SOME ASPECTS OF THE GEOLOGY AND ENGINEERING PROPERTIES OF THE HOLOCENE DEPOSITS AT THE BOTHKENNAR SOFT CLAY RESEARCH SITE

by

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This thesis is submitted to Heriot-Watt University in accordance with the requirements for the degree of DOCTOR OF PHILOSOPHY

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I DEDICATE THIS THESIS
TO MY FAMILY
STATEMENT OF ORIGINALITY

The research upon which this Thesis is based was undertaken during my employment as a Research Associate at Heriot-Watt University between October 1988 and August 1994. The work was funded by the Science and Engineering Research Council (grants GR/E74748, GR/H/14151, GR/J/56714) and formed part of the programme associated with the establishment of the Bothkennar Soft Clay Research Site.

The field and laboratory work described herein were undertaken by me personally unless noted otherwise in the text, in which case due acknowledgement is given. In these cases the interpretation of the results is my own.

I have previously published some aspects of my research in the peer-reviewed literature with my supervisor, Professor M.A. Paul. This material was produced as a collaborative effort and cannot be assigned uniquely to either individual. The publications are referenced in the normal way where appropriate and copies are included in a pocket at the inside back cover of this Thesis. In certain sections of Chapters Five, Six, Seven and Eight, use has been made of text or figures from these publications, with or without subsequent modification. In these instances the source is acknowledged at the start of the relevant section and I also acknowledge here the contribution of my co-author to the written form of the publication. The use of this material is made with his knowledge and agreement. However, the interpretation of the results in the context of this Thesis remains my own.

The overall text of the Thesis has benefitted from the advice of Professor Paul on matters of structure, balance and scientific style and I am pleased to acknowledge his guidance in these respects.

Beverley Frances Barras
9 May 2000
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ABSTRACT

This thesis describes the sedimentology and engineering geology of the Holocene estuarine sediments of the Claret Formation at the EPSRC Bothkennar Soft Clay Research Site near Grangemouth, Scotland. The composition, fabric and geotechnical characterisation of the deposits have been described using detailed borehole logging, X-ray densimetry, X-radiograph XRD, SEM and soil mechanical techniques, and these factors have been related in detail to changes in relative sea-level in the Bothkennar area during the period 5,000 – 3,000 ¹⁴C yrs BP.

The sediments can be divided into bedded, mottled and laminated facies on the basis of primar sedimentary structures and the nature of bioturbation. Within these facies it is possible to identify three electron microfabrics termed aggregate, cumulate and granular. The succession c facies has been controlled by the fall of relative sea-level since the Holocene maximum and records an emergent sequence from subtidal (probable water depth around 20m) to intertidal. It is possible to relate many minor discontinuities in the property profiles to probable fluctuations in the local depositional regime.

The facies are very similar in their overall particle size distribution, mineralogy and basic geotechnical properties. The sediment is of high plasticity (Iₚ ≤53%), has a complex water content profile at the centimetre scale and is lightly overconsolidated (YSR ~ 1.6). These characteristics can be explained by reference to the mineralogy, pore water composition and (marine) organic components, combined with the changing sedimentological conditions that arose from deposition in a shallowing water body.

A comparison of similar Holocene deposits (from Belfast, the Severn Estuary and from Drammen) with those at Bothkennar provides an initial framework within which to interpret the results of similar studies, taking as its focus the effect of relative sea-level change on the generation of complex physical property profiles via changes in the local sedimentary regime.
CHAPTER ONE

INTRODUCTION

1.1 PREAMBLE

1.2 HISTORY OF THE FACILITY AT BOTHKENNAR

1.3 ORGANISATION OF THE THESIS
CHAPTER ONE

INTRODUCTION

1.1 PREAMBLE

This thesis describes the engineering geological research undertaken by the writer between 1988 and 1994, during the establishment of the EPSRC (Engineering and Physical Sciences Research Council)\(^1\) soft clay research site at Bothkennar, near Grangemouth, Scotland (Figure 1.1). The opportunity for this research arose while the writer was employed at Heriot-Watt University as an SERC Research Associate specialising in Quaternary engineering geology.

The overall aim of the work was to understand the engineering geological setting of the Bothkennar site and to link this with the wider picture of Holocene sedimentation in the Forth estuary. The central concept has been that the present day physical properties of the sediments can be best understood by reference to their geological setting and history, via a consideration of their composition and sedimentary fabric at a variety of scales. Thus the research to be described here has centred on the detailed examination of the sediments from the site itself and the analysis of published geological and geotechnical data from the Forth area. These have been integrated into a palaeoenvironmental model of the site during the Holocene period, which in turn provides an explanation for the sedimentary fabric, facies architecture and geotechnical profiles at the site.

Soft estuarine clays of Holocene age are common around the indented coastline of Scotland due to the abundant availability of earlier sediments, usually of glacial origin, and the changes in relative sea-level that have occurred since the close of the Devensian glaciation. These latter events have included a major transgression in the mid-Holocene, during which marine embayments extended inland of the present coastline, and a later regression, during which the embayments were infilled with sediment and raised isostatically above the present sea-level. This has given rise to the extensive, flat lying ‘carselands’ which are found at the heads of nearly all the Scottish firths on both the east and west coasts and which are normally underlain by fine grained estuarine sediments, which may attain considerable thicknesses around their local depocentres.

\(^{1}\) Until the re-organisation of the Research Councils in 1996, the Bothkennar site was operated by the Science and Engineering Research Council (SERC) who were the funding body for the work described here.
Figure 1.1 Location of the Bothkennar Soft Clay Research Site (BSCRS) at NS920858 adjacent to the Forth Estuary, Scotland. Extracts from Ordnance Survey 1:50,000 map sheet 65 (Ordnance Survey, 1997) and 1:24,000 aerial photograph (Royal Commission on the Ancient and Historical Monuments of Scotland, 1988).

Crown copyright
The research presented in this thesis is concerned with the particular carse clays which lie towards the head of the Forth Estuary. This area is very suitable for a study of this kind, since there is already a large body of well-tested data on the Late Quaternary chronology of sea-level change which provides a framework for more detailed investigations such as those described here. This framework has been exploited to provide a very detailed palaeoenvironmental history of the study area which can be related directly to the sedimentary record and to the corresponding physical properties.

In order to meet the aims of this research, the following general lines of enquiry have been pursued.

- The overall setting and Late Quaternary geological history of the Forth estuary have been established from the literature, as has the history of previous geotechnical research in this area;

- The macrofabric, sedimentary facies and facies architecture of the sediments at the Bothkennar research site have been established from a programme of very detailed borehole logging, combined with X-radiography and high resolution densimetry;

- The microfabric of the sediments has been established by electron microscopy using freeze-dried, fractured surfaces examined in secondary electron mode. The selection and interpretation of the images has been made on a quantitative basis using statistical techniques designed to characterise the sedimentary facies;

- The composition of the sediments has been investigated by X-ray diffraction (XRD), laser granulometry and a variety of inorganic and organic geochemical techniques, and their geotechnical properties established by a laboratory programme of water content, plasticity, compression and strength testing;

- Examination of the published sea-level data, in conjunction with new radiocarbon assays from Bothkennar, have suggested a changing depositional environment that can be related, via changes in the sedimentary fabric, to the facies architecture and to profiles of geotechnical properties;

- The character and properties of the sediments at Bothkennar have been compared with those of other Holocene estuarine clays and the similarities and differences highlighted.

