unavoidable surge in the electrical mains supply. Henceforth and for clarity, only the mean value of the upward and downward scans is considered (Figure 7.3 (b)).

Following x-ray scanning, cores LMX-1 and LMX-2 were extruded from the sampling tubes and sliced open to allow sampling for litho-, bio- and chemo-stratigraphic analysis. The remaining cores (LMX-3 and LMX-4) were carefully extruded and prepared for subsequent oedometer testing.

Cores LMX-1 and LMX-2 were also photographed. Figure 7.4 displays these photographs, overlain by the relevant voids ratio profiles. The visual stratigraphy is also displayed. It can be seen that each lithological unit has its own voids ratio signature. This confirms that voids ratios reflect lithology, indicating the reliability of the technique and its potential as a non-destructive method for sediment characterisation. However, there is considerable within-unit variation in voids ratio, presumably reflecting smaller-scale differences in depositional conditions, moisture content and stress history, as observed in the geotechnical samples at the depositional surface (Chapter 5).

In the thicker and deepest humified stratigraphic unit (L1; 0.39 – 0.15 m), voids ratios display an upward rising trend. At 0.39 m, values of *e* begin at *c.* 2.5 and rise to *c.* 4 at 0.15 m. Occasional fluctuations in voids ratio are evident throughout unit L1 that cannot be visually attributed to any variations in lithology. Voids ratios fall to *c.* 2.5 in unit L2 (clay-rich silt with organic matter, 0.15 – 0.13 m) before rising again to *c.* 4 in unit L3 (humified organic silt; 0.13 – 0.12 m). In the overlying unit (L4; clay-rich silt with some organic matter, 0.12 – 0.07 m) voids ratios decrease and fluctuate between 2 and 3.5.

At the boundary to the uppermost unit (L5a; organic silt, *c.* 0.07 – 0.00 m), values of e rise to fluctuate between 3.5 and 6.5. This rise is more pronounced in core LMX-1, perhaps indicating a sharper stratigraphic boundary. Such rapid lateral and stratigraphic variations in voids ratio corroborate the previous suggestion (Section 6.3.5) that rapid spatial variations in initial voids ratio at the depositional surface translate into lateral and stratigraphic voids ratio variations. In core LMX-3, there is an obvious decreasing trend in voids ratio, from *c.* 5.25 at 0.07 m to *c.* 3.5 at 0.025 m. This trend then reverses, again increasing to a value of *c.* 5.25 at the low marsh surface. A similar, although less pronounced, trend is evident within unit L5a in core LMX-1. Values of *e* are generally higher in this core sample, and reach their maximum (*c.* 6.3) at *c.* 0.04 m.

7.2.2 Effective stress profile of the low marsh core

The effective stress profile of the low marsh core is displayed in Figure 7.5. In general, effective stress increases linearly with depth. However, variations in this trend occur and these can be attributed to lithological, and hence density, variations. The most obvious of these variations in gradient occurs at a depth of *c.* 0.07 m, which corresponds with the transition from unit L5a to L4.

7.2.3 Biostratigraphy of the low marsh core

The low marsh cores were sampled for foraminifera at a minimum resolution of 0.02 m. Foraminiferal assemblages (as a percentage of the total counted at each sampling depth) and total counts are displayed in Figure 7.6 in relation to depth, voids ratio profiles and visual stratigraphy. Cluster analysis was also performed on the foraminiferal assemblages using CONISS (Section 5.1.4; Grimm, 1987; 1993). Stratigraphic constraint was used during analysis, meaning assigned clusters must consist of stratigraphically contiguous samples.

Foraminifera counts are generally greater than 100, allowing reliable palaeoenvironmental reconstruction. However, between 0.08 and 0.14 m, counts do drop below 100. It is general practice to exclude such counts from palaeoenvironmental analysis; however, the monospecific assemblage (*Jadammina macrescens*) at these depths increases statistical confidence.

Cluster analysis split the samples into four foraminiferal zones. Zone LF1 (0.39 – 0.32 m) consists almost entirely of *Jadammina macrescens* (80 – 100%) with occasional *Trochammina inflata* (maximum of 20 %) and *Haplophragmoides* spp. (< 10 %).

In zone LF2 (0.30 – 0.16 m), *Miliammina fusca* becomes increasingly dominant between 0.30 and 0.16 m depth, constituting 20 – 40 % of the total foraminifera counted; this is accompanied by a commensurate decrease in the relative abundance of *Jadammina macrescens*. *Trochammina inflata* (< 10 %) and *Haplophragmoides* spp. (< 10 %) are also found within this zone.

Zone LF3 (0.14 – 0.04 m) is characterised by low species diversity (i.e. low number of species), again with *Jadammina macrescens* being dominant. The assemblage is monospecific between 0.12 and 0.10 m depth, and very nearly so at 0.08 m depth where *Miliammina fusca* constitutes < 5 % of total foraminifera.

In zone LF4 (0.07 – 0.00 m), species diversity increases and the dominance of *Jadammina macrescens* accordingly decreases. Particularly in the uppermost 0.04 m, *Trochammina inflata* becomes a more significant species, and even becoming dominant at 0.02 m.

All foraminifera present are agglutinated species, indicating that the entire core section accumulated within a saltmarsh environment (Horton, 1997; Horton, 1999; Horton and Edwards, 2000). Furthermore, the low species diversity is also typical of saltmarsh foraminiferal assemblages (in comparison with calcareous mudflat assemblages).

As well as being related to the visual stratigraphic units, the foraminiferal assemblages are also related to the voids ratio profile. Most noticeably, the highest (near-surface) voids ratios at the top of the core correspond with assemblages dominated by *Trochammina inflata*. Additionally, the lowest voids ratios (0.07 – 0.12 m depth) correspond with the monospecific *Jadammina macrescens* foraminiferal assemblage. Since the distribution of foraminifera in intertidal environments is a direct function of altitude (a surrogate for flooding duration and frequency) (Horton *et al.*, 1999), this provides strong additional evidence that the voids ratio of a material is related to the position at which it formed within the intertidal frame.

7.2.4 Lithostratigraphy of the low marsh core

Lithostratigraphic analyses are based on particle size and loss on ignition measurements, sampling at 0.01 m resolution. These variables, plotted in relation to depth in the core, are displayed in Figure 7.7 with the voids ratio profiles and the visual stratigraphy of the low marsh cores.

It can be seen that particle size and organic content variation are directly related to the stratigraphic units. The uppermost stratum (L5a; 0.00 – 0.07 m) is characterised by the highest organic contents. Unit L4 (0.07 - 0.12 m) has a lower organic content (*c.* 15 %). Variations in particle size also correspond with stratigraphic layers; shifts in mean particle size and the range of particle size distribution are accompanied by visual changes in lithology.

The loss on ignition profile is strikingly similar to the voids ratio profile. The highest voids ratios (*c.* 4 – 6) correspond to the maximum organic contents (20 – 30 %) in unit L5a (0.00 – 0.07 m). In the less organic layers (particularly L4; 0.08 – 0.12 m), loss on ignition values fluctuate around 15 %. This corresponds to voids ratios of *c.* 2 – 3. This strongly suggests that organic content exerts a significant control on voids ratio. Indeed, the two variables are strongly correlated ($r = 0.85$, $p < 0.001$) (Table 7.1).

There is also a less pronounced relationship between particle size and voids ratio. The deepest stratigraphic unit (clay-rich silt with some well-humified organic matter, 0.17 – 0.38 m depth) has the finest particles (modal particle diameter of 8 – 9 μm; a medium silt (Blott and Pye, 2001)). Particle size is fairly constant within this stratigraphic unit, although it displays a subtle coarsening-upwards trend that may be linked to the observed variations in voids ratios with depth. In the overlying strata typified by lower organic contents and lower voids ratios (0.07 – 0.12 m depth), the grain size distributions coarsen to a range of approximately 8 – 800 μm, before fining again in the uppermost (surface) stratigraphic unit. Here, the model grain size is *c.* 10 – 30 μm and the overall range is in the region of 4 – 300 μm. Throughout the sequence, silt is inversely and weakly (r = -0.367), yet significantly (p = 0.03), correlated with voids ratio. Clay is positively correlated with voids ratio, although only weakly ($r = 0.353$, $p = 0.04$) (Table 7.1).

7.2.5 Chemostratigraphy of the low marsh core

The stratigraphic distribution of redox sensitive elements is displayed in Figure 7.8. Stratigraphically constrained cluster analysis was performed on the geochemical data using CONISS (Grimm, 1987; 1993) and the low marsh core displays definite chemostratigraphic zonation. Five chemozones were assigned; these broadly agree with the visual (stratigraphic) observations. For example, chemozones LC1and LC2 correspond with lithological unit L1. Similarly, chemozones LC3 and LC4 overlap with lithological units L2 – L4. Chemozones LC4 and LC5 occupy the same depth range as lithological unit L5a.

Geochemical variations in chemozones LC1 (0.34 – 0.24 m) and LC2 (0.22 – 0.16 m) are generally linear, with both increasing and decreasing trends being displayed. For example, SiO_2 displays a gradual yet consistent falling trend from the base of the core upwards, from *c.* 46 wt % to *c.* 42 wt % at 0.16 m. In chemozones LC1 and LC2, a

Table 7.1 Correlations between geotechnical, lithological and geochemical variables for the low marsh core.

		Voids ratio, *e*	$\log_{10}\sigma'$ (kPa)	Loss on ignition (%)	Sand (%)	Silt (%)	Clay (%)	SiO_2 (% wt)	Fe_2O_3 (% wt)	MnO (% wt)	CaO (% wt)	P_2O_5 (% wt)	S (ppm)
$\log_{10}\sigma'$ (kPa)	*r*	**-0.757**											
	p	**<0.001**											
Loss on ignition (%)	*r*	**0.853**	**-0.605**										
	p	**<0.001**	**<0.001**										
Sand (%)	*r*	-0.021	0.082	0.095									
	p	0.905	0.639	0.586									
Silt (%)	*r*	**-0.367**	**0.717**	-0.182	**-0.350**								
	p	**0.030**	**<0.001**	0.294	**0.039**								
Clay (%)	*r*	**0.353**	**-0.725**	0.087	**-0.515**	**-0.623**							
	p	**0.037**	**<0.001**	0.618	**0.002**	**<0.001**							
SiO_2 (% wt)	*r*	**-0.507**	0.230	-0.466	**0.516**	-0.232	-0.307						
	p	**0.032**	0.358	0.051	**0.028**	0.354	0.215						
Fe_2O_3 (% wt)	*r*	-0.044	0.044	-0.162	**-0.737**	0.327	0.442	**-0.659**					
	p	0.864	0.862	0.521	**<0.001**	0.185	0.066	**0.003**					
MnO (% wt)	*r*	**0.668**	-0.450	**0.611**	0.347	-0.415	0.039	-0.402	0.099				
	p	**0.002**	0.061	**0.007**	0.159	0.087	0.879	0.098	0.697				
CaO (% wt)	*r*	**-0.562**	**0.810**	-0.334	0.257	**0.496**	**-0.730**	0.421	-0.342	-0.316			
	p	**0.015**	**<0.001**	0.175	0.304	**0.036**	**0.001**	0.082	0.165	0.202			
P_2O_5 (% wt)	*r*	**0.890**	**-0.629**	**0.810**	0.111	-0.355	0.221	**-0.585**	0.090	**0.636**	**-0.497**		
	p	**<0.001**	**0.005**	**<0.001**	0.660	0.149	0.378	**0.011**	0.723	**0.005**	**0.036**		
S (ppm)	*r*	**0.687**	-0.253	**0.817**	0.379	0.006	-0.392	-0.457	-0.223	**0.486**	-0.115	**0.682**	
	p	**0.002**	0.311	**<0.001**	0.121	0.981	0.107	0.056	0.373	**0.041**	0.649	**0.002**	
I (ppm)	*r*	0.387	0.121	**0.521**	0.080	0.295	-0.360	**-0.666**	0.333	**0.553**	0.008	**0.551**	**0.718**
	p	0.112	0.633	**0.027**	0.752	0.234	0.142	**0.003**	0.177	**0.017**	0.976	**0.018**	**0.001**

r = Pearson's correlation coefficient. *p* = significance. Significant correlations are in bold type.

Table 7.2 Correlations between geotechnical, lithological and geochemical variables for the mudflat core.

		Voids ratio, e	$\log_{10}\sigma'$ (kPa)	Loss on ignition (%)	Sand (%)	Silt (%)	Clay (%)	SiO_2 (% wt)	Fe_2O_3 (% wt)	MnO (% wt)	CaO (% wt)	P_2O_5 (% wt)	S (ppm)
$\log_{10}\sigma'$ (kPa)	r	**-0.694**											
	p	**<0.001**											
Loss on ignition (%)	r	**0.440**	-0.270										
	p	**0.002**	0.060										
Sand (%)	r	-0.235	**0.657**	**0.293**									
	p	0.105	**<0.001**	**0.041**									
Silt (%)	r	**-0.668**	**0.608**	**-0.354**	-0.021								
	p	**<0.001**	**<0.001**	**0.013**	0.886								
Clay (%)	r	**0.503**	**-0.861**	-0.113	**-0.901**	**-0.414**							
	p	**<0.001**	**<0.001**	0.438	**<0.001**	**0.003**							
SiO_2 (% wt)	r	-0.050	0.196	**-0.509**	0.027	0.122	-0.075						
	p	0.811	0.349	**0.009**	0.897	0.560	0.721						
Fe_2O_3 (% wt)	r	**0.721**	**-0.751**	**0.667**	-0.223	**-0.628**	**0.464**	**-0.526**					
	p	**<0.001**	**<0.001**	**<0.001**	0.284	**0.001**	**0.019**	**0.007**					
MnO (% wt)	r	**0.531**	-0.321	-0.064	-0.129	**-0.429**	0.295	**0.561**	0.197				
	p	**0.006**	0.117	0.762	0.538	**0.032**	0.152	**0.003**	0.344				
CaO (% wt)	r	-0.213	0.373	**0.518**	0.396	0.129	**-0.430**	-0.256	-0.073	-0.220			
	p	0.307	0.066	**0.008**	0.050	0.538	**0.032**	0.217	0.729	0.292			
P_2O_5 (% wt)	r	**0.561**	**-0.843**	0.031	**-0.623**	**-0.513**	**0.802**	0.015	**0.584**	**0.486**	**-0.553**		
	p	**0.004**	**<0.001**	0.884	**0.001**	**0.009**	**<0.001**	0.945	**0.002**	**0.014**	**0.004**		
S (ppm)	r	0.294	-0.214	**0.789**	0.146	-0.303	-0.019	**-0.802**	**0.663**	-0.335	**0.514**	-0.110	
	p	0.153	0.305	**<0.001**	0.486	0.141	0.928	**<0.001**	**<0.001**	0.101	**0.009**	0.599	
I (ppm)	r	-0.237	**0.639**	0.340	**0.806**	0.063	**-0.797**	-0.068	-0.132	-0.011	0.459	**-0.465**	0.176
	p	0.255	**0.001**	0.096	**<0.001**	0.766	**<0.001**	0.748	0.529	0.958	0.021	**0.019**	0.400

r = Pearson's correlation coefficient. p = significance. Significant correlations are in bold type.

gradual upward falling trend in CaO concentration can be observed (maximum of *c.* 2.75 wt % at 0.36 m). An upward rising trend in P_2O_5 is evident within chemozones LC1 and LC2, although only between concentrations of 0.1 – 0.3 wt %.

S and I have increased concentrations in chemozones LC1 and LC2 relative to the overlying chemozone, LC3 (0.14 – 0.08 m). In LC1 and LC2, I shows no persistent increasing or decreasing trends in concentrations, remaining between *c.* 75 and 125 ppm. S begins to decrease following its peak (*c.* 5500 ppm) at 0.16 m. The darker colour of lithological unit L1 does indeed seem to be related to the increase in S, reflecting the bacterial reduction of sulphate and the presence of black iron sulphides (Spencer *et al.*, 2003). However, the partly mottled nature of this stratum indicates that a stable redox zonation has not developed at this depth as a result of the fluctuating (over weekly to annual timescales) water table. The dynamic water table causes short-term fluctuations in the redox boundary (Casey and Lasaga, 1987).

Chemozone LC4 displays enrichment in MnO, P_2O_5, S and I in comparison to the underlying geochemical unit, LC3, and the overlying LC5. LC4 also displays relative depletion of SiO_2. Other than depletion in zone LC5 (values of *c.* 3.5 wt %), Fe_2O_3 displays little variation with stratigraphy or chemozone. Concentrations fluctuate around these values until a depth of *c.* 0.28 m, before dropping to 5 -6 wt % within chemozone LC1.

CaO concentrations largely correspond with the visual stratigraphy, falling from *c.* 1.75 wt % at the surface (top of chemozone LC5) to *c.* 0.5 wt % at 0.10 m (LC3). This is followed by a more rapid increase in concentration to levels similar to those observed at the surface in the lithological unit L2 (0.13 - 0.15 m depth).

MnO concentrations are generally constant downcore, fluctuating between 0.0 and 0.1 wt %, other than a peak at 0.04 m (0.45 wt %) in chemozone LC4. Similarly, P_2O_5 concentrations range from just *c.* 0.1 to 0.6 % by weight, and I levels never exceed 225 ppm. Variations in the geochemical profile are instead dominated by SiO_2 (*c.* 35 -55 wt %) and to a lesser extent by variations in Fe_2O_3 (3 – 7 wt %), S (2000 – 6500 ppm) and CaO (0 – 3 wt %).

SiO_2 and CaO are inversely correlated with the voids ratio variations (Table 7.1). In contrast, P_2O_5 and S show direct relationships with voids ratio. The correlation between

P_2O_5 and voids ratio in particular is very strong ($r = 0.890$, $p < 0.001$). Interestingly, I is not significantly correlated with voids ratio ($r = 0.387$, $p = 0.112$) despite displaying similarities in the form of the respective depth profiles. In contrast, MnO correlates fairly strongly and significantly ($r = 0.668$, $p = 0.002$) with voids ratio; such a relationship is perhaps not evident in the depth profiles (Figure 7.8).

7.2.6 Voids ratio profile of the mudflat core

The voids ratio profiles of cores MFX 1 -5 are displayed together in Figure 7.9. As in the low marsh cores, the root squared error (as displayed in Figure 7.9 a) is small and so only the mean value of the upward and downward scans is considered (Figure 7.9 b).

After the x-ray scanning procedure, cores MFX-1 and MFX-2 were extruded, horizontally split and photographed. Cores MFX-3, MFX-4 and MFX-5 were extruded with care to prevent disturbance and prepared for oedometer testing. Cores were also sampled for litho-, bio- and chemo-stratigraphic analysis.

Figure 7.10 displays photographs of the cores, overlain by the voids ratio profiles at each depth and the visual stratigraphy. There is substantial lateral and vertical variation in voids ratios within lithological unit M1, both within and between cores. These variations do not appear to be linked to any changes in lithology. There is an offset between cores obtained from the same sample depths (i.e. between cores MFX-1 and MFX-4; and between MFX-2 and MFX-3) that is not attributable to variations in the electrical mains supply between different scans, since each core scanning session was individually calibrated. The voids ratio profiles are also characterised by occasional 'upward' and 'downward' peaks at, for example *c.* 0.05 m (MFX-1), *c.* 0.10 – 0.12 m (MFX-1 and MFX-4) and at *c.* 0.14 m (MFX-4) (Figure 7.9 b). The low standard errors associated with the voids ratio profiles displayed in Figure 7.9 (a) indicate that these peaks are indeed 'real' and are not an artefact of sampling error. However, given the visual uniformity of lithology downcore, they cannot be directly linked to variations in grain size or organic content and may reflect the range of depositional structures caused by variations in depositional and/or subaerial conditions. An interesting feature of core MFX-5 is the decrease in voids ratio from *c.* 1.9 to *c.* 1.7 between 0.36 and 0.40 m, and the subsequent increase back to *c.* 2 at 0.46 m. Again, this feature cannot be attributed to any obvious lithological variation.