The organisation of this thesis in terms of these lines of enquiry will be described in section 1.3 of this introduction.
1.2 HISTORY OF THE FACILITY AT BOTHKENNAR

1.2.1 The Establishment of the Site

The Bothkennar Soft Clay Research Site (BSCRS; henceforth simply also called "the Bothkennar research site" or just "Bothkennar" when no ambiguity arises) was established by the former Science and Engineering Research Council (SERC) as a location at which large and full-scale ground engineering experiments could be undertaken. This was part of a larger strategy to establish a number of so-called 'test bed sites' on sediments of varying engineering character around the UK as a whole. An essential part of this strategy was to investigate very fully the geological setting and history of each site and to integrate this with an equally detailed investigation of the geotechnical properties of the sediments. In many ways the exercise undertaken at Bothkennar has proven an exemplar for activities of this type, as recently acknowledged by Professor D.W. Hight in his 1998 Rankine Lecture.

In 1984 the SERC commissioned the investigation of a number of possible sites around Britain with the purpose of purchasing a site in order to establish a large-scale in situ geotechnical testing facility. The site was to be situated on soft estuarine clays in order to complement the existing Building Research Establishment (BRE) stiff clay and glacial till sites at Cowden (Holderness) and at Garston (Watford) and the University of Bristol was awarded the contract to locate the most suitable site for this purpose. The chosen site had to conform to a number of geotechnical criteria (Nash & Lloyd, 1988a; Hawkins et al., 1989):

- It should be underlain by at least 10m of normally consolidated soft clay with an undrained shear strength <40kPa, plasticity index >20% and high compressibility;
- It should have a firm crust to allow access;
- Peat should be absent and organic material and/or sand intercalations should be minimal;
- The clay fabric should not be too laminated;
- The overall geotechnical variation across the site should be limited.

There were also additional operational requirements:

- The site had to be at least 5 hectares in size;
- It should be a greenfield site;
- The site should not flood regularly;
- There should be a low risk of vandalism on the site;
- No undermining should occur during the life of the site (anticipated to be 20 years).
Depositional environment, sensitivity, clay fraction, activity, porewater salinity and depth to water table were not specified (Nash & Lloyd, 1988a).

After an extensive desk study Hawkins et al. (1989) identified four possible sites: Brean on the Somerset levels (Severn Estuary), Newport on the Caldicot levels (Severn Estuary), Swale on the Thames Estuary and Bothkennar on the Forth Estuary. The Bothkennar site was considered to be the most suitable since it appeared to fulfil all of the criteria detailed above and so in 1987 the site was purchased by the SERC. In the six years following the purchase of the site (1988 to 1994), Heriot-Watt University was awarded three research grants in order to investigate the engineering geology at Bothkennar, in parallel with the geotechnical characterisation study which was undertaken at the same time (Hight et al., 1992a).

1.2.2 Introductory Description of the Bothkennar Research Site

The following section outlines the principal geographical and geological features of the site. These will be described in full in later chapters but are briefly presented here in order to orientate the reader.

The Bothkennar research site (Grid reference: NS920858, SW corner of site) is a greenfield site with an area of 11 hectares. It is situated immediately adjacent to the Forth Estuary on the south (locally the west) bank about one kilometre south of the Kincardine Bridge (Figure 1.1). The site is part of a previously intertidal area which was reclaimed for agricultural use in the late 18th century and is bounded to the north, east and south by flood embankments (Figure 1.2). A narrow strip of salt marsh now lies between the site itself and the estuary.

Within this reclaimed area the sediments of the Grangemouth Formation form a thin (about 1m) veneer which overlies between 15m and 20m of soft estuarine Holocene silty clays and clayey silts of the Claret Formation (Table 1.1). These comprise the local ‘carse clay’ at the site. The sequence overlies a prominent local gravel layer (the Bothkennar Gravel Formation) which in turn overlies about 5m of Late Devensian glaciomarine silty clays and sandy silts, which themselves rest in turn on glacial till over bedrock. The uppermost one to two metres of the deposit consists of a desiccated, weathered surface layer known informally as the ‘surface crust’.

The sediments of the Claret Formation form the subject of this thesis. As will be shown in later chapters, the Claret Formation is capable of subdivision into a number of distinctive and repetitive facies, the most significant feature of which is the preservation of primary sedimentary structure versus its destruction by bioturbation. At the Bothkennar site, a part of this repetitive sequence is cut out by sediments ascribed to a separate laminated facies of the Grangemouth

---

2 The term ‘carse clay’, although useful, is an informal term for an essentially diachronous sequence and does not have stratigraphical significance.
Table 1.1  Generalised stratigraphy at the Bothkennar Soft Clay Research Site.

<table>
<thead>
<tr>
<th>Geological Age</th>
<th>Stratigraphical Unit ¹</th>
<th>General description and probable origin</th>
<th>Approximate thickness at Bothkennar (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Modern</td>
<td>Grangemouth Formation</td>
<td>Weathered, massive to crudely stratified lagoonal clayey silt</td>
<td>0.5 - 1</td>
</tr>
<tr>
<td></td>
<td>(Saltgreens Member)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Holocene</td>
<td>Grangemouth Formation</td>
<td>Weathered, massive to crudely stratified intertidal clayey silt with numerous shells or shell fragments</td>
<td>0.5 - 1</td>
</tr>
<tr>
<td></td>
<td>(Skinflats Member)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Claret Formation</td>
<td>Soft to firm dark-grey to black micaceous silty estuarine clay</td>
<td>15 - 20</td>
</tr>
<tr>
<td>Late Devensian</td>
<td>Bothkennar Gravel</td>
<td>Dense sandy gravel with cobbles</td>
<td>2 - 3</td>
</tr>
<tr>
<td>(Late-glacial)²</td>
<td>Formation</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Loanhead Beds</td>
<td>Soft to firm brown to brownish grey glaciomarine silty clay ³</td>
<td>2 - 5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lockart Members</td>
<td>Stiff, red-brown or grey matrix-supported diamic: lodgement or meltout till derived from south or south-west ⁴, ⁵</td>
<td>15 - 20</td>
</tr>
<tr>
<td>Late Devensian</td>
<td>Park Burn Member</td>
<td>Compact dark grey fissile matrix-supported diamic: lodgement or meltout till derived from the west ⁴, ⁶, ⁷</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Carboniferous</td>
<td>Lower Coal Measures</td>
<td>Slightly to faintly weathered, white medium-grained sandstone</td>
<td>N/A</td>
</tr>
<tr>
<td>(Westphalian)</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

¹ D.G. Sutherland (1999) has suggested a revision to the nomenclature which reduces these formations to the status of member. This remains the subject of active discussion and so, in this Thesis, the earlier usage of Claret Formation etc. is retained.

² This refers to the time between the Late Devensian glacial maximum and the beginning of the Holocene around 10,000 BP.


⁴ Kirby (1968).

⁵ Kirby (1969a).

⁶ Kirby (1989b).
Formation, which appears to be associated with local channelling within the overall estuarine sequence.

The Claret Formation is terminated upwards by a major unconformity which probably represents a former intertidal surface. At the site this surface is partly marked by a in situ shell bed, principally containing the intertidal species Cerastoderma edule. Above this unconformity the sediments of the Grangemouth Formation are composed of a thin veneer of intertidal shell debris, intermixed with clayey silts noticeably coarser than the sediments of the underlying Claret Formation, and above these lie lagoonal deposits of the late eighteenth century reclamation, which extend to the modern ground surface. An irregular weathering front extends down to about the level of the shell bed, above which the sediments are desiccated and oxidised to form the surface crust. Thus the crust involves several stratigraphical units and should not be considered a stratigraphical term in itself.

1.3 ORGANISATION OF THE THESIS

This thesis comprises nine chapters, implicitly divided into three substantive parts: the background to the Bothkennar research site; the results from this study; the discussion of the results from Bothkennar and their context.

Chapter Two describes the detailed geological background at Bothkennar and its surrounding area in order to establish a geological framework at the site. The description is divided into three parts: the bedrock, the overlying glacial deposits and the Holocene estuarine infill. The bedrock is detailed with a view to suggesting a possible provenance for the fine-grained soft sediments which make up the estuarine deposits in much of the Midland valley of Scotland and at Bothkennar in particular. The glacial history which resulted in the erosion of bedrock and the deposition of glacial tills, moraine materials and outwash deposits in central Scotland is discussed, followed by a consideration of the late-Quaternary sea-level history in the Forth area. The chapter is concluded by an account of the programme of land reclamation which took place in the Forth valley.

Chapter Three reviews the geotechnical investigations conducted at the Bothkennar site by other workers and the earlier geotechnical investigations in the wider Forth area. Particular attention is paid to the concept of a geotechnically structured sediment, in which profiles of parameters such as water content, liquidity and void index and undrained shear strength may be expected to be controlled by some aspect of the sedimentary fabric.

Chapter Four discusses the design of the research strategy on which this Thesis is based and details the field work and sampling programme, the methods used for the subsequent conservation of the samples and in the subsequent laboratory work.
Chapter Five presents the sedimentary fabric data from this investigation. It includes a detailed macroscopic description of the sediments from the BSCR, including visual, X-radiograph and densimetry profiles. The results of electron microfabric investigations are also presented.

Chapter Six presents the compositional data from this investigation. These include the mineralogy of the sedimentary particles, their size distribution, the nature of the organic material within the sediment and some aspects of their inorganic geochemistry and pore fluid composition.

Chapter Seven presents the geotechnical data from this investigation. This has two elements. The first is a characterisation in the remoulded state which emphasises the importance of salinity and organic material as controls on the plasticity. The second is the description and analysis of property profiles (water content, density, undrained shear strength, liquidity index and void index) which are linked to a detailed consideration of the sedimentological log.

Chapter Eight discusses published sea-level data from the Forth area and, from radiocarbon dating of the aggrading sediment surface, deduces the contemporaneous water depth throughout the depositional period. From this history the facies architecture at Bothkennar can be interpreted in terms of the palaeoenvironmental history and related to the overall development of the geotechnical profiles. This chapter also reviews three analogous sites where studies have been carried out on soft Holocene clays. These are Belfast Lough in Northern Ireland, the Severn Estuary in south-west England and Drammen on the Oslofjord in southern Norway. The engineering geological setting and depositional history of these sites are detailed and their geotechnical characteristics compared and contrasted with Bothkennar in order to place the latter in a wider context.

Chapter Nine summarises the principal conclusions of this work and makes a number of recommendations for further study, both at Bothkennar itself and in its comparison with other soft clay sites.

Certain of the results presented in Chapters Five, Six and Seven have previously been published in the peer-reviewed literature in a series of papers co-authored by the writer. These publications are referenced as required and for completeness are also itemised here. Copies are also provided in the rear pocket of this thesis for ease of reference.

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Bibliographic details of these publications are given in the list of references. Wood is the maiden name of the writer.
CHAPTER TWO

GEOLOGICAL BACKGROUND TO
THE BOTHKENNAR SOFT CLAY RESEARCH SITE

2.1 INTRODUCTION

2.2 OUTLINE BEDROCK GEOLOGY IN THE BOTHKENNAR HINTERLAND

2.3 GLACIATION OF THE FORTH AND ITS HINTERLAND

2.4 HOLOCENE DEVELOPMENT OF THE FORTH ESTUARY

2.5 HISTORICAL LAND RECLAMATION IN THE FORTH VALLEY

2.6 SUMMARY
CHAPTER TWO

GEOLOGICAL BACKGROUND TO
THE BOTHKENNAR SOFT CLAY RESEARCH SITE

2.1 INTRODUCTION

This chapter describes the overall geological setting of the Bothkennar Soft Clay Research Site (BSCR). There are four principal aspects to be considered: the nature of the solid rocks underlying the hinterland from which the Bothkennar sediments were ultimately derived; the glacial history by which they were brought into the Forth drainage basin; the Bothkennar sequence itself together with the Holocene environmental changes that affected the Forth estuary whilst it was being deposited; and the artificial reclamation of the area in historical times, which has brought the BSCR to the condition in which it is seen today.

At the height of the Devensian Glaciation, the Forth Valley was occupied by a major eastward flowing ice stream, whose source area is believed to have lain in Rannoch Moor. In consequence the ice discharged across a variety of igneous, sedimentary and metamorphic rocks which provided the substrate for the glaciogenic sediments, principally tills, that now occupy most of the Forth lowlands. Subsequent re-erosion of these materials has provided the immediate source material for many of the Late-glacial and Holocene sediments of the Forth Valley, including those at the BSCR.

There is considerable evidence of relative sea-level change in the Forth basin. There is a wealth of previous research on this subject, which has led to one of the best-reconstructed sea-level histories of any area in Europe: this is a major bonus for this present research. Historical land reclamation in the Forth Valley has also been an important factor in the establishment of the present coastline of the River Forth. Reclamation on the banks of the Forth River and estuary (including the vicinity of Bothkennar) commenced during the 18th century, when large areas of mudflats became prime agricultural and industrial land. The overall outcome is a palaeoenvironmental model for the area around Bothkennar which forms the basis for the Holocene stratigraphy at the BSCR. This in turn provides a framework for the geotechnical work presented later in the thesis.

2.2 OUTLINE BEDROCK GEOLOGY IN THE BOTHKENNAR HINTERLAND

The term 'Bothkennar hinterland' is applied to the area from which glacial action is likely to have eroded the materials which now comprise the clay sediments of the Claret Formation. The basis on which this hinterland is defined is presented in the next section; it can be said here that there
is a probable pattern of former ice movement from the western Highlands to the Stirling area via Loch Lomond or Stratheyre.

Examination of the geological map of the central area of Scotland (Figure 2.1) gives an overview of the bedrock geology of the area surrounding the BSCRS and this hinterland. It can be seen that the rocks encountered range from Lower Palaeozoic (Dalradian) metamorphics (mainly low grade metasediments) to Upper Palaeozoic (Carboniferous) sedimentary formations, together with associated intrusive and extrusive igneous lithologies.

The Midland Valley of Scotland is bounded to the north by the Highland Boundary Fault (HBF) and to the south by the Southern Upland Fault (SUF), both north-east/south-west trending fractures (Figure 2.1).

2.2.1 Metamorphic Basement Rocks

To the north of the HBF lies a band (about 30km in width) of pelitic rocks (Figure 2.1: index numbers 18, 19 and 21). They are characterised by low grade, regional metamorphism and have been correlated with the Southern Highland Group (Upper Dalradian) of the Dalradian Supergroup (Johnson, 1991). The area to the west and east of Loch Lomond is characterised by the Ben Ledi Grits and Aberfoyle Slates (British Geological Survey, 1987) and consist of quartz-mica-schists, grits, slates and phyllites. Robertson & Henderson (1984) and Stephenson & Gould (1995) have reported a mineral content of quartz, plagioclase feldspar, alkali feldspar, muscovite, biotite, Fe-chlorite, kaolinite, hematite, calcite, dolomite and apatite for the phyllites, slates and shales of this area. This large volume of metamorphosed pelitic rock was probably a major contributor to subsequent glacially eroded and transported deposits.