In the uppermost (oxidised) stratum (M2; 0.03 – 0.00 m), voids ratios increase rapidly from *c.* 2.0 to 2.8, reflecting the more open structures associated with newly deposited sediment, variations in the presence of bioturbation structures and the higher occurrence of these structures towards the depositional surface.

7.2.7 Effective stress profile of the mudflat core

Despite minor variations in gradient that are attributable to small structural (density) fluctuations between cores, it is evident from Figure 7.11 that effective stress increases linearly downcore, reaching a maximum of *c.* 2.55 at the base of the core. This linear increase of effective stress with depth suggests downcore lithological and geotechnical homogeneity (overconsolidation) within of the mudflat core.

7.2.8 Biostratigraphy of the mudflat core

Samples were taken from the mudflat cores generally at a resolution of 0.04 m. Foraminiferal assemblages (as a percentage of the total counted at each sampling depth) and total counts are displayed in Figure 7.12 in relation to depth, voids ratio profile and visual stratigraphy. Species constituting less than 2 % of the total foraminifera counted are excluded from the diagrams for clarity.

The mudflat core is characterised by greater species diversity than the low marsh core. The majority of foraminifera (always greater than 90 %, and generally greater than 95 %) are calcareous, confirming that the core was formed in a tidal flat environment (Horton, 1997; Horton, 1999; Horton and Edwards, 2000). This is consistent with the lithological characteristics of the core. The remaining 5 -10 % of foraminifera (*Jadammina macrescens* and *Miliammina fusca*) are agglutinated and are likely to be allochthonous. Total counts are higher towards the base of the core, reaching 500 at 0.44 m. Counts drop below 200 above *c.* 0.15 m, and below 100 at *c.* 0.12 m. Environmental interpretations from these low (< 100) counts must be treated with caution.

Stratigraphically constrained cluster analysis using CONISS produced four foraminiferal zones. In all zones, foraminiferal assemblages are dominated by *Haynesina germanica*, which commonly constitutes > 50 % of the total. *Stainforthia fusiformis* is the second most abundant species and generally contributes between 20 and 30 % to the total foraminifera counted.

In the deepest zone (MF1; 0.44 – 0.36 m), the relative abundance of *Haynesina germanica* is at its lowest, falling below 50 %. There is a commensurate increase in the proportion of *Elphidium magellanicum* (*c.* 5 – 10 %) in this zone. The proportion of *Haynesina germanica* increases to *c.* 55 – 70 % in zone MF2.

Maximum counts of *Stainforthia fusiformis* relative to other species (> 25 %) are found in zone MF3 (0.22 – 0.08 m). There is a corresponding fall in the proportion of *Haynesina germanica* (*c.* 40 – 55 %). In zone MF4 (0.04 – 0.00 m), *Haynesina germanica* again increases in dominance to *c.* 75 %. This results in a fall in the proportion of *Stainforthia fusiformis* (< 20 % in MF4).

The remaining species in the core make only minor contributions. *Quinqueloculina* spp. and *Cibicides lobatulus* are the most significant of the minor species in zones MF3 and MF4.

There seems to be very little relationship between foraminiferal assemblage and the voids ratio profile. However, a subtle direct relationship may exist between variations in the abundance of *Haynesina germanica* and voids ratio, particularly at depths between 0.15 and 0.25 m. This may suggest an influence of relative elevation on the voids ratio profile, possibly caused by initial structural variations.

7.2.9 Lithostratigraphy of the mudflat core

Particle size and loss on ignition variation, sampled at 0.01 m resolution, are displayed in Figure 7.13 in relation to depth, the voids ratio profile and the visual stratigraphy of the mudflat cores.

Downcore particle size variations reveal three main units. The top unit, coinciding with the oxidised layer (0.00 – 0.03 m) is the coarsest section of the core. Particle size generally ranges from *c.* 8 – 600 μm, with a modal diameter of *c.* 50 – 80 μm (very fine sand). Between 0.03 and *c.* 0.18 m, the particle size distribution becomes finer, ranging from *c.* 4 – 300 μm, with a modal diameter of 8 – 10 μm (medium silt). At greater depths (> 0.18 m), particle size fines again, ranging from 0.9 to 100 μm, with a model grain size of 7 – 8 μm (fine/medium silt). This coarsening upwards sequence is indicative of increasing energy conditions through time, suggesting an increase in tidal water depths (positive sea level tendency). These subtle variations in particle size do not appear to be linked to the voids ratio profile.

The loss on ignition profile varies considerably, ranging from values typical of the contemporary mudflat (Section 5.1.2) (*c.* 12 %) at the surface of the core to a maximum of *c.* 19 % at 0.25 m depth. These values then fall to *c.* 13 % towards the base of the core (> 0.40 m) and remain largely constant. The higher loss on ignition values in the core are similar to those found in the contemporary low marsh sub-environment at Greatham Creek. From the visual description of the lithology and from the foraminifera present within the core (Section 7.3.5), it is evident that the material was not formed in a saltmarsh environment. Changes in loss on ignition values are likely to reflect variations in detrital organic matter in the core that has been transported onto the mudflat during times of high productivity, or maybe erosion, in the saltmarsh. Alternatively, the higher loss on ignition values may result from the combustion of chemical elements, such as sulphur or carbonate.

The considerable variations in particle size and loss on ignition are not reflected in the voids ratio profile, at least by visual observation of the data alone. The quantitative lithological data displayed in Figure 7.13 also suggest that the voids ratio profile of the core does not seem to be related to its lithology. Surprisingly, therefore, when the particle size distributions are split into sand, silt and clay fractions, voids ratio significantly correlates with silt (r = -0.668, p < 0.001) and clay (r = 0.503, p < 0.001) (Table 7.3). Variations in loss on ignition also correlate with those in voids ratio, albeit weakly (r = 0.440, p = 0.002).

7.2.10 Chemostratigraphy of the mudflat core

The depth distributions of SiO_2, Fe_2O_3, MnO, CaO, S and I are displayed in Figure 7.14 in relation to depth, the voids ratio profile and the visual stratigraphy of the mudflat core. The redox zonation of the core, as noted previously, is reflected in the geochemistry of the core. In particular, S shows depletion (*c.* 4500 – 7000 ppm) in the light brown oxic surface layer (chemozone MC3); 0.00 – 0.03 m) and enrichment (*c.* 10500 ppm) in the black anoxic zone (chemozones MC1 and MC2; > 0.03 m depth). This confirms previous speculations that sediment beneath the 0.03 m redox boundary remains permanently saturated, probably by capillary action, and is never exposed to oxygen. CaO also shows depletion in the oxic layer (chemozone MC3) and this may also reflect the stong redox zonation of the core. However, neither S (r = 0.294, p = 0.153) nor CaO (r = -0.213, p =

0.307) (Table 7.2) is strongly nor significantly correlated with voids ratio, suggesting the inferred redox mobilisation of these compounds has little effect on *in situ* voids ratio.

Other compounds display variation above and below the redox boundary; SiO_2, MnO and, to a lesser extent, P_2O_5 display enrichment in chemozone MC3. This is initially followed by sudden decreases in concentrations of this substance in the anoxic chemozone MC2. However, redox zonation cannot account for the fluctuations in the concentrations of these substances below the redox boundary within chemozone MC1, particularly the distributions of SiO_s and MnO. Indeed, SiO_2 concentrations do not appear to be related to the voids ratio profile, neither visually nor statistically (r = -0.050, p = 0.811), although the increase and subsequent decrease in SiO_2 between 0.35 and 0.48 m in MC1 is broadly coincident with the anomalous increase/decrease trend in the voids ratio profile at the same depths, noted previously. MnO also varies proportionally with the voids ratio profile at these depths. Indeed, MnO concentrations visually mimic the voids ratio variations downcore and the two variables correlate significantly, if not particularly strongly (r = 0.531, p = 0.006). The strongest correlation, however, is between Fe_2O_3 and voids ratio (r = 0.721, p < 0.001). The general decreasing trend in voids ratio downcore is also visible in the Fe_2O_2 profile, particularly in chemozone MC1.

7.3 MULTIPLE REGRESSION ANALYSIS OF VARIABLES AFFECTING IN SITU VOIDS RATIOS

On the basis of visual examination of the data and correlation coefficients (Tables 7.1 and 7.2), it is evident that relationships exist between the lithological and geochemical variables and the *in situ* voids ratios in the cores. However, observations and simple correlations do not reveal the relative contributions of different predictor variables on voids ratios. This can be undertaken by employing multiple regression techniques and considering the standardised beta coefficient, which is a measure of how strongly each predictor (independent) variable influences the dependent variable (voids ratio, e, in this case). It is measured in units of standard deviation; for example, a beta value of 2 indicates that a change of one standard deviation in the predictor variable will result in a change of 2 standard deviations in the dependent variable. Hence, the higher the standardised beta coefficient, the greater the impact of the predictor variable on the dependent variable. The standardised beta coefficient allows comparisons to be made regarding the strength of the relationship of each predictor variable to the dependent variable.

In the multiple regression analysis below, the adjusted *r*-square (r^2_{adj}) value is used, since this takes into account the number of independent variables in the model and the number of observations upon which the model is based. Conventional r^2 values tend to overestimate the predictive capacity of a model, and will always be increased by adding predictor variables to the model. r^2_{adj} is therefore considered to be a more useful measure of the relative 'success' of different regression models.

Mean values of the dependent variable, voids ratio (*e*), at 0.01 m resolution were taken to remove the effect of the observed rapid and often large variations therein. Similarly, mean values of effective stress were taken at a similar sampling interval to account for minor, inter-core variations in the effective stress profile.

7.3.1 Multiple regression analysis of the low marsh core

Initially, all independent variables (effective stress, and lithological and geochemical parameters) were simultaneously entered into a multiple regression model. An overall statistically significant model emerged ($F_{11, 6}$ = 17.897, p = 0.001) with a very high predictive capacity (r^2_{adj} = 0.92). Predictor variables are displayed in Table 7.3 (along with their significance), ranked according to their relative importance in predicting the voids ratio, *e*, as indicated by their respective standardised beta coefficients.

The most important predictor variable is S (ppm). This is followed by silt content. I (ppm), loss on ignition and sand content have similar relative predictive power. Interestingly, effective stress ($\log_{10}\sigma'$ (kPa)) ranks lowly, suggesting it has a relatively minor role in predicting voids ratios in the overconsolidated vadose zone. However, the validity of these rankings can be called into question because only two predictor variables (S and silt) are statistically significantly linearly related to voids ratio at the 0.05 significance level.

Regression based only on effective stress ($\log_{10}\sigma'$, kPa) still produces an overall statistically significant model ($F_{1, 33}$ = 44.40, p < 0.001). In addition, effective stress is a statistically significant predictor variable (t = -6.664, p < 0.001). However, the predictive capacity of the model is reduced (r^2_{adj} = 0.56).

If the lithological variables (loss on ignition, sand, silt and clay contents) are entered into the model with effective stress, the model remains significant ($F_{4, 30}$ = 34.57, p < 0.001) and increases substantially in predictive power (r^2_{adj} = 0.80). However, a regression model based only on lithology performs equally well ((r^2_{adj} = 0.79) whilst remaining significant ($F_{4, 30}$ = 34.57, $p < 0.001$). This suggests that effective stress is not a dominant control on *in situ* voids ratio and that material/lithological factors are playing a more important role.

When geochemical variables are used as predictors, an r^2_{adj} of 0.87 is obtained ($F_{7, 10}$ = 17.28, $p < 0.001$). This performance of the model does not improve when effective stress is included as a predictor variable (r^2_{adj} = 0.86; $F_{8, 9}$ = 13.81, $p < 0.001$).

Table 7.3 Multiple regression parameters and the significance of predictor variables obtained from multiple regression analysis of the low marsh sediment core. The dependent variable is voids ratio, *e*. Variables are ranked according to their relative importance in prediction of *e*, as indicated by standardised beta coefficients. Significant predictor variables are shown in bold.

	Unstandardised beta coefficient	Standardised beta coefficient	*t* statistic	*p*
Intercept (constant)	4.54	-	0.97	0.368
S (ppm)	**0.001**	**1.50**	**3.12**	**0.021**
Silt (%)	**-0.10**	**-0.81**	**-2.44**	**0.050**
I (ppm)	-0.01	-0.567	-0.57	0.352
Loss on ignition (%)	-0.12	-0.56	-0.56	0.067
Sand (%)	-0.06	-0.50	-0.50	0.251
CaO (wt %)	0.51	0.42	0.42	0.145
Fe_2O_3 (wt %)	0.48	0.41	0.41	0.275
P_2O_5 (wt %)	2.69	0.38	0.38	0.130
MnO (wt %)	2.234	0.26	0.26	0.131
$\log_{10}\sigma'$ (kPa)	-0.20	-0.09	-0.09	0.872
SiO_2 (wt %)	0.01	0.05	0.05	0.889

It is clear that the most powerful predictive model was created when all variables were used in the regression analysis (r^2_{adj} = 0.92). This suggests that all variables need to be considered in analysis of voids ratio variation in these overconsolidated vadose zone sediments. However, many of the predictor variables used were statistically insignificant, with the t statistics not exceeding the critical value at either the 0.01 or 0.05 significance levels. In this case, the t statistic is being used to test the hypothesis that each predictor variable is linearly related to the dependent variable. Furthermore, such a model, with a large number of predictor variables, may not provide optimum parsimony.

In order to overcome these problems, backward elimination multiple regression was employed. This stepwise regression procedure begins by regressing the dependent variable on all of the predictor variables. If any variables are statistically insignificant, the variable making the smallest contribution to predictive capacity (the variable with the lowest partial correlation with the dependent variable) is considered for removal. The partial correlation measures the strength of the association between the response variable and each of the predictor variables not currently in the equation after removing the effect of the variables currently in the equation. Predictors variables are only removed if the significance of the change to the F statistic exceeds 0.10 (the removal criterion). Otherwise, it remains in the model. This variable elimination procedure continues until there are no variables in the equation that exceed the removal criterion.

For the low marsh core, using the backward elimination multiple regression method, a highly significant model ($F_{9, 8}$ = 18.264, p < 0.001) with excellent predictive capacity (r^2_{adj} = 0.937) emerged. Indeed, this r^2_{adj} is greater than that obtained from all variables. The significant predictor variables are displayed in Table 7.4.

Perhaps most significantly, effective stress has been excluded from the model; its influence is not sufficiently significant to warrant its inclusion. According to the model, the most important predictor variable is S (ppm); its influence in predicting voids ratio is almost twice as important of that of the variable ranked second in order of importance in prediction (silt).

7.3.2 Multiple regression analysis of the mudflat core

Simultaneous entry of all dependent variables produced a significant model ($F_{11, 13}$ = 7.383, p = 0.001) with good predictive capacity (r^2_{adj} = 0.75). Predictor variables are displayed in Table 7.5 (along with their significance), ranked according to their relative importance in predicting the voids ratio, e, as indicated by their respective standardised beta coefficients. Loss on ignition is the most influential predictor variables, followed by CaO and MnO. As in the low marsh analysis, effective stress has a relatively minor role in predicting the voids ratio in the overconsolidated vadose zone. Additionally, only two predictor variables are statistically significant (loss on ignition and CaO), although MnO borders on statistical significance at the 0.05 significance level.

Using only effective stress ($\log_{10}\sigma'$, kPa) as a predictor variable, a significant model emerges ($F_{1, 47}$ = 43.63, $p < 0.001$). The predictive capacity of the model, however, is reduced (r^2_{adj} = 0.47). By adding the lithological variables into the model, statistical significance remains ($F_{4,44}$ = 17.60, $p < 0.001$) and the r^2_{adj} value increases to 0.58. However, when effective stress is not entered into the model and only the lithological predictor variables remain, the model performs no worse (r^2_{adj} = 0.58; $F_{3,45}$ = 22.814, $p <$ 0.001).

Table 7.4 Multiple regression parameters and the significance of predictor variables obtained from backward elimination stepwise multiple regression analysis of the low marsh sediment core. The dependent variable is voids ratio, *e*. Variables are ranked according to their relative importance in prediction of *e*, as indicated by standardised beta coefficients. Two predictor variables, I (ppm) and Fe_2O_3, are statistically insignificant (shown in bold) but remain in the model due to their significance to overall model performance.

	Unstandardised beta coefficient	Standardised beta coefficient	*t* statistic	*p*
Intercept (constant)	4.87	-	2.10	0.690
S (ppm)	0.001	1.51	5.07	0.001
Silt (%)	-0.10	-0.81	-3.45	0.009
I (ppm)	**-0.01**	**-0.66**	**-2.97**	**0.18**
Loss on ignition (%)	-0.10	-0.54	-2.88	0.020
Sand (%)	-0.05	-0.45	-2.33	0.048
Fe_2O_3 (wt %)	**0.51**	**0.43**	**2.11**	**0.067**
P_2O_5 (wt %)	2.89	0.41	2.81	0.023
CaO (wt %)	0.47	0.39	2.48	0.038
MnO (wt %)	2.28	0.27	2.47	0.039

Similarly, there is little difference in the predictive capacity of regression models based upon geochemistry alone (r^2_{adj} = 0.63; $F_{7, 17}$ = 6.947, p = 0.001), and geochemistry and effective stress in combination (r^2_{adj} = 0.65; ($F_{8, 16}$ = 6.487, p = 0.001). In the latter model, effective stress possesses the highest standardised beta coefficient, indicating its importance as a predictor variable.

Since the most powerful model is produced by using **all** variables simultaneously, it is clear that effective stress is not the only control on *in situ* voids ratios and the predictive capacity is increased by incorporating lithological and diagenetic variables.

Using the more statistically rigorous backward elimination method, a highly significant model ($F_{5,19}$ = 18.264, p < 0.001) with good predictive capacity (r^2_{adj} = 0.78) emerged. Indeed, with fewer predictor variables, the predictive capacity of the model is greater than that of the simultaneous entry method. Predictor variables are displayed in Table 7.6.

Table 7.5 Multiple regression parameters and the significance of predictor variables obtained from multiple regression analysis of the mudflat sediment core. The dependent variable is voids ratio, *e*. Variables are ranked according to their relative importance in prediction of *e*, as indicated by standardised beta coefficients. Significant predictor variables are shown in bold.