2.2.2 Sedimentary Rocks: Old Red Sandstone Strata

Southwards from the HBF, middle and upper Palaeozoic sediments occupy the remaining ground of the hinterland. Rocks of Devonian age crop out along the northern flank of the Midland Valley basin (Mykura, 1991; cf. Figure 2.1). A wide belt (20 to 30km, thinning to about 10km near Loch Lomond) occurs between the Ochil fault (OF) and the HBF and consists of Lower and Upper Old Red Sandstone (ORS) sediments, which have been folded along NE-SW trending fold axes. The dominant suite of minerals contains quartz, alkali-feldspar, muscovite, dolomite, microcline and plagioclase feldspar, with minor biotite, haematite, hornblende, zircon, epidote, apatite and chlorite (Francis et al., 1970) and their contribution to the Bothkennar sediments could have been considerable.

The Lower ORS in this area is divided into laterally continuous units (Armstrong & Paterson, 1970) of generally similar character along their strike (Wilson, 1980). They are a succession mainly of dark brown to purple coarse to pebbly sandstones and conglomerates, together with
Figure 2.1 Geological map of the central area of Scotland with lines of movement of the last Scottish ice sheet. Note position of the Bothkennar site (blue arrow). Basemap reproduced from the 1:625,000 United Kingdom (North) sheet, solid edition (British Geological Survey, 1979).
fine-grained mudstones, silty shales and grits. The sandstone lithologies are principally quartzitic with some arkose. Near the HBF the succession is dominated by conglomerates with contained pebbles of volcanic (andesitic) composition. To the east of the Stirling area a very considerable thickness of andesitic lavas and associated volcanic rocks is found. These thin westwards, becoming insignificant beyond about Callendar, although their pebbles are found in the conglomerates throughout the Lower ORS (Bluck, 1978; 1984).

In the north-west of the Midland Valley, the Lower ORS is succeeded directly by the Upper ORS which is of more limited extent than the lower one (Mykura, 1991). These consist generally of fine-grained, red-pink to brown sandstones, with numerous intercalated mudstones. Conglomerates are infrequent but when present contain pebbles of Lower ORS (mainly andesites and related volcanics) or Dalradian (mainly quartzite and schist) derivation. The overlying beds are rythmically bedded arenites consisting of pebbly sandstones and shales in well-defined cycles, which pass upwards into the overlying Carboniferous sediments.

2.2.3 Sedimentary Rocks: Carboniferous Strata

The Carboniferous rocks of the Forth hinterland occupy the northern flank of a major basin (the Kincardine Basin) whose axis trends approximately NNE-SSW (Figure 2.1: index numbers 78, 80, 81 and 82-3).

These rocks are of somewhat mixed lithology the dominant minerals being quartz, muscovite, biotite, plagioclase, orthoclase and microcline feldspar (fresh and weathered), dolomite and other carbonate minerals, plus heavy minerals: zircon, garnet, tourmaline, rutile (Francis et al., 1970). The lowermost units consist of grey or yellow-brown sandstones, shales and mudstones with some conglomerates, particularly in its upper part. The succeeding groups are largely similar to those at their base but contain subordinate limestones which are often used as marker horizons and to define the boundaries of various local subunits (Cameron & Stevenson, 1985). Coal-bearing cyclothem also become dominant higher up in this part of the sequence.

In the overlying rocks, the cyclothem of the lower units become obscured and grey-yellow feldspathic sandstones, often thick and sometimes pebbly, dominate. The pebbles are largely quartzite and appear to have been derived ultimately from a Dalradian source. There are also interbedded claystones with occasional reddening and the sequence is completed by pale grey to yellow sandstones, grey to dark grey mudstones, seatearths and coal beds.

2.2.4 Igneous Rocks within the Midland Valley

As noted above, igneous rocks make a substantial contribution to both the Devonian and Carboniferous sequences. They are of both extrusive and intrusive origin and include a wide variety of basaltic and calc-alkali lithologies.
Lavas of Devonian age crop out in a belt (up to about 20 km in width) immediately north of the Ochil Fault (Figure 2.1), although the outcrop narrows considerably westwards. They consist of andesitic and basaltic lavas and tuffs, with some rhyolite (Figure 2.1; index numbers 50 and 52).

Carboniferous lavas (largely basalts and spilites) are to be found to the west of the sedimentary basin (Figure 2.1; index number 53). They occur largely as flat-lying ‘plateau lavas’ and can be classified as being olivine basalts containing phenocrysts of labradorite and olivine (MacGregor, 1928).

Finally, quartz-dolerite dykes of late-Carboniferous to Permian Age (Figure 2.1; index number 35) are also known in the Midland Valley of Scotland. Here they are generally continuous, are about 30 m wide and generally trend east-west (Cameron & Stevenson, 1985; Smythe et al., 1995). The dominant minerals are plagioclase, augite and Fe-Ti oxides, while Ca-poor pyroxene (hypersthene and pigeonite), olivine, hornblende, biotite, chlorophaeite (chlorite), pyrite and apatite are common, but generally minor, constituents (Macdonald et al., 1981).

### 2.2.5 Implications for Quaternary Sediments of the Forth Area

The availability of the bedrock source materials described above has determined the character of the glacial sediments in the Forth area and so, indirectly, of the estuarine sediments derived from them. This theme is pursued later in this chapter and also in Chapter Six. However, it is useful at this stage to point out two features of the bedrock source materials that are directly reflected in the Forth deposits.

The most obvious characteristic is their colour. From the Dalradian through to the Carboniferous the sedimentary bedrock falls into one of two broad colour assemblages: grey or light coloured *versus* red-brown-purple. The grey or light coloured rocks include the Carboniferous sandstones (and related deposits) plus many of the Dalradian metasediments. The red coloured group include principally the rocks of the Old Red Sandstone series. This colour distinction is maintained in the glacial tills (section 2.3.3), which are divisible into a brown and a grey facies whose included clasts are largely rock fragments from the members of the relevant bedrock colour assemblage. The estuarine sediments also follow the same pattern: for example the sandy sediments of the buried raised beaches (Sissons, 1969; also this chapter, section 2.4) are reported to be either pinky-red or yellow-grey. The unweathered Claret Formation at Bothkennar initially appears black or very dark grey owing to the presence of iron monosulphide: on exposure to air it lightens to a pale grey. In general the colours in the estuarine sediments appear to follow a spatial pattern of reds around the basin margin and greys towards the depocentre: this same colour distribution is seen in the bedrock and (with less certainty) in the till facies.
The second characteristic is the mineralogical comparison between the bedrock and the Late Quaternary sediments. The minerals contributed by the bedrock are clearly extremely varied: indeed, nearly every common rock-forming mineral is represented somewhere in the area. Of particular note, however, and with reference to the assemblage seen at Bothkennar (Chapter Six, section 6.2), are the following:

- The low-grade phyllites and slates such as the Aberfoyle slates and Dunoon phyllite (both chlorite grade) are a possible source of chlorite;

- The mica schists prevalent throughout the region (such as the Ben Ledi Grit and the Beinn Bheula schist) are a possible source of both muscovite and biotite, although these minerals are also widespread in detrital form in the various sandstone units;

- The widespread Devonian and Carboniferous lavas are a possible source of both high grade ferromagnesian minerals and of both plagioclase and alkali feldspars. Feldspars derived from the lower Devonian lavas also occur in the arkosic sandstones of the Old Red Sandstone;

- The ‘Green Beds’ [metamorphosed volcanics] which occur in the Dalradian within the Beinn Bheula schist are a possible source of sodic plagioclase feldspars.