	Unstandardised beta coefficient	Standardised beta coefficient	*t* statistic	*p*
Intercept (constant)	1.27	-	0.76	0.463
Loss on ignition (%)	**0.05**	**1.00**	**2.33**	**0.036**
CaO (wt %)	**-0.20**	**-0.79**	**-2.44**	**0.030**
MnO (wt %)	36.82	0.70	2.15	0.051
Fe_2O_3 (wt %)	-0.13	-0.54	-0.823	0.422
S (ppm)	2.77×10^{-5}	0.45	0.90	0.385
P_2O_5 (wt %)	-1.35	-0.43	-1.12	0.285
Sand (%)	-0.01	-0.38	-0.82	0.427
$\log_{10}\sigma'$ (kPa)	-0.05	-0.30	-0.61	0.552
Silt (%)	-0.004	-0.06	-0.18	0.859
I (ppm)	-0.001	-0.06	-0.20	0.847
SiO_2 (wt %)	0.002	0.02	0.08	0.938

Table 7.6 Multiple regression parameters and the significance of predictor variables obtained from backward elimination stepwise multiple regression analysis of the mudflat sediment core. The dependent variable is voids ratio, *e*. Variables are ranked according to their relative importance in prediction of *e*, as indicated by standardised beta coefficients. All predictor variables significantly contribute to the model.

	Unstandardised beta coefficient	Standardised beta coefficient	*t* statistic	*p*
Intercept (constant)	1.06	-	2.87	0.010
$\log_{10}\sigma'$ (kPa)	-0.10	-0.68	-3.52	0.002
Loss on ignition (%)	0.03	0.59	4.48	<0.001
P_2O_5 (wt %)	-1.64	-0.52	-2.35	0.030
MnO (wt %)	26.53	0.51	4.51	<0.001
CaO (wt %)	-0.11	-0.44	-2.92	0.009

With this more robust method of multiple regression, the most important predictor variable is effective stress. The remaining predictor variables suggest minor, additional controls on *in situ* voids ratio are exerted by the organic content and geochemical concentrations.

7.3.3 Factors affecting in situ voids ratios in the sample cores

If the multiple regression models above are valid, then variations in voids ratio in the low marsh core are primarily controlled by S (ppm), silt, I (ppm), loss on ignition, sand and, to a lesser extent, the remaining redox sensitive elements. In the mudflat core, where lithological variation is minimal, voids ratio is most strongly controlled by vertical effective stress, with additional contributions from loss on ignition, P_2O_5, MnO and CaO. This suggests that existing compression models are not only insufficient in terms of mechanical compression, but also may fail to predict voids ratio changes sufficiently well due to the importance of diagenetically sensitive elements. However, before models of autocompaction are developed on the basis of these findings, a critical and sensible assessment of the multiple regression models is required.

Two main issues arise which suggest that geochemical factors are not necessarily useful predictors of voids ratio and hence cast doubt upon the findings of the multiple regression analysis. Firstly, if the relative concentrations of the most important predictor variables in the low marsh core are considered, the integrity of their predictive power can be questioned. The trace elements S and I are present in extremely low concentrations (2000 – 6000 ppm and 50 – 525 ppm). Similarly, MnO (0.1 – 0.5 wt %), P_2O_5 (0.1 – 0.6 wt %) and CaO (1 – 2 wt %) make minimal contributions to the geochemical composition of the sediment. It therefore seems doubtful that such low concentrations of these substances, and the low downcore variations therein, are likely to exert such a dominant influence on voids ratio in relation to lithological variations. In the mudflat core, geochemical compounds and elements are also only present in what are deemed to be geotechnically insignificant quantities; P_2O_5 varies between 0.1 – 0.4 wt % and MnO between 0.054 – 0.060 %. Again, not only are these quantities small but it is also difficult to envisage the small variations therein significantly contributing to voids ratio changes. CaO may have more of an influence on *in situ* voids ratio, ranging between more substantial values of *c.* 4 and 7 %.

Secondly, by considering correlations between predictors variables (Tables 7.1 and 7.2), the argument that geochemical factors may not be useful predictors of *in situ* voids ratios gains strength. Strong relationships exist between lithological and geochemical variables. In the low marsh core, the two most important geochemical predictors variables, S (r = 0.817, p < 0.001) and I (r = 0.521, p < 0.001), both strongly correlate with loss on ignition, as do P_2O_5 (r = 0.810, p < 0.001) and MnO (r = 0.611, p = 0.007). CaO shows close association with the silt (r = 0.496, p = 0.036) and clay (r = -0.730, p = 0.001) fractions and Fe_2O_3 correlates with sand content (r = -0.737, p < 0.001).

In the mudflat core, relationships exist between the various lithological fractions and the three geochemical compounds selected using the backward elimination multiple regression method. MnO (r = -0.429, p = 0.032) and P_2O_5 (r = -0.513, p = 0.009) significantly correlate with the silt fraction. Indeed, P_2O_5 is associated with all grain size fractions (sand: r = -0.623, p = 0.001; clay: r = 0.802, p < 0.001). CaO is significantly correlated with loss on ignition (r = 0.518, p = 0.008) and the clay fraction (r = -0.430, p = 0.032).

Such strong correlations between geochemical and lithological variables suggest that the observed distribution of redox sensitive elements and compounds is not a function of redox remobilisation and zonation. Rather it may be a result of downcore compositional changes and/or due to preferential adsorption of redox sensitive substances onto different grain size fractions or organic substrates. The latter mechanism has been observed previously, and is driven by the bacterially-mediated remineralisation or organic carbon in decaying marsh vegetation. The available electron acceptors involved in the oxidation of organic carbon are successfully utilised by microbes in terms of decreasing thermodynamic advantage (O_2 before NO^-_3 before MnO_x before Fe_2O_3 before seawater SO^{2-}_4 before methanogeneisis; Froelich *et al.*, 1979). When such oxyhydroxides enter anoxic conditions (beneath the permanent water table) during burial, they undergo reductive dissolution (Thomson *et al.*, 2002), releasing any associated/complexed elements (Zwolsman *et al.*, 1993). As a consequence, dissolved diagenetically sensitive geochemical substances become mobile in pore water solution and are free to move vertically or advectively through the sediment to precipitate as or with hydrous Fe and Mn oxides in the oxic (or at least periodically oxic, mottled) zone. Generally, this occurs at the top of the sulphate reduction zone. However, where redox zonation is not clearly defined, as occurs at depths associated with a fluctuating water table and hence lacking a strong redox boundary, diagenetically mobile elements (particularly S and I) may instead be adsorbed onto organic compounds which provide reactive sorption sites (Cundy and Croudace, 1995a; A. Cundy, *pers. comm.*; A. Plater, *pers. comm*; Plater *et al.*, 1998). Such a phenomenon has also been documented in the clastic component by Zwolsman *et al.* (1993) and Cundy and Croudace (1995). In conjunction with the significant correlations between geochemical and lithological variables, this mechanism suggests that the geotechnical importance assigned to geochemical substances is in fact an artefact of sediment composition and preferential adsorption of redox sensitive elements onto various components of the lithology/substrate. Consequently, the predictive importance assigned to geochemical variables by the multiple regression modelling procedure can be attributed to their close association with lithological parameters.

On the basis of these arguments involving (insignificantly) low concentrations of elements and preferential adsorption onto various lithological fractions, it seems that the dominant factor in causing *in situ* voids ratio change within the overconsolidated low marsh vadose zone sediments is lithology (in the low marsh cores obtained) and effective stress (in the mudflat cores obtained). It is highly likely, therefore, that the multiple regression modelling is misleading and the observed trends in geochemistry are likely to be an artefact of preferential binding of redox sensitive elements onto lithological carrier phases, particularly the organic fraction. Geochemical factors, therefore, are judged to be passive and their distribution reflects their association with the 'true' (lithological) controls on voids ratio.

By disregarding geochemical variables, variations in voids ratio can now only be explained by variations in lithology and effective stress. Hence, in the low marsh cores, where lithology was highly variable, lithology is the most important predictor variable. This does not indicate that an effective stress modelling approach is insufficient. Rather, it is a reflection of the observed downcore variations in lithology and strongly highlights the need to develop individual compression models for each lithology in the low marsh core for successful pre- and retro-diction of volumetric change.

In the mudflat core, where lithology is constant, effective stress is the most important predictor variable for calculations of voids ratio. It is therefore likely that a stress-based compression model is sufficient and that geochemical parameters need not be incorporated into the autocompaction model.

7.4 THE EFFECTS OF DIAGENETIC POINT CONTACT CEMENTATION ON COMPRESSION BEHAVIOUR

Other than by filling void spaces following their precipitation, diagenetically sensitive substances can also act as point-contact cements (Hawkins, 1984; Tovey and Yim, 2002) that have the potential to change material compressibility and alter future mechanical behaviour from that observed in the virgin materials obtained from the depositional surface.

In order to ascertain the significance of such effects in compression behaviour, it is necessary to undertake oedometer tests on the diagenetically altered materials obtained from depth within the stratigraphic sequences analysed above. However, due to the rapid stratigraphic variation in lithology in the low marsh cores, it is only the lithologically homogenous mudflat cores that can reliably be used to determine any effects of diagenetic cements on compression behaviour. Therefore, oedometer samples were obtained from cores MFX-2, MFX-3 and MFX-5 at depths that showed noticeable relative enrichment or depletion in diagenetically sensitive substances (Figure 7.14). These sampling depths, along with sample IDs, are shown in Figure 7.15 and in Table 7.7, which also displays the lithological characteristics and initial structural parameters. Specific gravity measurements were made at 0.25 m and 0.49 m to ascertain whether any changes occurred downcore. Values were similar (mean = 2.63) and equalled those obtained from the surface samples (Table 5.2). $e\log_{10}\sigma'$ are displayed collectively in Figure 7.16. Material properties obtained from these plots are displayed in Table 7.8. Samples were tested for their loading behaviour only according to low stress scenario 1 (Table 6.2), followed by the incremental loading steps displayed in Table 6.1.

The downcore homogeneity of lithology has been noted previously but is confirmed by the data presented in Table 7.7. Values of initial voids ratio from the core samples (1.80 – 2.17; Table 7.7) fall within the same range as those measured on the surface samples (1.71 – 2.68; Table 6.5).

Table 7.7 Physical properties of oedometer samples obtained from cores MFX-4, MFX-5 and MFX-6.

Sample I.D.	Approximate sampling depth (m)	Loss on ignition (%)	Sand (%)	Silt (%)	Clay (%)	Natural moisture content, *w* (%)	Initial voids ratio, e_i
MFX-1-S-OED-24	0.17	18.65	7.1	75.5	17.4	77.00	2.16
MFX-2-S-OED-24	0.22	18.15	16.6	71.7	11.7	78.29	2.17
MFX-3-S-OED-24	0.30	17.06	13.0	75.5	11.4	76.38	2.09
MFX-4-S-OED-24	0.34	16.87	17.1	72.8	10.1	76.49	2.10
MFX-5-S-OED-24	0.40	13.50	10.3	78.9	10.9	68.93	1.80
MFX-6-S-OED-24	0.46	13.39	14.5	75.3	10.2	71.52	1.93

Table 7.8 Material properties of oedometer samples obtained from cores MFX-4, MFX-5 and MFX-6.

Sample I.D.	C_r	C_c	σ'_c (kPa)
MFX-1-S-OED-24	0.07	0.63	18
MFX-2-S-OED-24	0.09	0.62	17
MFX-3-S-OED-24	0.09	0.60	18
MFX-4-S-OED-68	0.09	0.62	17
MFX-5-S-OED-NC	0.09	0.53	16
MFX-6-S-BPS-24	0.09	0.57	18

The $e\log_{10}\sigma'$ plots in Figure 7.16 display compression trends that are consistent with those obtained from the surface samples in their virgin state. All samples are again overconsolidated. However, preconsolidation stresses range from 16 – 18 kPa, compared to 8 – 14 kPa in the surface samples. Also, the range of recompression indices is lower in and more uniform in the core samples (0.07 – 0.09; Table 7.8) than the surface samples (0.02 – 0.25), perhaps indicating a removal, and hence decreased importance, in variations in depositional structure and, consequently, of compressibility in the overconsolidated section of the stratigraphic column. These variations in preconsolidation stress and recompression index may reflect the influence of the diagenetic point contact cement manifesting itself as an increased structural resistance to compression and a higher stress transition to virgin compression (Gutierrez and Wangen, 2005; Nygard *et al.*, 2004). However, this trend may equally be indicative of the closure with depth of the bioturbation hollows and an associated increase in the structural integrity of the sediment.

Post-yield, however, the compression behaviour of the core samples closely matches that of the surface samples. This is revealed by a similar range of values for the compression index in both surface (0.50 – 0.73; Table 6.7) and stratigraphic (0.53 – 0.63; Table 7.8) samples. This suggests that any diagenetic modifications to compressibility are associated with the overconsolidated part of the stress range. Once structural breakdown occurs, any influence of a point contact cement is removed and there is no evidence of any memory of this low stress increase in compressive strength.

Figure 7.17 compares the samples obtained from the cores (red lines) with those sampled from the depositional surface (tested in Chapter 6; black lines). From Figure 7.17, it can be seen that the overall influence of shallow burial and any post-depositional modification of the sediment is minimal. Compression stress paths are highly similar. Although compression parameters vary slightly, particularly in terms of the recompression index and the preconsolidation stress, the one-dimensional compression stress paths of the MFX core samples are very similar to those of the surface samples.

7.5 SUMMARY OF PRINCIPAL FINDINGS

The aim of this chapter was to determine if and how diagenetic processes operating in the overconsolidated vadose zone had any influence on *in situ* voids ratios, particularly relative to vertical effective stress, and on the one-dimensional compression behaviour of undisturbed samples obtained from depth. Two main post-depositional processes were identified in the vadose zone stratigraphies. Firstly, visual evidence of diagenetic remobilisation of redox sensitive substances was observed in both low marsh and mudflat cores. Secondly, humification was identified in the low marsh cores. In the absence of a suitable quantitative technique for measuring the degree of humification in predominantly mineralogenic intertidal sediments, humification was not able to be considered any further in the statistical prediction of *in situ* voids ratios.

Variations in voids ratio with depth were compared to the depth profiles of lithological, geochemical and geotechnical variables. In the low marsh core, lithology varies rapidly stratigraphically. This is in contrast to the lithologically homogeneous mudflat core. Geochemically, redox zonation in the low marsh cores seemed to be minimal. In the mudflat core, however, there is a pronounced redox boundary that is reflected to a certain degree by the depth profiles of geochemical elements, particularly S, MnO and CaO.

Multiple regression analysis was undertaken to determine the relative importance of effective stress and lithological and geochemical factors on *in situ* voids ratios. Backward elimination stepwise regression suggested that geochemical factors were the dominant control on voids ratios in the low marsh core. Geochemical variables were of secondary importance to effective stress in the mudflat core, yet were still statistically of some importance. However, the overall importance of these variables was questioned and, ultimately, rejected since concentrations of these elements were deemed to be too low to affect voids ratio changes. The distribution of geochemical substances is likely to be an artefact of preferential adsorption onto lithological substrates (particularly the organic fraction).

One-dimensional (oedometer) compression tests undertaken on mudflat samples obtained from depths in the cores that displayed noticeable enrichment/depletion in redox sensitive elements demonstrated similar compression stress paths to tests performed on virgin surface samples. However, higher preconsolidation stresses were noted in the stratigraphic samples and this was attributed to the effect of weak diagenetic point contact cements and/or the closure upon burial of bioturbation structures.

Research hypothesis 6 was defined in Section 3.10 as follows:

6. Diagenetic changes are unimportant and do not significantly modify mechanical compression behaviour.

On the basis of the data presented in this chapter (particularly those obtained from the mudflat core), effective stress remains the dominant control on voids ratio in the overconsolidated vadose zone. Also, post-depositional geochemical modification of the sediment does not significantly alter the compression behaviour of recently-formed sediments obtained from the depositional surface. Hence, hypothesis 6 cannot be rejected. However, the effect of humification in saltmarsh materials on voids ratio and compression behaviour remains a key unknown.

Each of the six research hypotheses have now been addressed following investigations into mechanical and (bio-)chemical autocompaction processes in intertidal vadose zone stratigraphies. It is now possible to develop an empirically-informed predictive autocompaction model. This is undertaken in Chapter 8.

CHAPTER 8: MODEL DEVELOPMENT, APPLICATION, AND DISCUSSION

8.1 MODELLING AUTOCOMPACTION IN MINERALOGENIC INTERTIDAL SEDIMENTS

8.1.1 Implications of the testing program for modelling autocompaction

Detailed investigations into the geotechnical environment in intertidal areas; the deformation mechanisms of mineralogenic intertidal sediments; and the influence of diagenetic processes on the behaviour of these sediments provide a firm empirical base for modelling autocompaction. The conclusions obtained from these investigations allow informed decisions to be made about the best way to model compression behaviour, namely:

1. Rather than using single values for the recompression and compression indices, a statistical model is needed for each material, incorporating the full range of structures and behaviours present;
2. A 'conditional' regression model in conventional $e\log_{10}\sigma'$ space is required, with values of the gradient and the intercept being conditional on whether a value of effective stress is less or greater than the initial 'yield' (not preconsolidation) stress;
3. In order to reflect variations in structure and behaviour pre- and post-yield, the error term must, if necessary, be allowed to vary on either side of the yield stress;
4. A time variable to describe creep is not necessary.
5. Redox-driven diagenetic processes affect neither *in situ* voids ratio nor compression behaviour of mineralogenic intertidal sediments enough to justify an increase in model complexity. Accordingly, autocompaction models will be developed solely on the basis of the surface compression behaviour.

8.1.2 Model development using Bayesian changepoint regression

Incorporating the model requirements outlined above requires an appropriate modelling technique. Accordingly, autocompaction models in $e\log_{10}\sigma'$ space have been developed using a well-established Bayesian statistical technique called changepoint regression modelling (Carlin *et al.*, 1992). Changepoint regression models allow the two linear segments of the compression stress paths (i.e. the overconsolidated, low gradient unload-reload hysteresis loop; and the higher gradient normal compression line) to be combined in a single statistical model. In this study, the changepoint corresponds to the yield stress and it describes the value of effective stress at which the two linear segments intersect. During analysis, Bayesian statistical techniques are firstly used to estimate the value of the changepoint/yield stress. Secondly, linear regression models are defined for effective stress values less than and greater than the modelled changepoint/yield stress. In addition, the changepoint model allows different parts of the dataset to obey different probability laws. Therefore, different error terms can be defined for each segment of the model – i.e. at effective stresses that are both less than and greater than the changepoint/yield stress. This facilitates adequate description of variations in structure and behaviour both pre- and post-yield.

Most recently, and indeed pertinently, changepoint regression has been used in sea level research by Parnell (2005) to determine whether the effects of well-studied periods of Holocene climate change, such as the 8.2 k event (e.g. Rohling and Pälike, 2005), the Little Ice Age (e.g. Matthews and Briffa, 2005) and the Mediaeval Warm Period (e.g. Esper *et al.*, 2002), can be identified in sea level records from the Humber estuary, UK. By employing changepoint analysis, Parnell (2005) was able to statistically determine whether these events resulted in rapid and distinct sea level changes or insignificant, diffuse changes. The exact timing of any such changes (i.e. the changepoint) was also estimated.

Parnell (2005) used the WinBUGS (Bayesian inference Using Gibbs Sampling) software package to undertake his analysis and the code used therein has kindly been made available for the changepoint regression modelling in this study. The model is based on a set of parameters: α, a constant, C_r, the recompression index and pre-yield stress gradient, C_c, the compression index and post-yield stress gradient and x_c, the estimated changepoint, equal to the common logarithm of the modelled yield stress. The model is formulated as:

$$e = \alpha + C_r(\log_{10}\sigma' - x_c) \textbf{ if } \log_{10}\sigma' \leq x_c \quad (8.1)$$

$$e = \alpha + C_c(\log_{10}\sigma' - x_c) \textbf{ if } x_c < \log_{10}\sigma' \quad (8.2)$$

where e is the voids ratio predicted by the model and σ' is a value of effective stress.

The performance of the regression models and the regression coefficients of each model developed are displayed in Table 8.1. High r^2 values (0.94 for the low marsh model and 0.81 for the mudflat) indicate that both models have good predictive capacity. This is also illustrated in plots of estimated against observed voids ratios (Figures 8.1 and 8.2). It is also encouraging to note that the modelled optimum values of the compression index compare favourably with the values presented in Table 6.5. Values of the recompression index differ to those in Table 6.5; this can be attributed to the larger variation of values of C_r prior to the structural breakdown at yield. As observed in laboratory tests, the modelled yield stress is less than the graphically estimated preconsolidation stress and represents the stress at which the compression stress path begins to move from the recompression line to the normal compression line.

Table 8.1 Performance of models and values of regression parameters of the changepoint regression models.

	r^2	Standard error of estimate	x_c	Yield stress (kPa)	α	C_r	C_c
Low marsh	0.94	0.25	1.10	12.62	4.18	-0.10	-1.61
Mudflat	0.81	0.24	0.72	5.22	2.24	-0.01	-0.612

Regression lines and the associated errors (95 % prediction limits) are displayed in Figures 8.1 (low marsh) and 8.2 (mudflat). Although changepoint models allow the standard error of the estimate to be varied on either side of the yield stress, it can be seen from Figures 8.1 and 8.2 that a single value sufficiently incorporates the majority of data when it is expressed as 95% prediction limits. At this stage of modelling, therefore, further analysis of the error term is not attempted.

8.1.3 Comparison of changepoint regression models with x-ray-derived sedimentation compression curves at low effective stresses

In order to ascertain the predictive accuracy of the changepoint regression models developed above, it is useful to compare the model predictions with sedimentation compression curves obtained from borehole measurements. This is because it is still of course possible that small variations in geochemical substances do exert some influence on structure, despite their perceived insignificance. In particular, Fe_2O_3 and CaO are present in both cores in sufficiently high levels to warrant further investigation into their effects. However, the changepoint regression models defined above provide empirically-informed descriptions of the compression behaviour of the low marsh and mudflat materials over an effective stress range of 0 – 1500 kPa. Obtaining sedimentation compression curves for these sediments over this stress range is logistically very difficult given the overburden depths required to cause such stresses. Furthermore, from Figures

7.1 and 7.7, it is evident that the contemporary low marsh material (upon which the changepoint regression model is based) only extends to a shallow depth within the core (*c.* 0.07 m). Arguably, from Figure 7.7, this depth may only be as shallow as *c.* 0.03 m, since loss on ignition increases from the surface value of *c.* 25 % to *c.* 30 %. Hence, it has not been possible to obtain analogous sediments for the full stress range described by the changepoint regression models. Comparison of the models with sedimentation compression curves is therefore only possible at shallow depths and hence very low effective stresses (*c.* 0 – 3 kPa) lower than the yield stress. Model predictions are therefore compared with the sedimentation compression curves derived from the x-ray scanning of the shallow cores examined in Chapter 7.

The low marsh changepoint regression model predicts well within the top section of the core (to a depth of *c.* 0.03 m, equivalent to *c.* 0.06 kPa), with *in situ* values of voids ratio generally plotting within the 95 % prediction limit error envelope (Figure 8.3). As lithology changes and the geotechnical properties change, the predictive capacity of the model diminishes rapidly, as expected.

Given the homogeneous nature of the lithology in the mudflat cores, the changepoint regression model developed on the contemporary mudflat material predicts well throughout the depth of the cores obtained (Figure 8.4). Variations in *in situ* voids ratio all plot within the 95 % prediction limit error envelope. Hence, any variations in voids ratio caused by geochemical enrichment that do occur are no greater than those arising from depositional structural variations. This finding leads to an increase in confidence in an effective stress-based compression model and justifies the rejection of geochemical variations as significant predictor variables in determination of *in situ* voids ratios. Consequently, the effective stress-dependent compression models are deemed to be sufficient in predicting voids ratios in the overconsolidated vadose zone in mudflat materials at low effective stresses.

8.2 PRACTICALITIES OF MODEL APPLICATION

The models developed above display good predictive capacity within the overconsolidated section of intertidal stratigraphies (i.e. at low stresses). Of greater interest, however, is the performance of the models in practical situations when they are used in pre- and retro-diction of volumetric change in marsh stratigraphies. Before this can be done, however, a number of interrelated issues must first be addressed. The first of these involves the compression behaviour of the materials that underlie the contemporary low marsh materials in the low marsh cores. Although *in situ* voids ratios are known through these cores (Section 7.2.1), this does not provide an understanding of how volume will change in response to applied effective stress. Secondly, since intertidal environments display transitional variations in sediment with altitude, it is important to understand how to group sediment types and their corresponding compression behaviour prior to statistical modelling.

8.2.1 Compression behaviour of saltmarsh sediments

The foraminifera contained within the low marsh core strata indicate that the entire core formed in saltmarsh environments (Section 7.2.3). More specific reconstructions of depositional sub-environments (i.e. marsh zone) can be made by considering previous work undertaken at Greatham Creek into the contemporary altitudinal distribution of foraminifera. Multivariate analyses undertaken by Horton and Edwards (2000) separate the observed surface foraminiferal assemblages into three zones, as summarised in Table 8.2. The first of these, representative of middle and high marsh sub-environments (3.24 – 2.22 m OD; Horton and Edwards, 2000), is a low species diversity, agglutinated assemblage dominated by *Jadammina macrescens* and *Trochammina inflata*.

The low marsh zone (2.22 – 2.04 m OD) is dominated by *Miliammina fusca*, and typically shows additional, yet low frequencies, of calcareous species. The third zone revealed by statistical analysis is the mudflat zone (2.09 – -0.35 m OD), which has a high diversity assemblage that is dominated by calcareous species including *Elphidium williamsoni*, *Haynesina germanica* and *Quinqueloculina* spp.. It is important to note that the foraminiferal zones are somewhat different to the floral zones displayed in Table 4.2. However, this does not preclude the use of the foraminiferal zones in reconstructing the elevation of the saltmarsh surface relative to the intertidal frame.

Table 8.2 The dominant foraminiferal taxa of Cowpen Marsh (Source: Horton and Edwards, 2000).

Mudflat	Low marsh	Middle and high marshes
Elphidium williamsoni	*Miliammina fusca*	*Jadammina macrescens*
Haynesina germanica		*Trochammina inflata*
Quinqueloculina spp.		

It seems likely that the low marsh cores record long-term (annual-decadal) variations in flooding frequency. At the base of the core, the dominance of *Jadammina macrescens* indicates that this sediment formed in a high marsh environment, close to the landward edge of the saltmarsh (after Scott and Medioli, 1980). The increase in species diversity and the appearance of *Trochammina inflata* and *Miliammina fusca* indicates an increase in flooding frequency and a fall in the elevation of the sediment surface relative to mean sea level. This may have resulted from desiccation of the high marsh (*Jadammina macrescens*-dominated) sediment as it rose out of the intertidal frame; this desiccation would result in a reduction in volume and a lowering of the relative marsh elevation.

Above 0.15 m, foraminiferal species diversity once again drops, and the near-monospecific *Jadammina macrescens* assemblages indicate a high marsh environment. This decrease in flooding frequency is likely to be associated with further sediment accumulation raising the relative elevation of the saltmarsh surface (Allen, 1990). At shallower depths in the core, towards the contemporary saltmarsh surface, there is a decrease in the dominance of *Jadammina macrescens* and an increase in *Trochammina inflata*, indicating a second submergence event (increase in flooding frequency), where the foraminiferal assemblage suggests a mid-marsh environment.

The low marsh cores therefore record a range of saltmarsh sediment types, from low to high marsh. However, the lithological parameters (particle size and loss on ignition variation) in the core (Figure 7.7) may appear to contradict the relative elevation history inferred from the foraminifera. For example, the monospecific, high marsh *Jadammina macrescens* assemblages coincide with the lowest loss on ignition values in the core (*c.* 15 %). Such low values are typical of mudflat environments, whereas the contemporary high marsh environment is typified by loss on ignition values of 60 – 85 % (Figure 5.3). This discrepancy can be explained by the increased exposure of the high marsh sediments to desiccating and humifying subaerial conditions that result in significant breakdown of organic compounds, which are subsequently bacterially utilised or leached away. Consequently, the preservation of the surface loss on ignition 'signature' is low. The darker, more humified nature of the lower stratum also suggests greater subaerial exposure and drier (frequently less flooded) conditions (Figure 7.1). This evidence of post-depositional behaviour severely complicates attempts to model the autocompaction behaviour of these sediments, since volumetric change no longer becomes solely dependent on effective stress. Other factors involving physical and chemical decay of the organic fraction also play important roles.

The hypothesised loss of organic content following humification may also have had an effect on the compression behaviour of the high- and mid-marsh samples, as revealed by oedometer tests on these sediments. These were obtained from cores LMX-3 and LMX-4. Sampling depths and IDs are displayed in Figure 8.5 and in Table 8.3, which also displays the lithological characteristics and initial structural parameters. Table 8.4 displays the marsh sub-environment of deposition, as inferred from foraminiferal assemblages.

Specific gravity measurements were taken at *c.* 0.10 m depth intervals beneath 0.07 m; values were found to be largely constant, ranging from 2.58 to 2.61 (mean = 2.60). $e\log_{10}\sigma'$ are displayed collectively in Figure 8.6. Material properties obtained from these plots are displayed in Table 8.5. Samples were tested for their loading behaviour only according to low stress scenario 1 (Table 6.2), followed by the incremental loading steps displayed in Table 6.1.

Table 8.3 Physical properties of oedometer samples obtained from cores LMX-3 and LMX-4.

Sample I.D.	Approximate sampling depth (m)	Loss on ignition (%)	Sand (%)	Silt (%)	Clay (%)	Natural moisture content, *w* (%)	Initial voids ratio, e_i
LMX-1-S-OED-24	0.05	28.50	0.2	71.8	28.0	217.07	5.57
LMX-2-S-OED-24	0.09	14.65	0.3	67.6	32.1	96.17	2.54
LMX-3-S-OED-24	0.13	15.02	2.0	72.0	26.0	108.68	2.95
LMX-4-S-OED-24	0.16	22.16	7.1	69.7	23.2	105.06	2.85
LMX-5-S-OED-24	0.22	22.72	6.4	79.0	14.6	111.34	2.96
LMX-6-S-OED-24	0.31	16.21	8.9	76.9	14.2	115.73	3.16
LMX-7-S-OED-24	0.35	15.49	8.8	78.6	12.6	99.25	2.71

Table 8.4 Inferred marsh sub-environments of oedometer samples obtained from cores LMX-3 and LMX-4.

Sample I.D.	Approximate sampling depth (m)	Inferred (foraminiferal) marsh environment of formation
LMX-1-S-OED-24	0.05	Mid marsh
LMX-2-S-OED-24	0.09	High marsh
LMX-3-S-OED-24	0.13	High marsh
LMX-4-S-OED-24	0.16	Low marsh
LMX-5-S-OED-24	0.22	Low marsh
LMX-6-S-OED-24	0.31	Low marsh
LMX-7-S-OED-24	0.35	High marsh

Figure 8.6 and Table 8.3 display that the majority of samples obtained from the low marsh cores have very similar initial voids ratios (2.54 – 3.16), are overconsolidated and display similar compression behaviour both (pre- and post-yield). The compression stress paths of these samples consequently plot at very similar locations in $e\log_{10}\sigma'$ space, despite variations in lithology (particularly loss on ignition) and the marsh zone in which they were formed (Tables 8.3 and 8.4). The exception to this clustering of behaviour is sample LMX-1-S-OED-24, which has a considerably higher initial voids ratio and, resultantly, a much steeper virgin compression line (C_c = 2.71).

Table 8.5 Material properties of oedometer samples obtained from cores LMX-3 and LMX-4.

Sample I.D.	C_r	C_c	σ'_c (kPa)
LMX-1-S-OED-24	0.24	2.71	25
LMX-2-S-OED-24	0.10	0.85	24
LMX-3-S-OED-24	0.14	0.98	23
LMX-4-S-OED-24	0.11	1.02	22
LMX-5-S-OED-24	0.12	1.08	24
LMX-6-S-OED-24	0.14	1.13	27
LMX-7-S-OED-24	0.14	0.93	28

It seems that the compression behaviours of the saltmarsh materials in this core are not directly related to the marsh sub-zone and the elevation relative to the tidal frame at which the sediments formed. This may result from post-depositional processes, particularly humification; the degradation and removal or organic matter has been known to result in structural collapse of organic intertidal soils (Delaune *et al.*, 1994). However, Tipping (1995) distinguishes between primary (syn-depositional) and secondary (post-depositional) humification. Hence, it may be possible that the observed loss on ignition values throughout the low marsh core are in fact unaltered since their time of formation. The *in situ* lithological condition may be a result of a combination of factors that initially lead to low organic contents in the mid and high marshes, such as poor conditions for *in situ* organic growth, a strong advected or aeolian mineralogenic component and/or strong humification of the organic component during deposition/formation.

8.2.2 Grouping material behaviour for statistical analysis

The compression models developed so far only consider two materials. In the transitional intertidal environment under a regime of consistently rising/falling sea level, any changes in relative marsh elevation will lead to transitional variations in organic content and hence compression behaviour within a stratigraphic column. Compression behaviour is therefore also unlikely to be organised into discrete material units. This raises the question of how to deal with gradational saltmarsh lithologies in terms of their compression behaviour – how can materials best be grouped together, on what basis (elevational, lithological or geotechnical) and what threshold values should separate 'different' material types?

It was noted (Sections 5.12 – 5.14; Figures 5.3, 5.6 and 5.7) that lithological variations in the mudflat zone are minimal, whereas organic content increases in quasi-proportionality with altitude in the saltmarsh zone. The development of floral zone-specific models of compression behaviour would, however, assume that marsh zones are fully discrete both lithologically and geotechnically and that no overlap exists between altitudinally contiguous samples. However, variations in lithology are diffuse and transitions between floral zones occur over a range of altitudes in a patchy 'mosaic' of vascular plant assemblages and their associated lithologies. Dividing the contemporary transect into distinctive zones consisting of largely homogenous sediments was not aided by unconstrained cluster analysis. Assigned material zones (clusters) overlapped considerably, reflecting the gradational nature of sediment variations with respect to altitude. Furthermore, a grouping approach based on marsh zones would assume that post-depositional changes to lithology and compression behaviour are insignificant. However, although geochemical effects as a result of redox zonation have been considered and deemed to be passive in the shallow stratigraphies considered, humification seems to have more of an effect on lithology and compression behaviour. Indeed, this has been illustrated by the similarity of behaviour observed in sediments that, on the basis of their microfossils contents, formed at different positions within the intertidal frame. As elevation relative to mean sea level increases, tidal influence decreases and a greater number of factors play a role in determining compression behaviour. These factors are related to the magnitude and frequency of subaerial exposure. Analysis of the bio- and litho- stratigraphies and the downcore geotechnical properties suggests that the compression behaviour of saltmarsh sediments is not necessarily linked to relative elevation. Hence, it seems more appropriate to group sediments for compression modelling together on the basis of *in situ* lithology rather than 'parent' marsh zone.

Figure 8.7 considers all of the one-dimensional compression tests undertaken in this study. Overall, samples with greater initial voids ratios are more prone to compression, as indicated by the extremely strong predictive relationship between e_i and C_c (r^2 = 0.96)

(Figure 8.8). The same trend was observed in the contemporary low marsh and mudflat sediments (Section 6.2). From the relative graphical 'locations' of different materials on Figure 8.7, it is also evident that a relationship exists between organic content, initial voids ratio and compression behaviour. The mudflat samples constitute the least organic group (loss on ignition of *c.* 13 – 18 %), and these plot beneath all samples obtained from saltmarsh environments. Plotting directly above the mudflat samples are those obtained from the low marsh core. These partially humified sediments have intermediate organic contents (loss on ignition of *c.* 15 – 22 %). The unhumified, contemporary sediments have the highest organic contents (*c.* 23 – 28 %) and plot highest with respect to the y-axis. This suggests that organic content has a fundamental control on structure; indeed, the relationship between loss on ignition and the initial voids ratios of the samples is strong (r^2 = 0.81) (Figure 8.9). Similarly, loss on ignition displays a strong control on the compression index, C_c (r^2 = 0.79, Figure 8.10).

These observations confirm that the compression behaviour is a result of *in situ* lithological rather than elevational controls. Since a statistical modelling approach requires sediments of similar behaviour to be grouped together, it seems that lithology provides a sensible criterion for doing so. Such an approach, however, is still problematic, since a continuum of lithologies and resultant compression behaviour exists (Figure 8.9). Therefore, decisions must be made regarding threshold values with which to differentiate between materials prior to changepoint modelling.

On the basis of evidence presented throughout this thesis, sediments obtained from mudflat environments are homogenous in many respects. The section of the contemporary mudflat environment that was sampled in the sedimentological transect in Chapter 5 covers altitudes between 0.34 m OD to 1.75 m OD. Within this 1.41 m altitudinal range, loss on ignition values remain constant at approximately 15 % (Figure 5.3). Similarly, grain size variations do not significantly vary throughout this zone (Figures 5.6 and 5.7). On the basis of one-dimensional compression tests on both surface and core samples, the geotechnical properties are also essentially constant, despite the core samples inevitably having formed at varying elevations relative to the tidal frame (as suggested by the foraminiferal assemblages; Figure 7.12). A change in the relative elevation of the depositional surface does not result in any dramatic changes in lithology or compression behaviour. Providing the relative elevation of the depositional surface does not increase to a level that permits autogenic organic growth, a potentially deep and continuous sequence of mudflat sediments can form, within which lithological and geotechnical properties remain unchanged. As such, compression models developed on surface samples are likely to be applicable to all mudflat sediments that formed at similar elevations within the intertidal frame. In addition, it is evident from Figure 8.7 that mudflat compression samples form a relatively discrete cluster of compression behaviour. It therefore makes sense to group these sediments together and separate them from the transitional zone of saltmarsh materials.

In contrast, it is the variations in organic content in saltmarsh sediments that cause problems in assigning different material groups. Just as it was difficult to assign thresholds to differentiate between lithological zones in the contemporary transect, the gradational nature of material property changes is likely to make the designation of groups on the basis of lithology and compression behaviour equally as difficult. Indeed, without additional data on the full range of organic lithologies (in both their virgin state and diagenetically modified to different degrees), such a partitioning of material groups is beyond the scope of this study.

For the purposes of model application, it is possible to split the remaining saltmarsh materials into two groups. The first of these ('Low Marsh 2') groups sample LMX-1-S-OED-24 with all remaining contemporary low marsh samples tested in Chapter 6. The reason for this grouping is that these samples have very similar lithological characteristics, particularly organic content (compare Table 6.5 with Table 8.3). They also have a similar stratigraphic setting, both being near-surface and, on the basis of visual observation, are largely diagenetically unaltered (Figures 7.1 and 7.4). Perhaps most importantly, the compression behaviours displayed by the samples are similar (Figure 8.7).

The second group ('Saltmarsh 1') involves the remaining samples obtained from the low marsh cores. These samples have been shown to display similar compression behaviour (Figures 8.6 and 8.7). Such a grouping becomes even more reasonable and valid if it is assumed that the observed lithological and geotechnical characteristics have resulted from syn-depositional, rather than post-depositional, processes. Hence, the observed structure and compression behaviour are likely to represent those since burial began. The statistical approach also increases the range of behaviour described by the model, increasing the accuracy of the models, albeit at the expense of model precision.