2.3 GLACIATION OF THE FORTH AND ITS HINTERLAND

In this section the glacial history and deposits of the Central Valley of Scotland are discussed, particularly those of the last (Devensian) glaciation. The Late Devensian amelioration and the short-lived readvance of the Loch Lomond Stadial, which resulted from a period of climatic deterioration immediately prior to the final deglaciation, are also included in this subsection. It is important to consider this history for two reasons. Firstly, glacial action was ultimately the mechanism that brought into the Forth area those sediments which, after reworking, now form the Holocene infill to the Forth estuary. Secondly, the removal of a large body of ice from Scotland has unloaded the crust and caused widespread isostatic uplift, whose timing and spatial pattern is a major control on relative sea-level change. These changes in turn are responsible for the architecture of the Holocene infill and are discussed in detail later in this chapter.

2.3.1 Patterns of Ice Movement

There is evidence for two major cold phases during the Devensian glacial period. Foraminifer populations from sediments on the Rockall Plateau (Ruddiman & Raymo, 1988) show these to be at about 70,000 and 20,000 BP (years before present), with an intervening warmer interstadial (Boulton et al., 1991). The presence of an extensive early Devensian ice sheet is suggested by data
presented by Sutherland (1981), Sissons (1983), Hall (1984) and Davies et al. (1984), although the extent of this glaciation is not known.

The Quaternary event which has left the most dominant terrestrial sedimentary record on the Scottish landscape is the last major ice sheet which was at a maximum about 18,000 to 20,000 BP and covered most of Britain (Figure 2.2). The ice flowed radially outwards from the Highlands and Southern Uplands of Scotland (cf. the ice flow lines on Figures 2.1 and 2.2) and in so doing removed most earlier glacial sediments from the low-lying areas. Thus, across Scotland as a whole, the depositional record of earlier glaciations has largely been lost and evidence of pre-Late Devensian glaciation is recorded principally by erosional features, although isolated terrestrial sediments do occur and more complete sequences are found offshore. The erosional features consist of U-shaped valleys and sea lochs, and corries of the high mountains of Scotland, whose depths are too great to have been the result of the most recent glaciation alone (Figure 2.3) (Boulton et al. 1991).

The study of erratic streams has also been of use in reconstructing patterns of ice movement in Scotland. Figure 2.4 shows the dispersal directions from distinctive bedrock outcrops in west-central Scotland. Of particular relevance to the Midland Valley of Scotland is an investigation of an outcrop of essexite near Lennoxtown in the Campsie Fells. In his study of this area, Shakesby (1978) examined field boundary walls that had been built from stones picked in adjacent fields, including erratics of porphyritic and non-porphyritic essexite which had a restricted outcrop and had never been quarried. The frequency of essexite erratics per unit area of wall was mapped and used as a measure of the erratic dispersion pattern, which revealed that the erratics fanned outwards in a generally eastward direction from their centre of dispersal.

Boulton et al. (1977; 1985) have reconstructed the British Devensian ice-sheet from field data in order to model the ice-sheet profile and dynamic and thermal regimes (Figure 2.2). Data for the earlier model used flow lines inferred from the distribution of erratics, till fabric studies, orientation of drumlins and glacial striae; climatic snowlines, the prediction of precipitation and distribution of accumulation. The later model incorporated data from ground surveys to reconstruct a pattern of longitudinal and transverse land features which reflect lines of ice flow and glacial retreat.

This reconstruction (Figure 2.2) shows that the ice-shed lay towards the western side of Scotland and that most of the ice was lost towards the Atlantic margin: presumably by calving from ice-cliffs along, or landwards of, the line of the continental shelf-break (Stoker et al., 1993) or, at later stages, via tidewater glaciers in the west coast fjords. Of particular importance, however, for the study of the Forth area is the eastward drainage of the ice sheet and here the
Figure 2.2  Reconstruction of the Late Devensian ice-sheet of the British Isles (source: Boulton et al., 1991: figure 15.16b).
Figure 2.3 The distribution of major glacially-eroded throughs in Scotland (source: Boulton et al., 1991: figure 15.5).
Figure 2.4 The main erratic trains found in Scotland which indicate the later flow directions of the Late-Devenian ice (source: Sutherland 1984: figure 6).

Key to numbered features:
1. Granite/ foliated granite;
2. Particular sedimentary outcrops;
3. Minor igneous intrusions;
4. Large igneous bodies;
5. Erratic transport paths;
6. Margins of erratic distributions;
7. Directions of ice flow.
reconstruction suggests that discharge occurred via major ice streams that occupied the Firth of Forth and the Moray Firth and, to a lesser degree, the Strathmore depression.

Ice appears to have exited the Highlands via the Loch Lomond basin and to have been deflected eastwards by ice from the southern uplands, the combined stream then discharging via the Forth. Direct visual evidence for this pattern is provided by the strongly aligned streamlined forms visible throughout the Lothians (Burke, 1969) and by the crag and tail features associated with the upstanding volcanic necks common in this area. At the height of the glaciation the terminus of this stream lay well to the east, possibly marked by the morainic bedforms of the North Sea Wee Bankie Formation (Gatliffe et al., 1994). During later retreat stages the ice stream may have terminated as a tidewater glacier in the Firth of Forth, perhaps similar to those seen in, for example, western Spitsbergen and Alaska at the present day. This stream would have provided the main route by which glacial sediments entered the Forth Basin and so the lithology of these sediments would largely have reflected its substrate.

2.3.2 Events during the Loch Lomond Stadial

During the rapid climatic amelioration that occurred around 13,000 BP (the Windermere Interstadial) the Forth region was deglaciated. Some lines of evidence, including marine molluscan assemblages (Peacock & Harkness, 1990) and coleopteran assemblages (Atkinson et al., 1987) indicate that conditions were only marginally cooler than at the present day. However, this amelioration faltered around 11,000 BP when cold climate conditions (the Loch Lomond Stadial) were re-established over much of northern and western Europe. It is now accepted that this event was linked to a southward displacement of the Atlantic polar front (Ruddiman & McIntyre, 1973) and the consequent deflection of the warm North Atlantic Drift into the Bay of Biscay. The cause of this change is believed to lie in salinity changes in the western Arctic due to melting of the Laurentian ice sheet and the consequent weakening of the Atlantic 'heat conveyor' (Broeker & Denton, 1989; Boulton et al., 1991) although the details remain largely speculative. The stadial was itself short-lived: the climate again ameliorated from about 10,000 BP onwards or slightly earlier, although this is difficult to date accurately due to the existence of a radiocarbon 'plateau' in the early Holocene, as discussed by Tipping (1987) and by Ammann and Lotter (1989). However, the final deglaciation appears to have been complete by 9600 BP. Again, this change is thought to have been brought about by the northward migration of the Atlantic polar front and the renewed influence of the North Atlantic Drift at higher European latitudes.

In Scotland, one major response to this event was the regeneration of a limited mountain ice-sheet over the western Highlands (Sissons, 1974b; 1981; 1983; Sutherland, 1984) (Figure 2.5). Outlet glaciers from this ice-sheet occupied many Highland valleys and extended part-way into the lowlands (a diachronous series of events collectively termed the Loch Lomond Readvance).
Figure 2.5 Area covered by the Grampian ice-cap and other glaciers of the Loch Lomond Stadial (source: Sissons 1983a: figure 14.4).
The Loch Lomond basin was re-occupied and ice discharged eastwards along the line of the former Forth ice stream, creating a major drumlin field in the Drymen area south of Loch Lomond; at its maximum extent the ice reached Menteith, to the west of Stirling, as evidenced by a major moraine ridge known as the Menteith moraine. This feature shows up to \( \sim 18\text{m} (60\text{ft}) \) of relative relief (Sissons, 1966) and extends for \( \sim 20\text{km} \) (Sissons, 1974b), and is largely constructed of fluvio-glacial materials and transported shelly clays. Throughout this period the ice discharged sediment and meltwater into the Forth drainage system, from which have been constructed a series of glaciofluvial features in the area eastwards from Menteith to the contemporaneous marine limit.