Changepoint regression parameters for each of these models are displayed in Table 8.6. Due to the range of voids ratios observed in the sedimentation compression curves (Figure

8.3), the modelling code was manipulated to allow the error term to be varied at effective stresses greater and less than the yield stress. By doing so, both models fully utilise the capabilities of the changepoint regression modelling approach.

Comparison of the two low marsh models in Tables 8.1 and 8.6 reveals that the model has retained its good predictive capacity with the inclusion of sample LMX-1-S-OED-24 (r^2 = 0.90). Post-yield regression parameters (C_c and standard error of the estimate) are identical. Values of the recompression index, C_r, differ due to the inclusion of sample LMX-1-S-OED-24. This is due to the steeper pre-yield compression stress path of the sample. Similarly, the estimated yield stress (and changepoint) has decreased. The largest change, however, has occurred in the standard error of the estimate, which has increased from 0.25 to 0.68. The 'Saltmarsh 1' model displays excellent predictive capacity (r^2 = 0.96) (Table 8.6, Figure 8.11). Values of the modelled recompression and compression indices compare favourably with those obtained experimentally.

These models are compared with sedimentation compression curves derived from the x-ray scanning of the cores in Figure 8.12. Standard errors of the regression models are expressed as 95 % prediction limits. The model Low Marsh 2 is applicable to sediments above 0.07 m, equivalent to effective stresses lower than *c.* 0.17 kPa. The model Saltmarsh 2 is applicable to sediments below 0.07 m (stresses greater than *c.* 0.17 kPa). For both models, different values of the standard error at stresses greater and lower than the yield stress is evidently justified, since the range of *in situ* voids ratios for each material plot largely within the error envelopes of their respective changepoint models. Despite this, there is a strong decreasing trend in voids ratios in cores LMX-2 and LMX-4 (Figure 8.12). This suggests a more rapid reduction in voids ratio following effective stress application than is predicted by the model. However, the decreasing trend occurs in sudden 'jumps' rather than gradually; this is particularly evident in the sedimentation compression curve of core LMX-4. Hence, it is likely that the reduction is a result of lithological changes within each stratum. Furthermore, since the model is composite, based on a number of similar lithologies, the modelling process averages and smoothes the modelled curves and this results in the shallower gradient (low value of C_r) of the model predictions compared to the *in situ* curves.

Table 8.6 Performance of models and values of regression parameters of the changepoint regression models developed on low marsh core materials. S.E.E. 1 = standard error of the estimate at effective stresses less than the yield stress/changepoint. S.E.E. 2 = standard error of the estimate at effective stresses greater than the yield stress/changepoint.

Material/model	r^2	S.E.E. 1	S.E.E. 2	x_c	Yield stress (kPa)	α	C_r	C_c
Low marsh 2	0.90	0.68	0.25	0.91	8.13	4.52	-0.12	-1.61
Saltmarsh 1	0.96	0.48	0.15	0.92	8.39	2.92	-0.09	-0.96

8.3 Implications and application of the empirically-informed autocompaction models

In a purely geotechnical sense, the field and laboratory investigations have yielded interesting results in their own right. However, in order to address the original thesis rationale, these data must be considered practically and applied to intertidal stratigraphic sequences in order to evaluate their predictive capacity. This section discusses some of the implications of the changepoint regression models to the various problems associated with the autocompaction of intertidal sediments before applying the model to a short stratigraphic sequence.

8.3.1 Implications for elevation adjustment in low energy upper intertidal environments

Validating and applying the changepoint regression models in a predictive capacity to determine future changes in marsh surface elevation is not possible in this thesis for two main reasons. Firstly, a long-term elevation monitoring approach would be required, involving the use of a Sedimentation-Erosion Table (SET), differential GPS or a remote sensing technique. Secondly, prediction of marsh surface elevation adjustments requires that the full sediment column is considered; this would require the development of a further autocompaction models for a wider range of (potentially highly organic) lithologies which, again, is beyond the scope of this investigation. Nonetheless, on the basis of the findings presented in this thesis, it is possible to speculate on their implications for elevational adjustment in intertidal areas.

Cahoon and Lynch (1997) suggested that it is near-surface sediments that contribute to the majority of volumetric reduction within a stratigraphic column and, hence, surface elevation change. However, this finding was based on measurements of vertical accretion and autocompaction in a mangrove substrate in southwestern Florida, USA. Nonetheless, it is interesting to consider the implications of the initial overconsolidation of mineralogenic sediments for estimates of surface elevation change in the less organic marshes of northwest Europe. If the low marsh deposit (modelled yield stress = 12.62 kPa) were buried by a saturated fine sand of constant density with depth (bulk density ≈ 2 g/cc; unit weight ≈ 20 kN/m^3; values from Powrie, 2004), the low marsh soil would not undergo virgin compression until the overburden thickness exceed *c.* 1.25 m (calculations based on Equations 3.1 and 3.2). In a mudflat deposit (modelled yield stress = 5.22 kPa), the overburden thickness would have to exceed *c.* 0.50 m before virgin compression commenced. However, if buried by a low density material obtained from the intertidal zone (bulk density ≈ 1.4 g/cc; unit weight ≈ 14 kN/m^3), the low marsh material would not deform until the overburden depth exceeded 3.0 m. In the mudflat, this depth is approximately 1.24 m.

Clearly, these are highly idealised conditions, but the calculated overburden depths that would be required to initiate virgin compression indicate that near-surface sediments are generally resistant to large strain compression. Consequently, it is reasonable to assume that in predominantly mineralogenic sequences, major contributions to reductions in thickness of a sediment column come not from the near-surface sediments, but from those at depths at which effective stresses exceed the yield stress. This finding is also contrary to that implied by Terzaghi's compression law, the logarithmic nature of which suggests that it is recently deposited materials at low effective stresses that undergo the most rapid compression following burial.

8.3.2 Implications for reconstructions of late Holocene sea level

Massey *et al.* (2006) provide one of the few attempts to quantify autocompaction and correct for its effects using geotechnical theory (Paul and Barras, 1998, which uses Terzaghi's compression law as its rheological model; Section 3.4). Geotechnical corrections were used to apply an altitude correction to Holocene coastal back-barrier sediments from North Sands and Blackpool Sands, south Devon, UK. Vertical corrections ranged from < 0.1 m in argillaceous sediments situated directly above incompressible basal substrates to > 1.0 m at regressive contacts between mineralogenic and partially organogenic facies, to > 2.0 m in peats. Using the corrected (decompacted) data, Massey *et al.* (2006) considered the implications for estimations of late Holocene rates of sea level rise. They found that relative sea level rise in south Devon during the mid – to late Holocene is 1.27 mm yr^{-1} for the *in situ* (uncorrected for autocompaction) data and 1.16 mm yr^{-1} for decompacted (corrected) data – a total error of 0.11 mm yr^{-1}.

Although this geotechnical work by Massey *et al.* (2006) has further highlighted the issue of autocompaction, the lack of empirical input into the governing rheological model (Terzaghi's compression law) may have resulted in revised estimates of rates of relative sea level change that are no more accurate than the uncorrected data. The overconsolidated nature of the predominantly mineralogenic materials considered in this study suggest that altitudinal corrections will not be as great as those predicted from a backward projection of the normal compression line. A decompaction procedure based on the findings of this study would lead to some degree altitudinal correction. However, refining the work of Massey *et al.* (2006) would require a considerable degree of further investigation, since the additional parameters required for the changepoint regression models developed in this thesis (x_c and C_r) have not been obtained for even the mineralogenic sediments in the south Devon cores. Even with these parameters, only a partial altitudinal correction could be applied since corrections for autocompaction are fully dependent on the cumulative compression of all underlying sediments. The lack of empirically-informed, phenomenologically correct autocompaction models of peat facies preclude a full decompaction of the sediment columns. Any resulting conclusions drawn regarding rates of mid- to late Holocene sea level variations could be described as speculative at best and highly erroneous and misleading at worst.

Without a full understanding of the autocompaction of the full range of materials in the core, only a general conclusion can be drawn on the basis of the findings from investigations into mineralogenic intertidal sediments: the 'correct' altitude of the sea level index points obtained from the south Devon cores is likely to lie somewhere between the *in situ* compacted state and that obtained from the decompaction using Terzaghi's compression law.

8.3.3 Application of the autocompaction models to recent sediments

Due to the various problems outlined above, validation of the autocompaction models has not been possible in the contexts described. However, since the near-surface overconsolidated sediments of the vadose zone have been the focus of much of this thesis, applying the autocompaction models developed to the short stratigraphic sequences analysed in Chapter 7 seems a suitable choice. Furthermore, by considering these short sequences only, problems associated with the autocompaction behaviour of deeper and poorly understood organic facies are removed.

Hence, the aim of this section is to validate the predictive capacity of the 'new' autocompaction model developed in this thesis and compare it with that of Terzaghi's compression law by comparison with compaction-free, instrumental tide-gauge records. Following this, some implications for recent sea level records extracted from mineralogenic intertidal sediments are discussed.

Before this can be done, it is important to realise that comparison and validation of models is critically dependent upon the quality of the microfossil assemblages contained within the cores. Also, since only *c.* 0.40 – 0.50 m cores are used, there is likely to be an altitudinal offset of reconstructed mean sea level with that recorded by the tide gauges, since no correction for autocompaction in the underlying strata is considered. However, it is not the absolute altitude that is of interest; rather, it is the reconstructed rate of sea level change observed in the geological record. Since compaction reduces the relative vertical distance between sea level index points, and since the rate of sea level is typically considered in mm yr^{-1}, the observed rate in the sediments is prone to modification by autocompaction. Hence, the following application of the decompaction models considers the *in situ* rate and compares it with the decompacted rates obtained from each autocompaction model.

8.3.4 Compaction-free instrumental records of twentieth century mean sea level variations

Two tide gauge records have been considered to obtain a compaction-free, instrumental record of relative sea level change against which the recent, near-surface geological records can be compared. The first of these was obtained from the Hydrographic Surveyor (Mr. J.A. Robinson) of the Tees dockland authority, PD Teesport Limited. The tide gauge apparatus is located at the entrance to the Tees dockland area and so has direct access to marine tidal conditions. Surprisingly, given the long history of the Tees estuary as an industrial port (Plater *et al.*, 1998), only a 15 year record is available, stretching from 1988 to 2003.

The second, longer-term and hence more useful record was obtained from the North Shields tide gauge record (data obtained from the Proudman Oceanographic Laboratory (POL)/Permanent Service for Mean Sea Level (PSMSL) website: http://www.pol.ac.uk/psmsl/psmsl_individual_stations.html). The North Shields tide gauge record dates back to 1896 and is largely continuous. It provides one of the few UK records that began prior to the start of the twentieth century. This second tide gauge record was selected due to its geographic proximity to the Tees estuary (*c.* 50 kilometres). Comparison of the two tide gauge records indicates that sea level variations are similar in the two areas (Figure 8.13). The exception to this is from 1999 onwards, when mean sea level in the two areas diverge. This may reflect the influence of dredging activities in the Tees estuary. Nonetheless, the two sites have similar reference water levels; the contemporary altitude of mean sea level at River Tees Entrance is 0.35 m OD, compared to 0.30 m OD at North Shields (Admiralty Tide Tables, 2005). Furthermore, due to their proximity to the isostatic 'hinge line' of the British Isles, between-site isostatic variations are negligible (Shennan and Horton, 2002). It is therefore assumed that sea levels and their variations recorded by the North Shields tide gauge sufficiently reflect a regional sea level signal. Hence, the geological record of relative sea level from the Tees estuary can be directly compared with the North Shields instrumental record.

Despite short-wave (decadal) fluctuations in mean sea level, the North Shields tide gauge records a long-term rate of sea level rise of 1.9 mm yr^{-1}.

8.3.5 Transfer functions and reconstructing relative marsh elevation

In order to assist in the analysis of palaeo-surface elevation relative to the intertidal frame, two transfer functions (numeric, multiple-regression based procedures) have been employed. On the basis of modern data (termed the 'training set'), highly accurate and precise (in a statistical sense at least) estimations can be obtained of the elevation relative to mean sea level (metres) (the indicative meaning) at which fossil (core) sediments formed.

Two foraminiferal training sets have been employed. The first of these is a local training set based on contemporary (surface) distributions of foraminifera at Greatham Creek, previously developed by Horton (1997; 1999) and incorporates loss on ignition content and sand, silt and clay fractions, expressed as a percentage. The second is based on a national dataset (Horton and Edwards, 2006). This combines modern foraminiferal distributions from a wide range of sites (*n* = 15). Since sites with differing tidal ranges are combined to form this single, 'national' transfer function, it becomes necessary to standardise the altitudes of formation in order to take these differences in vertical range into consideration. This can be achieved by converting the altitude of formation of each sample to a standardised water level index (SWLI) (Horton, 1997; Horton and Edwards, 2006):

$$SWLI = \left(\frac{Alt_{ab} - MLWS}{MHWST_b - MLWST_b} \right) \times 100 \tag{8.3}$$

where:

Alt_{ab} = the altitude of sample *a* at site *b* (measured in m OD)
$MLWST_b$ = the mean low water spring tide level at site *b* (m OD) and
$MHWST_b$ = the mean high water spring tide level at site *b* (m OD).

Using the local hydrographic reference water levels in the Tees (Admiralty Tide Tables, 2005), SWLI can easily be converted to altitude *via* algebraic substitution and rearrangement of Equation 8.3 for altitude (Alt_{ab}).

Employing two foraminiferal transfer functions may initially seem superfluous. However, Horton and Edwards (2006) advocate the use a national training set because the varying physical, biological and hydrographic characteristics of each site, when combined, maximise the range of palaeoenvironments that can be reliably interpreted by the transfer function data (Horton and Edwards, 2006). Consequently, this decreases the occurrence of 'no analogue' situations, which occur when fossil assemblages have no modern equivalent in the training set; greater dissimilarity between fossil and modern assemblages forces the transfer function to extrapolate, rendering the reconstruction increasingly prone to error (Horton and Edwards, 2000). A 'no analogue' situation may result from the following (Horton and Edwards, 2006):

1. significant taphonomic modification of foraminiferal assemblages, such as the dissolution of calcareous foraminifera;
2. a reduction in the influence of relative elevation (a proxy for flooding frequency and duration) on species assemblages, possibly due to an increase in the relative importance of extremes in salinity or temperature on assemblages;
3. sediment reworking or a particularly mobile substrate.

In response to this claim, Allen and Haslett (2002) state that a transfer function based upon a national dataset may not perform well when applied to a single (local) core due to the importance of local conditions (particularly tidal range) on foraminiferal assemblages and significant differences between the faunal compositions of modern marshes and fossil cores. With both arguments in mind, it is considered necessary to consider both the local and national training sets to allow assessment of their relative performance in terms of reconstruction.

8.3.6 Transfer function performance

Typically, the coefficient of determination (r^2) between the observed and the predicted variable (altitude or SWLI) root mean square error of prediction are used to assess the predictive capacity of the transfer function (Horton and Edwards, 2006). Each regression model produces a number of 'components'. As the component number increases, so too does the mathematical complexity of the model. The choice of component for use in the transfer function depends on the relative values of the prediction statistics (r^2 and RMSEP) and the principle of parsimony (i.e. the lowest component that gives an acceptable model – Horton and Edwards, 2006). Accordingly, component two is used for each of the transfer functions based on the different training sets because it performs significantly better than component one, with only modest improvements thereafter (after Horton and Edwards, 2006). The prediction statistics of the two transfer functions employed are displayed numerically in Table 8.7. The performances of these models are depicted graphically in Figures 8.14 to 8.15.

Table 8.7 Prediction statistics of the two transfer functions employed in this thesis.

Model	r^2	RMSEP (m)
Local foraminiferal transfer function	0.84	0.22
National foraminiferal transfer function	0.71	0.22

The root mean squared error of prediction (RMSEP) and the r^2 correlation are based on the 'bootstrapping' cross-validation method. These statistics were computed using the C^2 computer program (Juggins, 2003). All models are based on Weighted Averaging-Partial Least Squares regression techniques using component 2 of the model output.

Both models give a satisfactory statistical performance, although the local foraminiferal transfer function provides the best predictive capacity (highest r^2) (Table 8.7). The lower r^2 in the national dataset is to be expected, given the range of environments that it describes.

Unsurprisingly, the scatter of data on the plot of observed versus predicted altitudes for the national transfer function (Figure 8.15) is greater than that of the local foraminiferal transfer function.

8.3.7 Assessing transfer function reliability using the Modern Analogue Technique (MAT)

Each reconstruction of altitude from fossil data must be assessed for its reliability in terms of its (dis-)similarity to the modern training set. Dissimilarity is calculated using the Modern Analogue Technique (MAT) feature in the C^2 computer program (Juggins, 2003). The squared chord distance was selected as the dissimilarity coefficient (Birks *et al.*, 1990; Edwards and Horton, 2000; Overpeck *et al.*, 1985; Prentice, 1980) since it maximises the signal to noise ratio when used with percentage data (Birks, 1995) (all foraminiferal and lithological data are entered into C^2 in this (%) format).

Deciding on whether a fossil sample has a 'good' or 'no close' modern analogue is undertaken by comparing the minimum dissimilarity coefficient (minDC) of a fossil sample to some threshold value. If the minDC of the fossil sample exceeds this value, the associated reconstruction is typically discarded. However, no consensus exists in the literature regarding this threshold value. Birks (1995), Birks *et al.* (1990) and Horton and Edwards (2005) take a dissimilarity coefficient below the minimum 10th percentile of dissimilarities calculated between all modern samples as a 'good' analogue between modern and fossil samples. This means that 10 % of the training set could be analogues for each other and that the remaining 90 % of contemporary samples represent different environments. Horton (1997) uses the 20th percentile as a cut off, and other values (2nd, 3rd and 5th percentiles) have also been employed. Indeed, MAT only provides a general guide to the relative similarities between fossil and modern data and that defining a single percentile value as the threshold between 'good' and 'no close' analogues is misleading, since values vary depending on the characteristics of the modern and fossil datasets and the ecological scale of analysis (Jackson and Williams, 2004). Accordingly, Woodroffe (2006) uses the largest minDC calculated between all modern samples as a general indication of whether a fossil sample has a good contemporary analogue in the training set. The same reasoning is employed here; any reconstructions with a minimum dissimilarity coefficient higher than the largest value for the modern data set are treated with extreme caution. These statistics are shown in Table 8.8 and 8.9.

Encouragingly, all reconstructions from the low marsh core have good modern analogues (Table 8.8). In contrast, the reconstructions obtained from the local foraminiferal transfer function in the mudflat core possess no modern analogues apart from at 0.36 m depth (Table 8.9). Accordingly, reconstructions from the mudflat core that are based on the local foraminiferal transfer function are treated as potentially erroneous. The value of the national foraminiferal transfer function in reducing the number of no analogue situations is highlighted here; all samples have good modern analogues when calibrated with this larger dataset.

Table 8.8 Minimum dissimilarity coefficients for calibrated samples from the low marsh core. Any values in **bold** indicate that the minimum dissimilarity coefficient value is greater than the highest minimum dissimilarity value for the modern training set, used as a threshold for modern analogues. Such samples should be treated cautiously.

	National foraminiferal transfer function	Local foraminiferal transfer function
	Highest MinDC value in training set = 46.40	Highest MinDC value in training set = 24.83
	Minimum dissimilarity coefficient	
Depth in core (m)		
0.00	2.58	3.83
0.01	3.90	5.98
0.02	0.36	15.35
0.04	0.01	4.76
0.06	0.20	1.60
0.08	1.85	3.05
0.10	0	0
0.13	0	0
0.14	2.31	2.31
0.16	3.14	10.27
0.18	1.13	5.09
0.20	2.05	4.30
0.22	0.70	7.99
0.24	1.61	4.37
0.26	3.09	4.27
0.28	5.22	5.38
0.30	2.53	5.31
0.32	1.81	1.81
0.34	6.57	11.78
0.36	0.06	1.98
0.38	0.61	4.47

Table 8.9 Minimum dissimilarity coefficients for calibrated samples from the mudflat core. Values in **bold** indicate that the minimum dissimilarity coefficient value is greater than the highest minimum dissimilarity value for the modern training set, used as a threshold for modern analogues. Such samples should be treated cautiously.

	National foraminiferal transfer function	Local foraminiferal transfer function
	Highest MinDC value in training set = 46.40	Highest MinDC value in training set = 24.83
	Minimum dissimilarity coefficient	
Depth in core (m)		
0.00	9.43	**25.06**
0.04	22.97	**37.82**
0.08	27.33	**34.60**
0.12	26.09	**26.09**
0.16	22.10	**32.22**
0.20	26.72	**33.31**
0.22	23.54	**30.13**
0.24	25.70	**31.63**
0.28	18.21	**30.32**
0.32	27.29	**31.94**
0.36	19.53	23.76
0.40	21.61	**30.56**
0.44	29.63	**30.28**

8.3.8 Variations in relative marsh elevation above mean sea level

By plotting depth in the core against reconstructed palaeo-surface elevation relative to mean sea level, the multiple foraminiferal samples collected from the low marsh (Figure 8.16) and mudflat (Figure 8.17) cores can be used to produce a sequence of palaeo-surface elevation change. Before constructing a sea level curve, it is necessary to analyse these plots in order to determine the accuracy of the transfer function reconstructions.

Also marked on the palaeo-surface elevation diagrams (Figures 8.16 and 8.17) are the contemporary elevations relative to mean sea level at which the sediments were sampled (+ 1.91 m in the low marsh and + 0.71 m in the mudflat). By comparing the predicted elevation of these near-surface sediments to the known contemporary elevation, it is possible to determine whether the transfer function is performing with sufficient accuracy.

The agglutinated saltmarsh foraminiferal assemblages of the low marsh core are consistent with its organic lithology. However, the reconstructions from the low marsh core that are based on the national training set display decreased sensitivity to variations in foraminiferal assemblages, since reconstructed elevations are largely constant throughout the depth of the core at *c.* 2.3 m above mean sea level. However, analysis of the foraminiferal assemblages in Section 8.2.1 and comparison with the contemporary distributions (Horton and Edwards, 2000) suggests that the observed variations in foraminiferal content should reflect greater changes in relative elevation. Reconstructions using the near-surface samples based on the national training set also fail to predict the contemporary elevation. As a consequence of this, confidence in the reliability of the 'downcore' reconstructions of elevation is low. The reconstructions of the low marsh foraminiferal assemblages based on the local transfer function perform better, displaying a greater sensitivity to variations in foraminiferal assemblages.

The reconstructions obtained from the foraminiferal assemblages contained within the mudflat core do not seem to be reliable. It has already been noted that the reconstructions calculated from the local training set are potentially erroneous due to a lack of modern analogues for foraminiferal assemblages in the core. Reconstructions from the near surface sediments based on both training sets fail to successfully predict the contemporary elevation; indeed, predictions are *c.* 1.2 m too high from the local dataset and *c.* 1.5 m too high from the national training set. Furthermore, the elevations that are being predicted correspond to contemporary altitudes that are characterised by saltmarsh growth. Hence, the lithostratigraphy within the core does not corroborate the statistical estimation of formed elevation.

These problems are common and are to be expected. The value of mudflat foraminiferal assemblages in providing high-resolution records of relative elevation change is limited due to poor zonation of microfossils relative to the intertidal frame. As elevation relative to the tidal frame decreases, the driving forces behind vertical zonation (i.e. magnitude and frequency of flooding duration) weaken and complicating factors such as sediment erosion, transport and redeposition will increase due to increased tidal energy (Horton and Edwards, 2006). In intertidal sub-environments located at elevations lower than the pioneer marsh (i.e. mud- and sand-flats), sediment mobility therefore results in assemblages consisting or both allochthonous and autochthonous components and broader vertical ranges occupied by microfossil (particularly foraminiferal) species (Horton and Edwards, 2006). Although a relationship between mudflat microfossil assemblages does usually still exist, any reconstructions that are made based on such assemblages will lack the high precision of those based upon the tightly constrained assemblages that are characteristic of saltmarsh environments.

8.3.9 Age-depth modelling using radioisotopes

Chronologies for the low marsh and mudflat cores are based on two radionuclides with short half-lives: ^{210}Pb and ^{137}Cs (Section 4.3.7). All isotopic data are presented in Appendix II.

Following subtraction of the background ('supported') component (0.020 Bq/g in the low marsh core and 0.017 Bq/g in the mudflat core, as determined from gamma spectrometric analysis of ^{226}Ra activities) from the total ^{210}Pb activity in each core, depth profiles of the excess (unsupported) ^{210}Pb activity can be plotted (Figures 8.18 and 8.19). Despite an anomalous peak at 0.16 m, a relatively log-linear decrease of excess ^{210}Pb activity with depth is evident in the mudflat core (Figure 8.19), indicating a relatively constant rate of accumulation through time. There is a larger distortion to the 'ideal' log-linear decay of excess ^{210}Pb activity with depth in the low marsh core between 0.04 and 0.12 m depth (Figure 8.18). Such dramatic changes in the slope of the profile are common (Cundy and Croudace, 1996; Olsson, 1986) and can be attributed to bioturbation, physical mixing or redox-driven migration of the ^{210}Pb isotope. The latter process in particular has been shown to be common in saltmarsh environments (Cundy and Croudace, 1995); ^{210}Pb is known to be prone to mobilisation following reduction of carrier phases in the anoxic zone beneath the water table and its re-precipitation with hydrous Fe and Mn oxyhydroxides in the oxic layer above the redox boundary. This process does not necessarily preclude the use of ^{210}Pb as a useful chronological tool, providing that peaks arising from redox mobility and concentration are identified and eliminated by comparison with other redox sensitive data (Cundy and Croudace, 1995).

By comparison with the depth distributions of Fe_2O_3, MnO and S in each core, an assessment can be made regarding the possibility of diagenetic remobilisation of ^{210}Pb (Figures 8.20 and 8.21). Although redox zonation was not found to be particularly significant in the low marsh core (Section 7.1), the observed enrichment of ^{210}Pb in the low marsh at 0.04 m is coincident with a pronounced peak in MnO. Furthermore, the depth distribution of S is similar to that of the ^{210}Pb, suggesting a potential redox or lithological control on the distribution of the ^{210}Pb isotope. The depth distribution of ^{210}Pb in the mudflat core does not seem to be related to any of the three dominant participants in redox processes, suggesting that redox-driven mobility of ^{210}Pb has been limited. Hence, the minor peak in excess ^{210}Pb may reflect bioturbation or physical mixing of the sediment.

Following comparison with redox-sensitive geochemical data, data points that display significant enrichment in ^{210}Pb due to redox processes or bioturbation have been removed from subsequent analysis (after Cundy and Croudace, 1995). Using these profiles, sediment accumulation rates were determined according to the 'Simple' model of Robbins (1978). Here, the sedimentation rate is given by the slope of the least squares fit for the natural logarithm of the excess ^{210}Pb activity plotted against depth. The radioactive decay equation states that:

$$A = A_0 e^{-\lambda t} \tag{8.4}$$

where A is the activity at depth x, A_0 is the initial activity, λ is the radioactive decay constant for ^{210}Pb (= 0.0309) and t is time. In addition, the sediment accretion rate is

$$s = \frac{x}{t} \tag{8.5}$$

where s is the accumulation rate (mm yr^{-1}), t is time in years (before 2004) and x is depth (m or mm). Taking natural logarithms to solve the radioactive decay equation and substituting $t = z/s$ gives:

$$\ln A_z = \ln A_o - \left(\frac{\lambda}{s}\right) z \tag{8.6}$$

Hence, when excess $ln^{210}Pb$ (equivalent to $\ln A$) is plotted against depth, an approximately straight line is produced. Fitting a least-squares regression line to the data gives an algebraic expression of the form $y = mx + c$, where lnA_0 is the intercept (constant) c and λ/s is the gradient m. Therefore, the sediment accretion rate can be calculated from the following equation:

$$s = 0.0309/m \tag{8.7}$$

Using this procedure, the best estimate sediment accretion rate in the low marsh core is 3.3 mm yr^{-1} (95 % confidence interval = 2.7 - 4.4 mm yr^{-1}). In the mudflat core, sediment is estimated to be accreting at a rate of 4.4 mm yr^{-1} (95 % confidence interval = 3.5 - 5.8 mm yr^{-1}). By dividing the depth in the core by each estimate of the accretion rate, the number of years prior to that of core collection (2004) is calculated. This allows a year of deposition to be assigned to particular depths in each core (Figures 8.22 and 8.23).

A combination of dating tools is employed in accordance with Smith (2001), who states that an uncritical use of ^{210}Pb age-depth models has increased in recent years and that corroboration of such chronologies must be performed with at least one independent technique that separately provides an unambiguous chronostratigraphic horizon. Due to the diagenetic alteration of the exponential ^{210}Pb decay profile and the removal of spurious data points, such a validation of the ^{210}Pb chronology becomes even more important. ^{137}Cs is a particularly useful radionuclide because it is very strongly sorbed onto clay particles and hence is virtually non-exchangeable during shifts from oxidising to reducing conditions (and *vice versa*) (Cundy and Croudace, 1995).

^{137}Cs profiles for the low marsh and mudflat are displayed in Figure 8.24. In the low marsh core, marker horizons are found at depths of 0.14 m (corresponding to the 1963 weapons testing peak) and 0.07 m (corresponding to the ~1980 Sellafield discharge maximum). From equation 8.5, accretion rates of 3.4 mm yr^{-1} and 3.3 mm yr^{-1} are obtained. These rates are in very close agreement with those calculated using the ^{210}Pb age-depth model.

It is more difficult to draw conclusions regarding chronology from the ^{137}Cs profile of the mudflat core, since the maximum peak occurs at a very shallow depth (*c.* 0.04 m). This would suggest a sedimentation rate of 1.9 mm yr^{-1}. This is significantly below even the lower estimate of accretion rate derived from ^{210}Pb. Furthermore, accretion rates can be expected to be inversely proportional to elevation relative to the intertidal frame as a result of increased sediment delivery lower in the intertidal frame due to the increased frequency and duration of tidal inundation (Allen, 1990). Hence, given the lower altitude of the mudflat core relative to that of the low marsh, the accretion rate can be expected to he higher in the tidal flat.

The cause of the unusual ^{137}Cs profile could be attributed to a number of processes, such as bioturbation, sediment reworking or erosion. It is also evident from Figure 8.24 that ^{137}Cs activities are low in relation to those measured in the low marsh core and so may reflect lithological issues, such as lower clay and organic contents in the mudflat which may provide reactive sites for adsorption. However, if the introduction of ^{137}Cs into the system (0.24 m) is dated to 1954 (the date at which large scale introduction of ^{137}Cs into the atmosphere occurred; Plater and Appleby, 2004) then an accretion rate of 4.8 mm yr^{-1}

is calculated – a rate not too dissimilar to the best-estimate obtained from the ^{210}Pb model. However, it must be noted that the calculation based on the first 'appearance' of ^{137}Cs will give a maximum accretion rate, since downward mixing and/or diffusion of the radioisotope would lead to its presence at depths corresponding to ages older than 1954 (Long *et al.*, 2002).

There is therefore generally good agreement between the accretion rates/chronologies obtained from both radionuclides and so the age-depth models presented in Figures 8.22 and 8.23 are deemed to be reliable. Encouragingly, the accretion rates compare favourably with those of Plater *et al.* (2000; 1998), which are also based on radioisotope chronologies. Plater *et al.* (1998; 2000) obtained accretion rates of 3.74 ± 2.65 mm yr^{-1}, 2.78 ± 1.03 mm yr^{-1} and 3.29 ± 0.87 mm yr^{-1} from shallow cores samples from altitudes of 2.50 m OD, 2.00 m OD and 1.84 m OD respectively. However, due to the unusual mudflat ^{137}Cs profile, the mudflat chronology should be treated with extreme caution.

8.3.10 Sediment decompaction procedure

The decompaction routine reconstructs the previous stress conditions experienced throughout a given stratigraphic column (or section thereof) and, using the compression models developed previously, calculates values of voids ratios (a volumetric parameter) that existed prior to the deposition of overlying material. Providing both the *in situ* (compacted) thickness and voids ratio of a layer are known, the previous thickness can also be calculated using the following equation:

$$z_1 = z_2\left(\frac{1+e_1}{1+e_2}\right) \tag{8.8}$$

where e_2 is the *in situ* compacted (fossil) thickness of a layer of near-uniform lithology and geotechnical properties and z_2 is the measured thickness of that layer; e_1 is the estimated (model-derived) decompacted voids ratio of the layer following the simulated 'removal' of the overburden layers and z_1 is the corresponding thickness of that layer (Figure 8.25). This equation is based on the fact that 1 + *e* (termed the specific volume) is proportionately related to volume (Powrie, 2004).

Due to variations in lithology and structure, it is necessary to divide the stratigraphic column into a number of layers, within which lithological and geotechnical properties are constant. Each layer is then treated separately; the individual compression/volumetric behaviour in each layer is treated independently. The results for each layer are then summed to provide an estimate of total compression since overburden sedimentation and, hence, the altitudinal correction factor.

Following geotechnical testing of the materials in the stratigraphic sequence of interest, and using the method described generally above, the decompaction routine used in this section involves a number of steps:

1. the stratigraphic column is divided into sub-layers, the number of which depends on the resolution of the sampling technique employed. Whilst the x-ray technique provides very high resolution measurements of layer thicknesses (sub-mm), the constituent decompaction calculations would place considerable demands computational time and capacity. Indeed, in a relatively short 0.50 m core, 500 layers would require consideration and the number of calculations cannot be undertaken in Microsoft Excel due to limitations in the size of the spreadsheets. A 0.01 m layer thickness resolution is chosen. Within each layer, mean values of voids ratio are calculated. Any variations in structure and compression behaviour in each of these layers is considered and accounted for by the statistical modelling approach.
2. for each layer, values of the following parameters are designated on the basis of lithology, compression testing and changepoint regression modelling:
   a. depth in the core (m);
   b. layer thickness (m);
   c. e_2 (*in situ*, compacted voids ratio);
   d. σ', the effective stress at the midpoint of the layer;
   e. C_r, the modelled recompression index;
   f. C_c, the modelled compression index;
   g. x_c, the modelled changepoint;
   h. α, the modelled 'alpha' value required by the regression equation;
   i. the standard errors associated with C_r and C_c.
3. The next stage involves calculations of the effective stress that existed at each level below the depositional surface that occurred prior to the deposition of overburden sediments. This involves subtracting the effective stress at the level of,

for example, a sea level index point from that found in each underlying layer under consideration.

4. Common logarithms of these 'previous' values of effective stress must be calculated before they can be used in the changepoint regression models.
5. Using the changepoint regression models and the layer-specific input values, the voids ratio within each layer prior to the deposition of overburden sediments is estimated.
6. Using Equation 8.8, the 'previous' thickness of each layer prior to overburden sedimentation can be calculated.
7. In order to calculate changes in individual layer thicknesses as a result of overburden sedimentation, the *in situ* layer thickness (z_2) is subtracted from the decompacted thickness (z_1).
8. To calculate the total correction for each sea level index point (i.e. when each depth throughout the core was at the depositional surface and was not buried by overlying sediments), the individual layer corrections are simply summed. This provides the altitude correction which should be added to the *in situ* altitude of a sea level index point to return to the altitude at which it was deposited.

The procedure described above provides the best-estimate of the amount of compression experienced by a sediment column. In order to calculate the errors associated with the thickness changes, the following additional stages are required:

9. Following Stage 5 above, the upper and lower 95 % prediction limits are calculated from the mean value of estimated voids ratio by adding 1.96 $\times$ the relevant standard error of the estimate.
10. Using Equation 8.8, the maximum and minimum thicknesses of each individual layer can be estimated from these upper and lower voids ratio predictions.
11. In order to calculate the upper and lower predictions for thickness changes, the *in situ* thickness is then subtracted from the upper thickness value and the lower thickness value is subtracted from the *in situ* thickness.
12. In order to convert these thickness changes as ± errors relative to the mean prediction, the mean thickness change estimate is subtracted from the upper thickness change estimate for each layer. Similarly, the lower thickness change estimate is subtracted from the mean thickness change estimate for each layer.
13. To prevent unhelpful amplification of the error term with increasing height from the base of each core, the upper and lower thickness change estimates for each layer

are squared. These values are then summed and a square root is taken, expressing the cumulative error terms as root squared errors (RSEs).

14. To obtain the predicted upper and lower limits of the total thickness correction, the upper and lower RSEs are added and subtracted from the mean predictions.

For each material, two different rheological models are used in the decompaction routines. The first type of model employed is the changepoint regression models developed previously in Section 8.1.3 on the basis of the empirical research undertaken in this thesis. The second type of model for each material is based on Terzaghi's compression law and involves a backwards extrapolation of the normal compression line to the minimum value of effective stress observed in the core. Terzaghi's compression law model can still be applied using the changepoint regression framework; it simply requires replacing the value of C_r with that of C_c. To allow a more equal, unbiased comparison between models, error terms are also applied to the Terzaghi model. The relevant parameters for each model and material are presented in Table 8.10. Models are compared graphically in Figures 8.26, 8.27 and 8.29.

Table 8.10 Performance of models and values of regression parameters of the changepoint regression models used in the decompaction procedure. S.E.E. 1 = standard error of the estimate at effective stresses less than the yield stress/changepoint. S.E.E. 2 = standard error of the estimate at effective stresses greater than the yield stress/changepoint.

Material/model	r^2	S.E.E. 1	S.E.E. 2	x_c	Yield stress (kPa)	α	C_r	C_c
Low marsh 2	0.90	0.68	0.25	0.91	8.13	4.52	-0.12	-1.61
Low marsh 3 (Terzaghi)	0.85	0.25	0.25	1.10	-	4.18	-1.61	-1.61
Mudflat 1	0.81	0.24	0.24	0.72	5.22	2.24	-0.01	-0.61
Mudflat 2 (Terzaghi)	0.78	0.24	0.24	0.72	-	2.24	-0.61	-0.61
Saltmarsh 1	0.96	0.48	0.15	0.92	8.39	2.92	-0.09	-0.96
Saltmarsh 2 (Terzaghi)	0.88	0.15	0.15	1.11	-	2.77	-0.98	-0.98

Mean, upper and lower estimates of the vertical corrections are displayed for the low marsh and mudflat cores in Figures 8.29 to 8.32. For each depth at which the *in situ* (compacted) sea level index points lies, the relevant correction for autocompaction can be obtained from these graphs. The graphical form of the decompaction correction curves depends not only on the input parameters; it is also dependent upon values of *in situ* voids ratios and whether these are greater or less than the modelled best-estimate at a given effective stress. The statistical nature of the model, however, ensures that the full range of voids ratio that have been observed at a given effective stress are described in the procedure.