The effects of these events in the Bothkennar area are three-fold. Firstly, the severity of the climate initiated a phase of shoreline erosion, evidenced by a planation surface that is known to occupy an area of at least 28 km\(^2\) in the Forth area (Sissons, 1969) and which has been possibly correlated with rock platforms in the west of Scotland (Sissons, 1974a), although this is not certain since such features may be of composite origin (Browne & McMillan, 1984). Secondly, onto this planated surface was deposited an extensive sandy gravel or boulder horizon which occurs immediately beneath the Holocene deposits at Bothkennar. This was originally named the Bothkennar Gravel (Browne et al., 1984) and later placed in a more formal stratigraphy as a facies of the Bothkennar Gravel Formation (Paul et al., 1995). The sediments were originally believed to have been deposited as a gravel lag at or very near the contemporaneous sea-level (Sissons, 1969) and so have been used to constrain the Forth sea-level curve; however, this view has recently been challenged by Peacock (1998) who argued that the sediments may be ice-rafted (an origin first proposed by Cadell (1883)) and if so are not definitively connected to a former sea level.

Thirdly, and of great importance in the development of the sedimentary infill of the Forth estuary, the load imposed by the Loch Lomond ice-sheet interrupted the isostatic recovery that followed the main deglaciation from 13,000 BP onwards. In areas close to the ice-sheet, including the Forth area, this caused an actual reversal of relative sea-level change and allowed an inundation of the estuary as far as about Stirling (although the magnitude of this reversal is now brought into question by the above queries over the origin of the Bothkennar Gravel Formation). During the subsequent isostatic recovery in the early Holocene, a series of raised ‘beaches’ (in reality tidal flats somewhat analogous to the mid- to late-Holocene carse deposits) were formed and subsequently inundated by the mid-Holocene (Flandrian) transgression. These deposits and the sea-level events to which they are related are discussed in section 2.4 below.
2.3.3 Description of Glacial Tills in the Forth Area

Above the marine limit, the predominant surface sediments in the Forth area are terrestrial glaciogenic deposits that largely obscure the solid bedrock over most of the low ground. They are shown on official British Geological Survey maps as boulder clay and/or morainic drift (both are effectively synonyms for till and are distinguished by the nature of their geomorphology at outcrop) or as variations on the theme of glaciofluvial sands and gravels; a broad classification that effectively obscures their genetic relationships. An extract (Figure 2.6) from the 1:625,000 drift geological map (British Geological Survey, 1977) shows their nomenclature and distribution in the Forth area. The tills have the more widespread distribution and have largely provided the material from which the Holocene infill to the estuary has been constructed.

Despite their widespread occurrence, the detailed sedimentology and genesis of these tills appear not to have been widely reported in the literature, other than in the publications of the British Geological Survey (see later in this section). This work, together with additional information kindly supplied by Professor M.A.Paul (pers. comm.), allows certain generalisations to be suggested, although it is stressed that these deposits have not been the subject of detailed study by the present author.

In their unweathered state the tills can be divided into two broad facies, which for convenience can be called informally the brown facies and the grey facies, although it should be noted that the grey facies is brown-weathering at surface. The grey facies is a relatively fine-grained, matrix dominant till (in the terminology of McGown & Derbyshire, 1977), which contains clasts of igneous lithologies such as basalts and related rocks, various low-grade metamorphic rocks and occasional limestones. The brown facies is a matrix dominant to well-graded sandy- to silty-clay till with frequent clasts of red-brown sandstone, probably derived from the Devonian Old Red Sandstone series, together with numerous clasts of those igneous lithologies (largely andesites and related rocks) typical of the Devonian sequences. In its uppermost one to two metres the till is commonly reddened and sandstone clasts are decomposed. Both facies normally extend to the ground surface, where they exhibit a streamlined morphology and are thus interpreted as lodgement tills. It is hypothesised that the brown facies is mainly derived from the Devonian rocks in the north of the area and the grey facies mainly from the Carboniferous rocks of the Midland valley, both facies having an identifiable element from the Dalradian lithologies of the Loch Lomond basin. Thus the grey facies appears to represent a component that has travelled along the axis of the Forth ice stream while the brown facies forms a more lateral component.

If the two facies are correctly identified as lodgement tills of differing provenance, one may expect, by analogy with better studied examples, that they will possess certain sedimentological characteristics. They will be relatively consistent in their composition in terms of both mineralogy and grading. Evidence for this in the area of the Forth Estuary has been provided by
Figure 2.6 Drift deposits of the central area of Scotland. Note position of the Bothkennar site (blue arrow). Basemap: extract from the 1:625,000 Quaternary Map of the United Kingdom, North sheet. (British Geological Survey, 1977).
the work of the British Geological Survey (Smellie, 1981; Costelow & Brown, 1986) where it has been shown that the till here forms a single geotechnical unit. Studies by Browne & McMillan, (1985; 1989) have identified an analogous deposit, which they term the Wildnerness Till, that forms a composite lithological unit over the ground to the east of Glasgow and towards Stirling. The local character of this deposit is dependent on the bedrock from which it is derived: a brown and a grey lithofacies are again developed on Devonian and Carboniferous substrates respectively. Elsewhere, in East Anglia, there is evidence (Perrin et al., 1979) that, although lodgement till is likely to contain a complex suite of minerals, this suite will be similar through a till sheet. Similarly, the particle size distribution of the fine fraction in a matrix dominant lodgement till sheet is often uniform throughout the deposit (Dreimanis & Reavely, 1953; Chryssafopoulos, 1963; Scott, 1976; Sladen & Wrigley, 1983), unless some other process such as aqueous transport has intervened. Furthermore, the overall size distribution of a lodgement till is typically polymodal and these modal sizes again appear consistent: indeed, some authors (Dreimanis & Vagners, 1971) have argued that this is a consequence of the homogeneity of a till sheet, since after some minimum transport distance each mineral type is reduced to a terminal size characteristic of that mineral and finer than which it does not become.

If the till facies do possess the above characteristics, then certain features of the sedimentology of the Claret Formation at Bothkennar (to be described in Chapter Six) are readily explained. This latter deposit has been found to be very uniform in its mineral composition both laterally and vertically across the BSCRS and this uniformity is also seen in the Claret Formation sediments of the Stirling area (Holloway, pers. comm.). The particle size distribution is also fairly uniform, although obviously affected by some aqueous sorting, and the fine end of the distribution is particularly consistent. The distribution also contains characteristic modal sizes which recur in most of the samples, although the relative sizes of the modes can vary. These characteristics are explicable if the Claret Formation is essentially the fine fraction of the (grey) regional till facies, eroded and redeposited in the Forth estuary following the main Holocene transgression.

2.4 HOLOCENE DEVELOPMENT OF THE FORTH ESTUARY

The infill of the Forth estuary records a complex sequence of changes in relative sea-level, which were initiated during the deglaciation from the Devensian glacial maximum and continue throughout the Late-glacial and Holocene. This changing sea-level brought changes in the palaeogeography of the estuary and hence in the wave and tidal regimes, which in turn caused changes in the sediment budget and transport patterns.

2.4.1 Relative Sea-level Change: General Ideas

During a glaciation–deglaciation cycle, events occur on both the local and the global scale that cause sea-level to change relative to the land surface. In glacierised areas, these are related, in
both space and time, to movements of adjacent ice margins and to the patterns of sedimentation around these margins. The mutual relationships of these features provides the basis from which the sea-level history can be reconstructed.