For each core, decompaction corrections are expectedly greatest when based upon Terzaghi's compression law. In the low marsh, the maximum correction required is *c.* 0.16 m in the surface layer (Figure 8.30). This contrasts with the predictions of the 'overconsolidated' models, where the maximum correction is 0.02 m, required at a depth of 0.25 m (Figure 8.29). A similar pattern occurs in the mudflat cores; the maximum correction using Terzaghi's compression law is 0.11 m at the surface (Figure 8.32). Using the Mudflat compression models that considers early overconsolidation, the maximum correction is just 0.04 m (Figure 8.31).

The fact that greatest correction is required in the uppermost layer using Terzaghi's compression law reflects the cumulative compression in all underlying layers and the fact that the *in situ* voids ratios are considerably below the model predictions, hence requiring larger corrections.

8.3.11 Sea level reconstruction

It was noted in Section 8.3.3 that validation of the autocompaction models in a stratigraphic context is very much reliant on the quality of the sea level indicators contained within the low marsh and mudflat cores. Accordingly, sea level reconstructions and validation of autocompaction models will be based upon the local foraminiferal training set. The foraminiferal assemblages in the mudflat core and the transfer functions employed to calibrate them for relative elevation changes do not provide a sufficiently accurate record for direct application. Indeed, the dubious conclusions drawn from the ^{137}Cs profile also suggest that the resultant age-depth model may be erroneous and hence would not allow model validation.

Reconstructing variations in mean sea level is a simple matter of subtracting the elevation of the palaeo-surface relative to mean sea level (obtained from the transfer function) from the contemporary altitude of the sample. Through use of the age-depth model for the low marsh developed above, the obtained altitudes can be placed into a chronological framework. Age-altitude graphs can then be plotted. In order to facilitate ease of comparison between the high-resolution record of the tide gauge data and that provided by the low marsh geological record, from which the average resolution of the foraminiferal record is six years, a six year running mean was calculated from the North Shields tide gauge record. The reconstructed *in situ* (compacted) sea level curve obtained from the low marsh is plotted with the 'smoothed' tide gauge record in Figure 8.33. The error term describes the sample-specific RMSEP of the transfer function (*c.* 0.20 m) and the levelling error (0.016 m). When combined as a root-squared error, the error term is *c.* 0.22 m.

It is immediately obvious that the geological record displays a *c.* 0.40 m negative offset from the instrumental record. This was expected on the basis of the autocompaction of the underlying strata. An additional contribution may also result from a tidal damping effect. The narrow inlet that separates Greatham Creek from Seal Sands (Figure 4.1) may result in a restriction of the in- and out-flow of tidal flood waters. Such impeded tidal ventilation is known to alter the tidal flooding characteristics compared to those experienced in open marine conditions (Bird, 2001; van der Molen, 1997). Tidal amplitudes are typically diminished, mean sea level is suppressed and hence this phenomenon may account for the lower value of mean sea level recorded by the low marsh core. In any case, it is the rate of relative sea level change that is of interest in this section of the investigation.

The geologically reconstructed rate of sea level prior to the addition of the decompaction correction factors is 2 mm yr^{-1} (Figure 8.33). This compares favourably with the long-term rate observed in the tide gauge data (1.9 mm yr^{-1}). Indeed, some of the decadal scale changes are also reproduced by the reconstruction.

When decompacted using the autocompaction model that is based on Terzaghi's compression law, the relative position of the sea level index points changes. Little relative movement of index points occurs at the bottom of the core section considered; conversely, larger corrections occur towards the top of the core. As a result, the decompacted rate of relative sea level rise increases to 3 mm yr^{-1} (Figure 8.34) - a considerable increase in comparison to the 1.9 mm yr^{-1} recorded by the instrumental record.

In contrast, when decompacted using the autocompaction models developed in this thesis that consider low stress overconsolidation, the relative position of the sea level index points remains largely unchanged. As a result, the *in situ* and decompacted rates also remain largely unchanged (Figure 8.35). Indeed, the decompacted rate now exactly matches that obtained from the instrumental record. The root-squared error term now also considers the upper and lower estimates of the decompaction corrections that are required at each level.

On the basis of this evidence, it seems that the autocompaction models developed in this thesis provide the most accurate predictive capacity. The overconsolidation observed at low stresses and described by these models results in very little relative movement between sea level index points, provided that the part of the record of interest is contained within this overconsolidated zone. As a result, rates of relative sea level change obtained from the overconsolidated section of mineralogenic stratigraphies do not seem to require a correction for autocompaction. Furthermore, the increase in the rate of relative sea level change that is calculated following decompaction of the sediments using Terzaghi's compression law highlights the need to undertake detailed geotechnical investigation of the autocompaction behaviour of the sediments contained within a given stratigraphic section. Only following empirical input can a model be expected to accurately describe autocompaction within intertidal stratigraphies.

Although the mudflat autocompaction models could not be validated in a practical way due to the poor quality of the reference water levels reconstructed from the sea level indicators within the core, the overconsolidation described by the models suggests that similar findings would be obtained.

8.3.12 Implications of the decompaction of near-surface sediments

The results obtained from the decompaction approaches offer potentially useful implications for investigations into near surface sediments. These sediments are of particular interest because they contain sea level data that can potentially link instrumental records from the last 100 years with geological records from the late Holocene (Gehrels *et al.*, 2002). This period is critical to an understanding of the potential effects of global warming on sea levels. Tide gauge databases generally record a faster rate of sea level rise than is suggested by geological records obtained from the mid Holocene (Church *et al.*, 2001; Shennan and Horton, 2002; Woodworth *et al.*, 1999). However, since sea level index points from the late Holocene are generally few in number, it is difficult to determine when this acceleration in 'eustatic' sea level actually began. If it can be demonstrated that a worldwide acceleration in 'eustatic' sea level signal began in the middle of the nineteenth century, it becomes more likely that they have resulted from the increase in greenhouse gas emission associated with the Industrial Revolution (Gehrels *et al.*, 2002). As a consequence, a number of recent studies have attempted to use near-surface saltmarsh records of sea level change to date the onset of this acceleration in sea level rise (e.g. Gehrels *et al.*, 2002; Gehrels *et al.*, 2005). However, such records rest critically on the assumption that autocompaction has not altered the recorded rate of sea level change.

The fact that the *in situ* rate of sea level rise matches that observed in the compaction-free tide gauge record (as a result of overconsolidation of near-surface sediments) suggests that relative sea level records obtained from mudflat and mineralogenic low marsh sediments do not require a decompaction correction providing:

1. the same overconsolidating mechanisms are occurring in a given intertidal zone;
2. the historical period/timeframe of interest is contained within the overconsolidated section of the sedimentary profile (i.e. provided that the yield and preconsolidation stresses associated with surface overconsolidation have not been exceeded by overburden sedimentation).

Gehrels (2000) obtained similar rates from a saltmarsh-derived record in Maine, USA and a local tide gauge. He found a strong correlation between the geological and instrumental records (r = 0.91, p = 0.005) without applying a geotechnical correction for autocompaction. However, the geological reconstruction is based upon microfossils from the highly organogenic saltmarsh sediments that form on the east coast of the USA. Therefore, it could be possible that the same overconsolidating mechanisms are operating in the peat facies of Scarboro and Machiasport in Maine, preventing the need for a geotechnical correction in near-surface sediments. Whether this is correct or not, it raises an important question regarding the transferability of the autocompaction models developed in this thesis.

8.4 MODEL TRANSFERABILITY

Despite the improvement to our understanding of the autocompaction behaviour of these sediments, it is clear that a variety of different mineralogenic and organogenic materials exist within the intertidal zone that have not been explicitly addressed in a comprehensive manner. It is therefore important to future research to speculate on the transferability of the models that have been developed on the basis of the research undertaken in this thesis.

8.4.1 Transferability within the intertidal zone

In order to consider the potential for transfer of the autocompaction models developed in this thesis, it is necessary to consider them conceptually at a basic level. The model consists of four main components:

1. the initial voids ratio, e_i;
2. the preconsolidation and yield stresses;
3. the recompression index, C_r;
4. the compression index, C_c.

Even for the two lithologies studied in this thesis, values of each of these components have been shown to vary. Indeed, the material and structural properties observed in these lithologies are likely to form part of a continuum of material property variation throughout the intertidal zone that is essentially driven by differences in relative elevation, which controls (and is also controlled by) flooding frequency and duration. Within the intertidal zone, general variations in lithology with elevation result from the intertidal energy gradient, with coarser (sandy silts) deposited at lower elevations where tidal energy is higher. With increased elevation, tidal energy diminishes and this is accompanied by a decrease in particle size. In these wholly mineralogenic environments, initial voids ratios are also determined by depositional factors such as water chemistry (which affects floc size) and flow velocity (e.g. Burland, 1990; Lintern, 2003; Sills, 1998). More dramatic changes in lithology occur between MHWNT and MHWST, where salinity and flooding conditions permit colonisation of mudflat deposits by halophytic saltmarsh vegetation. The mineralogenic component decreases as HAT is approached due and the increase in *in situ* organic growth and the decreased opportunity for advected sediment deposition associated with a reduction in duration and frequency of tidal flooding (Allen, 1990). Each lithology has its own distinctive initial structure and subsequent compression behaviour. Indeed, the greater the initial voids ratio, the more prone the material to compression (Figure 8.8). Even small increases in organic content resulting from autogenic organic growth in the low marsh are sufficient to significantly increase the initial voids ratio and compression indices of granulometrically identical sediments (Figures 8.9 and 8.10; Section 5.3).

In addition to affecting the elevational distribution of lithologies, flooding frequency and duration also play a considerable role in modifying these lithology-specific structural and compression 'templates'. During times of subaerial exposure, a number of processes result in increases in effective stress in the intertidal sediments. In particular, desiccation can cause significant reductions in moisture content and subsequence volumetric reductions (Fredlund and Rahardjo, 1993a; 1993b). Similarly, a reduction in the groundwater level can also result increased effective stress *via* capillary suction stresses (Head, 1988; Powrie, 2004). These increases in effective stress at the depositional surface are manifested as increases in the preconsolidation and yield stress. In addition, the cyclicity of effective stress as a result of tidally-driven variations in water loading and groundwater levels have been shown to minimise the operation of creep processes which could, under static effective stress conditions, lead to continuous and significant volumetric deformation.

As a combined result of these processes, variations in values of initial voids ratio, the recompression index, yield stress and compression index can be expected to occur throughout the intertidal zone. At subtidal levels, where the geotechnical environment is considerably less dynamic and where subaerial exposure and groundwater variations do not occur by definition, the lower end member of the continuum exists. Overconsolidation is unlikely here and so the compression behaviour in $e\log_{10}\sigma'$ space is described by Terzaghi's compression law (Figure 8.36, graph 1). Within the fully mineralogenic section of the intertidal zone, similar materials form but their compression stress paths are modified by subaerial desiccation and groundwater variations. Hence, with increased elevation above LAT, the yield stress can be expected to increase (Figure 8.36, graphs 2 and 3). In the partially organic low marsh environment, the *in situ* production of organic matter leads to an increase in initial voids ratio which, in turn, leads to steeper recompression and normal compression lines. Furthermore, the greater frequency and duration of subaerial exposure increases the yield stress (Figure 8.36, graph 4).

Each lithology at each elevation therefore has its own 'signature' compression behaviour and although careful calibration of each material is required to determine the exact values of each of the model components, the basic autocompaction model developed within this thesis can be applied to these mineralogenic sediments. In mid-marsh environments, where mineralogenic sedimentation still contributes to lithological make-up, it is possible that the material property continuum continues, with higher organic contents creating more open, compression prone structures with higher initial voids ratios, compression indices and yield stresses (Figure 8.36, graph 5). However, towards the higher elevations within the intertidal zone, the application of these models is likely to be more uncertain due to higher organic contents; the autocompaction behaviour of organogenic materials is considerably more complex than those of mineralogenic materials (Allen, 2000a). The lack of empirical research into the environmental conditions in mid- and high-marshes and the deformation behaviour of the materials that form here means that it is only possible to speculate on how these materials autocompact (Figure 8.36, graph 6):

Due to their location at the landward edge of marine influence, conditions are increasingly terrestrial and the low duration and frequency of tidal flooding reduces environmental dynamism. Hence, it is hypothesised that the preconsolidation and yield stresses are considerably higher than those experienced in even the lower marsh environments due to prolonged exposure to desiccating subaerial conditions. Alternatively, the accumulating mass of decaying marsh vegetation may act as a 'protective' covering from such conditions. This would potentially reduce desiccation and hence estimates of preconsolidation and yield stresses become increasingly uncertain without direct investigation. Due to the increased distance inland, the amplitude of fluctuating groundwater can be expected to be significantly diminished. As a result, variations in effective stress will be less dynamic and this may permit creep processes to act in these sediments. Critically, processes of biological and chemical decay may provide the dominant controls on volumetric reduction. In essence, a greater number of independent variables would be required to predict structural and volumetric changes in high marsh sediments.

Application to organogenic sediments of the autocompaction models developed on the basis of empirical research into mineralogenic sediments may be highly erroneous. An upper limit of their application must exist; further research is required to determine the exact elevation and/or material property characteristics of this limit.

8.4.2 Temporal and spatial transferability

The intertidal zone at Greatham Creek is just one example of the range of diverse intertidal environments found within low energy estuarine sites throughout the world. As a result of the interaction between local geomorphological, hydrographic, sedimentological and climatic conditions, the exact nature of intertidal landforms varies both spatially and temporally. Consequently, application of the autocompaction models developed on sediments at Greatham Creek may not necessarily be valid elsewhere.

In terms of the predominantly mineralogenic intertidal systems of contemporary Northwest Europe (Allen, 2000b), variations in tidal range may lead to subtle changes in the degree of overconsolidation experienced at different elevations. In the larger estuarine systems of southwest Britain, including that of the River Severn (extreme tidal range of 14.8 m; Allen, 2000), overconsolidation towards HAT may increase significantly due to desiccation during summer neap tides and as groundwater levels fall. In contrast, a smaller tidal range may result in a lesser degree of overconsolidation; surface sediments are likely to be in closer proximity to tidal-controlled and recharged groundwater, allowing continued saturation of the near-surface sediments *via* capillary action.

In larger sedimentary systems such as the Ganges – Brahmaptura delta, India/Bangladesh, South Asia, sedimentation rates are significantly higher than those generally observed in contemporary UK saltmarshes as a result of rapid tectonically-driven relative sea level rise (Goodbred and Kuehl, 2000). Here, Allison and Kepple (2001) obtain a recent sedimentation rate of 11 mm yr^{-1} on the basis of ^{137}Cs dating. It is possible that these high rates of process activity reduce the opportunity for neap tide and seasonal drying and overconsolidation, since previously deposited sediments are covered by more recent sediments before they can dry out and undergo effective stress increases. Similar conclusions can be drawn regarding variations in sedimentation rate through time which result from differences in marsh maturity at a particular location. Sedimentation rates on recently initiated saltmarshes that are in their 'youthful' stages have been shown both empirically (French, 1993; Pethick, 1981) and *via* numerical experiments to be greater than those on 'mature' marshes (Allen, 1990). Hence, in their early stages, intertidal sediments may be expected to be less overconsolidated than their 'mature' counterparts that are of the same lithology.

More significant changes to the overall compression behaviour of saltmarsh sediments may occur as a result of inter-estuary variations in sedimentology. Crooks (1999) and Crooks and Pye (2000) found that saltmarshes in sites at Northey Island and North Fambridge, Essex contain low quantities of calcium carbonate (typically less than 1 %). These sediments are prone to deflocculation when submerged in non-sodic meteoric waters (Section 3.9.9). This can result in the formation of an aquiclude which hinders sub-surface drainage, and so sediments forming above these overconsolidation barriers have high moisture contents and voids ratios. The calcium-rich soils of the Severn Estuary do not disperse, and so a denser initial structure is created. Hence, the geochemical properties of sediments in various locations are likely to create different initial structures and subsequent compression behaviours.

Despite an obvious requirement to obtain site-specific estimates of the compression indices and soil structural values, application of the model to different northwest European mineralogenic sediments is still likely to be valid providing the general site and lithological characteristics are similar. However, as the spatial and temporal framework of consideration is further expanded, a greater number of environments and processes that are not explicitly considered by the conceptual model outlined above are encountered.

Again, the predominance of the organic component in coastal intertidal soils in different areas around the globe prohibits valid model application to these areas, such as North America or Ireland (Allen, 2000), for reasons involving decay of organic matter and an associated structural collapse. Such processes have been observed in both contemporary sediments (e.g. Delaune *et al.*, 1994) and their influence is strikingly evident within the stratigraphic record (Long *et al.*, 2006).

In lower latitude mangrove environments, climatic influences may lead to considerable departures from the general elevational trends in compression behaviour that are likely to occur in temperate areas. Not only will the increased temperatures increase evapotranspiration from the soil, leading to increased preconsolidation and yield stresses, but considerable potential for unsaturation may occur. Unsaturated tests were undertaken in the material testing program of this study, but the degree of unsaturation was not sufficient to cause any observable change in compression behaviour. In highly unsaturated intertidal soils in tropical climates, a fundamental change in soil compression behaviour may occur (Fredlund and Rahardjo, 1993a; 1993b).

8.4.3 The influence of extreme events

High magnitude, low frequency events can also have a significant influence on coastal sediment autocompaction. For example, Cahoon *et al.* (2003) undertook analysis of SET-based sediment elevation and marker horizon-based accretion dynamics in the mangrove forests of Honduras that were directly impacted by Hurricane Mitch (1998). They concluded that the mass tree mortality that resulted from high wind speed impact lead to a collapse of the peat substrate due to organic decomposition and no recovery or re-colonisation of the mangrove forests. This collapse process lead to an elevation loss of 11 mm yr^{-1}. They concluded that mass mortality of wetland vegetation (regardless of cause) can lead to similar elevational reductions for any organic soils with loss on ignition values of greater than 40 %.

High energy events such as hurricanes and tsunamis also have the potential to cause significant erosion in intertidal wetland environments (Conner *et al.*, 1990; Walker *et al.*, 1987). Small and localised erosional events at a saltmarsh edge are unlikely to effect significant changes to stress conditions. However, larger erosional events associated with high wind and wave energy may remove significant proportions of overburden sediment, reducing the effective stress in underlying layers. The effect of this is to overconsolidate the materials. In the sediments studied in this thesis, such an erosional event would not invalidate the use of the model providing that the effective stress decrement caused by erosion is less than the value of the preconsolidation stress calculated for the surface materials (i.e. providing overburden sedimentation had not increased effective stress in the soil to a level greater than the preconsolidation stress observed at the surface). However, in different environments and at different times where overconsolidation was not so significant due to more rapid sedimentation rates, models which do not describe this overconsolidation would lose their predictive capacity if the accommodation space created by erosion was not filled sufficiently quickly.

Coastal areas prone to earthquakes (e.g. Hamilton and Shennan, 2005a; 2005b; Long and Shennan, 1994; 1998; Shennan *et al.*, 1999; Zong *et al.*, 2003) are likely to undergo significantly different processes of autocompaction, such as soil liquefaction, that are not described by the models presented in this study.