2.4.1.1 Isostatic sea-level change

Ice loading of the continental crust causes a localised isostatic depression of the land surface. It is well-known than an ideal ice cap has an approximately parabolic profile (i.e. it is dome-shaped) and is thus thickest at its centre: the area of maximum loading and greatest depression (Paterson, 1981 and references therein). This depression decreases radially outwards from the centre and in consequence laterally extended, originally horizontal, features such as shorelines exhibit a differential uplift that increases towards the centre of loading. Thus they will 'tilt' upwards towards the isostatic centre. This tilt increases with time as rebound progresses, since it is caused by deep, slow flow of mantle material toward the centre. Thus, at any instant, the older shorelines will generally have a greater slope than the younger ones, a feature which has been used to categorise and group shorelines of differing age (Sissons & Smith, 1965b; Smith, 1965; 1968; Sissons et al., 1966).

There is evidence in the Moray Firth area that the Main Lateglacial Shoreline (MLG), formed during the Loch Lomond (Younger Dryas) Stadial, has a steeper slope than the earlier Late Devensian shoreline, with similar, though less well-developed, evidence in the Forth valley (Firth, et al., 1993). The steeper nature of the younger slope is thought to be due to the redepression of the crust during the Loch Lomond Stadial (Sutherland, 1984; Firth, et al., 1993) and this is contrary to rheological models of Lambeck (1991) which suggest a minimal effect on crustal rebound from the Loch Lomond Stadial readvance.

The simple picture may in practice be further complicated by the presence of a raised forebulge or arch adjacent to the area of loading which subsequently collapses following unloading. However, since it has been predicted (Lambeck, 1996) that this effect would be minimal in Scotland, due to the relatively small dimensions of the British ice sheet, the possibility will not be considered further here.

The cycle of depression and uplift causes a relative rise in sea-level (transgression) followed by a fall (regression). However, since the rate of deglaciation normally exceeds that of the corresponding isostatic recovery, the withdrawal of the ice from coastal areas leads to a temporary transgression which is reversed in due course as a result of continuing isostatic rebound.
2.4.1.2  Eustatic sea-level change

On the global scale, ice sheet formation and deglaciation reduces and increases respectively the volume of water present in the oceans. Where the ice sheet is of sufficient size, this produces a measurable change in global sea-level, which is termed a eustatic sea-level change. However, at any given location the isostatic and eustatic components of sea-level change are not likely to be synchronous. Isostatic change is controlled by changes in local glacierisation, whereas eustatic change is a worldwide response to a change in glacierisation anywhere on the globe. Variations in local climatic response, in the internal dynamics of the ice-sheet and in the coupling of the ice-mantle response are thus very likely to have prevented changes in the two components being synchronous.

In practice, the behaviour of the North American (Laurentide) ice sheet dominated eustatic changes in the Devensian and probably in earlier glacial periods, since it alone is estimated to have accounted for around 60% of the increase in ice volume during the glaciation, perhaps equivalent to 74m of world-wide eustatic sea-level fall (Flint, 1971). By contrast, the effect of the Scottish ice-sheet on world-wide sea-level was negligible. Thus, in Scotland, the two components of relative sea-level change effectively operated independently, coupled only loosely through a common response to changes in global climate.

2.4.1.3  Reconstruction of relative sea-level change

A key tool in this reconstruction is the sea-level curve. This is a time-elevation graph which takes on its y-axis the measured height, with reference to present OD, of various geomorphological or sedimentological features whose elevation relative to the contemporaneous sea-level can be established, and on its x-axis the corresponding ages of these features, normally quoted in $^{14}$C (radiocarbon) years BP (before present, conventionally 1950 AD, corrected by a reservoir age of 405±40 years in the case of shell samples). A calibration to calendar years may also be applied (Stuiver et al., 1986; Pilcher, 1991). These graph points are termed index points and they allow a smooth curve (the sea-level curve) to be drawn showing changing sea-level with time.

Both these measured parameters contain uncertainties¹ and so the index point is in reality an error box. Other dated samples of known OD elevation but unknown relation to the sea level can then be plotted on this graph and their relation to the contemporaneous sea-level inferred. For example, the water depth during the depositional period can be determined from the elevation difference between a dated benthic sample and the sea-level curve at that date.

¹ Uncertainty in the $^{14}$C date arises from the statistical error in the measurement of decay counts (conventional method) or isotopic ratio (AMS method). The uncertainty in the elevation arises from several causes, including principally compaction of the underlying sediments and uncertainty of the relationship of the sample location to mean sea-level or to the tidal frame (cf. Chapter 8, section 8.2).
2.4.2 Effect of Changes in Relative Sea-Level in the Forth Area

The generally accepted sea-level model for the Forth area is based on investigations carried out by J.B. Sissons (formerly of the University of Edinburgh) and his co-workers between about 1962 and 1983. This has led to a very large body of publications (Sissons, 1962; 1963; 1966; 1969; 1971; 1976; Smith, 1965; 1968; Sissons et al., 1966; Kemp, 1971; Robinson, 1993) which was summarised both by Sissons himself (1983) and later by Ballantyne and Gray (1984).

Table 2.1 summarises the main sea-level events, the principal features of the resulting morphology and the sediments associated with them. Also shown are the principal stratigraphical divisions proposed by various authors, including the present writer (Sissons, 1969; Browne et al., 1984; Gostelow & Browne, 1986; Paul et al., 1995; Peacock, 1998; Barras & Paul, 1999). In general, the preferred terminology is that of Paul et al. (1995) or, for those deposits not discussed by Paul et al., that of Browne et al. (1984).

2.4.2.1 Morphological features associated with sea-level change

Between Stirling and St Andrews there are three separate series of raised shorelines which can be recognised on morphological grounds (Sissons, 1983: figure 2). These are reproduced in Figure 2.7.

1. A set of Late-glacial raised shorelines which record the sea-level fall that followed the deglaciation of the Scottish main Devensian ice-sheet from around 13,500 to 11,000 BP (Figure 2.7: points 1 and 2).

2. A series of early-Holocene buried shorelines (the High, Main and Low buried shorelines) and associated sediments (Sissons, 1983: figure 3; Figure 2.7: points 4 – 6) that have been recognised in boreholes around Stirling and can be traced eastwards for some distance. These record sea-level events between about 10,300 BP and approximately 8,500 BP, which followed the withdrawal of Loch Lomond ice and represent the regression that ended at the onset of the main Holocene transgression.