A period of extreme temperature during drought conditions may be great enough to fully desiccate near-surface sediments. Since the air entry values of the soils considered in this thesis range from *c.* 100 to 450 kPa (Figure 5.28), full desiccation is theoretically likely to render intertidal lithologies fully incompressible within the effective stress range experienced in Holocene intertidal stratigraphies. Consequently, application of the conceptual modelling approach developed in this thesis without sufficient testing of material properties would lead to poor accuracy of prediction.

It is important to remember that a model is a simplified representation of reality. Those developed in this study are based on a limited range of lithologies and perhaps highly idiosyncratic environmental conditions. Hence, as with any model, the conceptual approach for modelling the compression behaviour of intertidal mineralogenic lithologies is only strictly applicable to the sediments upon which it is based. Indeed, Oreskes (2003) argues in general that models can never fully specify the systems that they describe and so long-range deterministic predictions are likely to be erroneous as the range of independent variables (and uncertainties therein) increases. Hence, uncritical 'black box' application of the autocompaction models presented here should be avoided. Nonetheless, the conceptual modelling approach developed in this thesis and the material testing program which led to its development will provide a valuable reference point for guiding research methodologies that aim to develop an understanding of sediment autocompaction in different lithologies and depositional environments.

8.4.4 Implications for model application

The largest obstacle to pre- and retro-diction of volumetric and surface elevation changes in intertidal environments is the autocompaction behaviour of organogenic facies. Indeed, future research will need to focus on the physical, biological and chemical processes that control the volumetric state of these sediments and the timescales over which they operate. Until such detailed investigations have been completed, it seems that accurate fore- and hind-casting of surface elevation change in intertidal environments can only be undertaken on relatively thin, predominantly mineralogenic sequences.

However, even once organogenic sediment autocompaction is understood, problems of model application and extrapolation through time will remain difficult as an increasing number of processes will affect sediment volume. In addition, the rates at which these processes occur may be very different to those modelled on the basis of field or laboratory data (after Statham, 1977). In a predictive capacity associated with saltmarsh management in the face of rising sea levels, these problematic temporal aspects may not prove to be as great an issue, since the timeframe of interest (perhaps 10 – 100 years) is considerably shorter than that associated with the retrodictive, palaeoenvironmental context (100 to up to 8000 years, for example). With the benefit of a continuous monitoring program running concurrently with modelling predictions, any deviations from such predictions can be noted when forecasting elevation change and the model can be refined or the input parameters can be re-quantified (should significant diagenetic alteration have occurred, for example).

When attempting to decompact sediments, significantly more challenging problems occur in determining previous thicknesses of sedimentary deposits, particularly in organic facies. The simple effective stress-based approach is no longer sufficient and it becomes difficult to determine when (in terms of depth of burial/effective stress) various biochemical alterations to such organogenic fabrics occurred. Any geotechnical models developed on *in situ* (i.e. stratigraphic, at depth) behaviour would result in hindcasting of compression behaviour of the material in its current state and not that at the time of formation. Furthermore, with such dramatic alterations to the material properties of a strongly humified sediment, compression parameters of the unaltered, near-surface sediment, such as the initial voids ratio, yield stress and compression indices are extremely difficult to estimate. The challenge will therefore be in synthesising geotechnical and palaeoenvironmental approaches to develop generic models of change for a given lithology. Depositional elevations can be estimated using the microfossil sea level indicators contained with diagenetically modified sediments. Through compression testing of the equivalent, contemporary sediments, the early stage compression behaviour can then, within error, be reconstructed. Through further microtextural and geochemical analysis, further insights may be gained into the relative chronology of the operation of various autocompaction processes. Needless to say, opportunities for future research in this field are plentiful.

8.5 Overconsolidation in Low Energy Intertidal Environments: Geomorphological Implications

It was noted in Section 8.4.3 that extreme events can potentially alter the autocompaction behaviour of mineralogenic intertidal sediments. An interesting concluding consideration involves the subtle feedbacks that may occur between autocompaction and extreme events, and any consequent implications for the geomorphological functioning of saltmarsh and mudflat systems.

Conventionally, storm surges are often implicitly considered to be the most significant high magnitude, low frequency events in driving geomorphological change in intertidal environments (Jennings *et al.*, 1995; Pethick, 1996; Reed, 2002). Such high energy events mobilise and redistribute sediment; this mechanism permits the land- and sea-ward migration of the shoreline in response to varying energy conditions over a variety of timescales (Crooks, 2004; Pethick, 2001; Pethick and Crooks, 2000). However, if these sediments were to undergo heavy overconsolidation by severe desiccation, the resultant denser sediment structures and the increased degree of interparticle contact would lead to increased critical shear strengths. Consequently, the kinetic energy of turbulent storm floodwaters may no longer be of a sufficient magnitude to erode these sediments. Sediment fluxes during high energy storm events may thus be reduced and upper intertidal landforms may lose their capacity to migrate in response to varying energy conditions.

It is possible, therefore, that a combination of low spring tides, high levels of solar insolation and an absence of precipitation may constitute a significant extreme 'event' that is capable of altering saltmarsh morphodynamic functioning. Hence, direct and intense solar inputs into low energy intertidal systems may prove to be as important as high energy storm events in determining the future behaviour of estuarine systems *via* their role in determining the stress history (i.e. whether under-, normally- or over-consolidated) of intertidal sediments.

8.6 CHAPTER SUMMARY

This chapter has synthesised the findings of the detailed field and laboratory investigations into autocompaction processes and behaviour undertaken in this study to develop an empirically-informed model of autocompaction using a Bayesian statistical regression technique. Comparison with the sedimentation compression curves obtained from x-ray core scanning revealed that the models predict voids ratios well for the lithologies for which they were developed.

The practicalities of model application have been addressed. This focused on the best way in which sediment types can be grouped together for purposes of statistical modelling. Since mudflat sediments have very similar lithological properties and display very similar compression behaviour, a decision was made to group these sediments together. Difficulties arise when *in situ* organic growth begins to complicate matters by causing a continuum of structures and compression behaviours. Furthermore, biochemical processes, particularly humification, can change the compression behaviour of these partially organic sediments so that the *in situ* structure is no longer directly linked to depositional elevation in the tidal frame or the structural and behavioural characteristics observed at the time of formation. It was therefore decided that such sediments should be grouped for modelling according to their *in situ* compression behaviour.

Using autocompaction models based on the appropriately grouped sediments, a short stratigraphic sequence was decompacted. Reconstructions of mean sea level were provided by foraminiferal assemblages that were calibrated by a transfer function. These reconstructions were placed into a chronological framework developed using radionuclides. Together, these data allowed a twentieth century age-altitude plot to be produced for both the *in situ* (compacted) sea level index points and those that were decompacted using two different rheological models: the newly-developed models of this study and Terzaghi's compression law. By comparison with compaction-free tide gauge records of mean sea level, the accuracy of the predictive capacity of each of these models was determined. It was found that the new models perfectly reconstructed the twentieth century rate of sea level change, whereas the Terzaghi model overpredicted by 1.1 mm yr^{-1}. Validation of the newly-developed models in this manner justifies the investigation into autocompaction processes and mechanisms and further suggests that uncritical, 'black-box' application of models without sufficient adaptation is a potentially dangerous practice.

Finally, the transferability of the model was also discussed. The conceptual modelling framework developed throughout this thesis is likely to be valid to similar mineralogenic sediments deposited within similar northwest European low energy intertidal zones. However, as the spatial and temporal scales are expanded, varying lithological (particularly organogenic sediments), hydrographic, geomorphological, diagenetic and climatic factors are likely to alter autocompaction processes and behaviour in such a manner that the modelling approach developed in this study becomes increasingly inapplicable.

CHAPTER 9: CONCLUSIONS

9.1 ORIGINAL CONTRIBUTIONS TO KNOWLEDGE

The principal aim of this study has been to develop a quantitative model of the autocompaction behaviour of contemporary and recent mineralogenic intertidal sediments based on field and laboratory data. This aim has been addressed by adopting a multidisciplinary perspective *via* the combination of theory and techniques from geomorphology, soil mechanics and palaeoenvironmental studies. Such an approach has enabled a re-evaluation of the previously employed, entrenched and inflexible models that were originally developed by civil engineers concerned with minimising foundation settlement. This study has contributed to knowledge in an original way by combining the results of field and laboratory investigations. The dynamics of effective stress variation within the intertidal zone have been quantified and conventional laboratory tests have been modified to account for these variations. This synthesis of data has permitted the development of phenomenologically accurate models of autocompaction in predominantly mineralogenic intertidal sediments.

The results of this study can be summarised by revisiting the research objectives defined in Chapter 1.

- To review current understanding of the autocompaction of intertidal sediments in order to identify existing knowledge gaps

Following a review of the literature from a wide range of cognate disciplines, it was concluded that existing stress-based compression models have been applied to problems of intertidal autocompaction with no regard for the specific conditions of the dynamic intertidal environment and the materials that form there. The transferability of models such as Terzaghi's compression law and consolidation theory had been entirely assumed. A complete lack of empirical data regarding the basic one-dimensional compression behaviour of intertidal sediments was also recognised. Similarly, the effects of the relevant non-mechanical diagenetic processes on structure and compression behaviour were also key unknowns.

The identification of key research issues that confound the development of accurate models of autocompaction allowed six research hypotheses to be defined to guide the research within the study. These hypotheses challenged the existing assumptions of previously employed autocompaction models and they were subsequently addressed throughout the empirical and experimental stages of the study.

- To investigate and quantify the key causes of effective stress changes within intertidal environments.

A local tide gauge was installed at a low elevation within the intertidal frame at Greatham Creek to determine the local hydrographic characteristics of the site. In order to determine the magnitude and periodicity of variations in the groundwater table beneath the low marsh surface, a piezometer was installed. Using hydrostatic principles, and under the assumption that the tidal waters were hydraulically isolated from groundwater due to an impervious horizon, the effective stress variations through time were calculated for two altitudes within the intertidal frame. These altitudes were 1.06 m OD (mudflat sampling site) and 2.26 m OD (low marsh sampling site). It was found that effective stresses acting throughout intertidal stratigraphies are never constant due to the combined operation of tidal water loading and capillary suction stresses. Further causes of effective stress variation, likely to be greater in magnitude than those caused by hydrological variables, resulted from subaerial exposure and subsequent desiccation.

- To undertake a detailed geotechnical testing program that has been tailored to the specific effective stress conditions of the dynamic intertidal environment.

Following the field investigation, a detailed laboratory testing program was devised. This program adapted the methods employed in established geotechnical procedures, since these did not reflect the reality of conditions experienced by sediments in the intertidal zone. Firstly, to reflect the unsaturated nature of the *in situ* sediments, the saturation stage of conventional oedometer tests was not carried out to determine how this affected compression behaviour. Secondly, a series of tests using a new, custom-built apparatus (the back-pressured shearbox) were undertaken on the mineralogenic sediments. These tests involved applying dynamic loads to simulate tidally-driven effective stress variations on surface samples. Subsequently, a combination of incremental (oedometric) and dynamic loading scenarios were employed. The boundary conditions for both types of dynamic loading tests were based on the results of the field monitoring program. Conventional oedometer tests were also carried out.

- To identify the main settlement processes that cause volumetric variations in mineralogenic intertidal sediments.

Primary consolidation dominates settlement in homogenous saturated clays in static effective stress environments. As a result, models that describe the volumetric evolution in these sediments do not require a time-dependent creep component. It was critical to determine whether creep contributed significantly to settlement in the mineralogenic low marsh and mudflat sediments studied in this thesis.

Conventional oedometer tests revealed that the creep component can contribute significantly to settlement if the associated settlement trend is extrapolated over centennial and indeed millennial timescales. However, further insights into the real-world behaviour of these sediments were revealed by the dynamic loading tests; the conclusions reached on the basis of these data contradict those obtained from conventional, static tests. The dynamic surcharge loading of the sediments by tidally-driven processes overconsolidates the sediments and increases structural stability when exposed to a stress lower than the maximum previous stress applied; this significantly reduces the rate of creep. Furthermore, creep occurs at constant effective stress; the continuous variations in effective stress observed in the field monitoring program result in a corresponding elastic cycling of the unload-reload hysteresis loop and creep deformation is prevented from occurring. Hence, the most important settlement process in the mineralogenic intertidal sediments tested is primary consolidation. This conclusion provides a firm, empirical rationale for the exclusion of creep processes in models of autocompaction of these mineralogenic sediments.

- To consider the importance of diagenetic processes in affecting the compression behaviour of these sediments.

Analysis of short cores obtained from the low marsh and mudflat sampling altitudes revealed two post-depositional processes in the vadose zone: diagenetic remobilisation of overconsolidating, concretionary redox-sensitive materials and humification of organic matter. These diagenetic modifications to the soils in the vadose zone can decrease the accuracy of models of autocompaction if unaccounted for. The absence of a suitable quantitative technique for determining the degree of humification in predominantly mineralogenic saltmarsh materials meant that the effects of this process on autocompaction remain unquantified.

Multiple regression analysis to ascertain the relative importance of various lithological and geochemical variables in predicting *in situ* voids ratios suggested that diagenetically mobile geochemical substances are the most important factor. However, the importance assigned to them by the statistical procedures was deemed to be an artefact of their preferential adsorption onto different grain size and organic fractions.

Oedometer samples were also obtained from the mudflat cores to determine whether diagenetic enrichment of geochemical substances affected compression behaviour in response to loading. The results of these oedometer tests revealed that variations in structure and compression behaviour of the diagenetically-altered sediments obtained from depth were not greater than those observed in the virgin samples. The only exception to this was a slight increase in the preconsolidation stress. Diagenetic processes therefore affect neither *in situ* voids ratio nor compression behaviour enough to justify an increase in model complexity.

- To develop an empirically-informed predictive model of the autocompaction behaviour of mineralogenic intertidal sediments.

The results of the field and laboratory investigations have permitted the development of phenomenologically accurate models based on a set of empirical observations. The geotechnical testing program demonstrated that both the low marsh and mudflat specimens tested were overconsolidated and displayed considerable variation in initial structure, which translated directly into differences in compression behaviour. The dynamic loading tests revealed that a time variable is not necessary to account for creep processes due to their inactivity. Similarly, a temporal analysis of pore pressure dissipation and consolidation is not possible in the sediments analysed due to the simultaneous operation of consolidation and creep processes and the consequent graphical deviations from conventional Terzaghi plots of time-settlement curves; this prohibits the successful use of established curve-fitting procedures. In addition, low load increment ratios lead to similar deviations in graphical form. Hence, Terzaghi's consolidation theory cannot be applied to these sediments.

The degree of saturation did not significantly affect the compression behaviour of the sediments so its influence does not require inclusion as a predictor variable. Similarly, effective stress was judged to be the dominant control on voids ratio within a single lithology in the overconsolidated vadose zone; diagenetic influences are deemed to be unimportant.

Since only one predictor variable (effective stress) significantly affects the dependent variable (voids ratio – a volumetric parameter) for each lithology, the models could still be develop within the conventional $e\log_{10}\sigma'$ framework providing that each of the necessary modifications outlined above is made. A statistical model was developed to describe the full range of variations in structure and compression behaviour observed. Also, a 'conditional' regression model was employed to account for the overconsolidation in the sediments; values of slope constants vary at effective stresses above and below the yield stress (i.e. the stress at which the rate of compression increases from the shallow gradient recompression line to the normal compression line).

- To apply the model to a stratigraphic sequence and compare the accuracy of its predictions to those of previously employed models of autocompaction.

The newly-developed autocompaction models were applied to a short stratigraphic sequence from which reconstructions of twentieth century variations in sea level were made on the basis of foraminiferal assemblages and radionuclide dating techniques. The resulting sea level curves were decompacted using both the newly-developed model and Terzaghi's compression law (a previously employed rheological model). The rates of relative sea level rise obtained from these curves were compared with that obtained from the North Shields tide gauge – a compaction-free instrumental record. It was found that the geological rate obtained from the reconstructions decompacted with the models developed in this study matched that of the tide gauge record perfectly. In contrast, decompaction using the Terzaghi model resulted in an overprediction of the rate of sea level rise by 1.1 mm yr^{-1}. These practical applications suggest that the previously existing models are insufficient and lack the predictive accuracy of the empirically-informed alternatives developed in this thesis.

9.2 Recommendations for Future Research

The final objectives of this study were to consider the broader implications of the newly-developed autocompaction model, speculate on its transferability and to identify areas for future research. Indeed, the different components of this objective are closely linked since many of the broader implications remain speculative and can only be verified through additional empirical research. Despite the need for geotechnical testing to determine the exact values of the required model parameters, the conceptual modelling framework developed throughout this thesis is likely to be valid to similar mineralogenic sediments deposited within analogous northwest European low energy intertidal zones. However, it may be possible to improve on the accuracy of the models through further research into the effects of specific processes and conditions:

- The models are based on laboratory tests that were undertaken using distilled water. It would be interesting to determine whether variations in pore water chemistry (i.e. increased salinity) significantly alter compression behaviour.
- The laboratory data strongly suggested that creep is a negligible process in the volumetric reduction of mineralogenic intertidal materials. However, the compression models are based on oedometer tests which inevitably incorporate some degree of creep. Further tests are required within which compression only occurs by consolidation; this can be done in a triaxial cell within which lateral deformation is constrained by a radial strain gauge. Vertical effective stress can then be increased monotonically at different rates and excess pore pressures can be monitored; when these reach zero, further stress is applied to prevent creep settlement from occurring.
- This study undertook dynamic loading of materials based on the upper stress boundary conditions observed in the field. Also, dynamic loads were interspersed with periods of constant effective stress. It would therefore be of interest to determine how the materials behave under constantly changing effective stress of lower magnitudes (e.g. 1 – 5 kPa).
- Since subaerial exposure has been shown to increase the effective stress acting on a sediment sample, it would be of particular interest to determine the magnitude and duration of a desiccation event that would be required to render a sediment fully incompressible. Desiccation events of varying magnitude and duration could be investigated through the use of an environmental cabinet in which temperature, humidity, and the intensity of insolation can be simulated. This is of particular interest

given the predicted increase in the return intervals and extremity of drought events in a warmer climate.

- Further model validation is also needed; this could involve the use of Sedimentation-Erosion Tables (SETs) whereby predictions of surface elevation change from autocompaction models are compared with directly observed, quantitative measurements. This may assist in identifying further mechanisms and trends of volumetric change, particularly in terms of the temporal dimension.

However, perhaps more pressing issues exist that require more prompt attention. As the spatial and temporal scales are expanded, varying lithological, hydrographic, geomorphological, diagenetic and other factors are likely to alter autocompaction processes and behaviour in such a manner that the modelling approach developed in this study becomes inapplicable. Hence, research is needed into different sediments in different environments. Perhaps most obviously and importantly, first-order, inductive research is required into autocompaction mechanisms of organogenic materials that form higher in the intertidal frame and also supratidally. Such sediments are not only prone to significant volumetric changes as a result of effective stress variations; their structure and volume are also directly dependent on biological and chemical decay processes. Only with an understanding of the autocompaction behaviour of these sediments can volumetric change within mixed organogenic and mineralogenic stratigraphies be accurately fore- an hind-cast.

Despite the challenges that are presented by predominantly organogenic sediments, the issues identified in this study regarding the need for multidisciplinarity and the synthesis of field observations and laboratory tests will unquestionably prove to be significant in unravelling the complexities of the autocompaction processes operating in these sediments. Indeed, this study is part of ongoing research into autocompaction processes within intertidal areas.

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