3. Four mid-Holocene raised beaches (Figure 2.7: points 7 - 10) resulting from the culmination of and regression from the main Holocene transgression, an event that was sufficient temporarily to overcome the ongoing isostatic rebound following the disappearance of the much smaller Scottish Loch Lomond Stadial ice cap. The culmination of the transgression occurred around 6,500 BP at the head of the estuary (Robinson, 1993; Sissons & Brooks 1971), at which time the highest Holocene shoreline (FG1) was formed. The subsequent regression across the carse surface left a series of minor shoreline ridges (termed PG2-PG4), as yet undated, which have been studied more completely on the north side of the Forth (Smith, 1965; 1968).
Table 2.1  Subdivision of the Late Quaternary sediments in the area of the Forth estuary.

| Time Period         | Radiocarbon Age (yrs BP) | Principal geomorphic features | Principal sedimentary units | Environment of formation | Reference on Figures 2.7 & 2.8 | Stratigraphical nomenclature | Sissons, 1969 | Browne et al., 1984 | Gostelow & Browne, 1988 | Paul et al., 1995 |
|---------------------|--------------------------|-------------------------------|----------------------------|--------------------------|-------------------------------|------------------------------|---------------------------|-------------------|-------------------|---------------------|----------------------|
| ~11,000 to 10,300   |                          | Sandy gravel layer            | Ice rafted debris          | B                        | Buried gravel layer           | Bothkennar Gravel            | D                         | Bothkennar Gravel   |                   |                     |                      |
| ~10,500 Nominally   |                          | Buried/poorly exposed cliff-line | Erosion under cold climate conditions | 3                        | Main Lateglacial shoreline    | -                            | -                        | -                 | -                 | -                   |
| ~13,000 to 11,000   |                          | Stoney or laminated clays     | Cold water probably glaciomarine | A                        | -                             | Abbotsgrange, Kinnell Kerse, Loanhead beds | E, F                     |                   | -                 | -                   |                      |
| ~10,300; 9,600; 8,800|                          | Peat                          | Saltmarsh to freshwater bog | D                        | Sub-carse peat                | -                            | -                        |                   |                   | -                   |
| ~6,500              |                          | Prominent cliff               | Culmination of transgression | 6                        | Main Postglacial Shoreline    | -                            | -                        |                   |                   | -                   |
| ~4,000              | Subdued steps on coarse surface |                              | Regressional beach levels | 7-9                      | PG2-PG4                       | -                            | -                        |                   |                   | -                   |
| 8,300 to 3,000      | Dark coloured silty clays to clayey silts | Subtidal becoming intertidal |                              | E                        | Carse clays                   | Claret Beds, Letham Beds    | A(ii)                    | B, C              | Claret Formation   |                      |
| 10,300; 9,600; 8,800|                          | Fine sand to silty clay       | Sheltered littoral to sublittoral | C                        | High, Main and Low Buried Beaches | Letham Beds                  | C                         |                   |                   | -                   |
| ~3,000              | Thin clayey silts with shells | Thin clayey silts with shells | Beach veneer to lagoonal   | F                        | -                             | A(i)                         | Grangemouth Formation      |                   |                   |                     |
Figure 2.7 Height-distance diagram of shorelines in south-east Scotland illustrating their intercorrelation with the stratigraphy and sea-level history shown in Figures 2.8 and 2.9 (source: Sissons, 1983b: figure 2 with correlations added). The enlarged section (above) shows dislocations in the Main Postglacial Shoreline (upper) and Main Buried Shoreline (lower) surfaces from the western Forth valley (source: Sissons, 1983b: figure 6).

Figure 2.8 Schematic illustration of the morphology and stratigraphy of sediments at the head of the Forth estuary, illustrating their intercorrelation with the shorelines and sea-level history shown in Figures 2.7 and 2.9 (source: Sissons 1983b: figure 3 with correlations added).
2.4.2.2  Sediments associated with sea-level change

The sea-level events that created the surface morphology in the Forth area also introduced the sedimentary infill that now occupies the upper estuary. By contrast with the rather complex surface morphology, the architecture of this infill is fairly simple and gives rise to a useful stratigraphical framework. Figure 2.8 shows this architecture in schematic form: on this figure the sedimentary units have been lettered A-F for convenience. The sediments fall naturally into three groups which relate to the various beach levels and thus to the history of relative sea-level change. Due to ongoing isostatic uplift some later units have been incised into earlier units and so the units do not always follow the normal rules of stratigraphical superposition, younger units being found at lower elevations than older ones.

The first group of sediments, shown undifferentiated as [A], are of Late Devensian age and directly overlie till. Their deposition commenced when the Forth was in direct contact with glacier ice. They have been divided into the Loanhead, Kinneil Kerse and Abbotsgrange beds by Browne et al. (1984). The Loanhead Beds are soft to firm brown to brownish grey, micaceous plastic silty clays with thin sandy bedding and contain a cold-water fauna. They have been recorded in the Bothkennar area in all boreholes that have reached the till basement. They are succeeded by the Kinneil Kerse Beds, which are mainly grey, soft to firm clayey silt together with thin bands of silty clay, some mottling and sandy laminations. The fauna indicates depositional conditions similar to the Loanhead Beds. The Kinneil Kerse Beds are absent from those boreholes near the margins of the basin. In the area around Grangemouth are found the Abbotsgrange Beds, which are loose, well laminated black to grey and brownish grey silts with thin bands of soft to firm, dark grey silty clay, that contain a temperate fauna.

The succession was deposited under an increasing amelioration of the climate from cold water conditions (Loanhead, Kinneil Kerse and base of Abbotsgrange beds) to more temperate climatic conditions but still much colder than those at the present (Browne et al., 1984). The Loanhead Beds are considered to be the product of fjord-based glaciomarine sedimentation and the Kinneil Kerse Beds to have been deposited from intermittent melt water discharge, transported down a delta front into a deep water extra-delta basin. The Abbotsgrange Beds are interpreted as the prodelta deposits of a seaward-prograding delta, in shallower water than Loanhead and Kinneil Kerse beds.

These sediments are cut by the planation surface of Loch Lomond Stade age, which also cuts across till and bedrock. On this surface rest sediments of the Bothkennar Gravel Formation [B] which comprise coarse gravel, cobble and boulder sized clasts with a matrix of loose sand to clayey sandy silt. This deposit marks the top of the Late Devensian sequence in the area and is believed to date from between 10,300 and 11,000 BP. It has been recently suggested to be of
ice-rafted origin (Peacock, 1998), in contrast to the previous interpretation as a lag or beach deposit (Sissons, 1969).

The sediments of the second group [C] are found only at the head of the estuary around Stirling, where they rest directly on the Buried Gravel Formation (BGF). They comprise a succession of silt and fine sand, on whose surface are developed a series of buried steps known as the High, Main and Low buried beaches. These beach surfaces are in turn overlain by saltmarsh and freshwater peats [D] that have been inferred to date from between 10,300 BP and 10,100 BP (High), 9,600 BP (Main) and dated at 8,690±140 BP (Low) (Sissons & Brooks, 1971). Beneath the High Buried Beach the sands are pink in colour, elsewhere they are grey to pale yellow. Beneath the low beach the sands are overlain by grey clays. The sediments are interpreted as estuarine beach deposits, grading down to tidal flat muds beneath the low beach. Browne et al. (1984) have proposed the name Letham beds for these and similar deposits but unfortunately they use the name indiscriminately for any slightly sandy deposit that rests on the Bothkennar Gravel Formation. This introduces problems of correlation and it is considered that the name should instead be restricted to those sediments on the estuary margins which underlie peat of early-Holocene age and are thus demonstrably part of the buried beach suite.

The third group of sediments [E and F] overlies the buried beaches and elsewhere rests directly on the Bothkennar Gravel Formation. They are silty clays and clayey silts, variously bedded, mottled or laminated, that are collectively referred to by the informal name of 'carse clay'. These later Holocene sediments have been assigned formally to two formations (Paul et al., 1995): the Claret Formation [E], which comprises the main body of bedded or mottled silty clays/clayey silts and the Grangemouth Formation [F], which comprises a mixed suite of tidal channel, intertidal and modern reclamation deposits which lie unconformably on the Claret Formation. These deposits extend to the present-day surface unless concealed by made ground. The sediments of the Claret Formation were deposited during and after the main Holocene transgression and subsequent regression; their deposition being initiated between about 8,000 BP and 7,500 BP at successively later times the farther up the estuary (Sissons & Brooks, 1971). In the Bothkennar area itself the majority of the sequence was deposited later than about 5,000 BP (Paul et al., 1995). Apparently little deposition occurred prior to this date and therefore the transgressive phase is not well-represented at Bothkennar. Deposition ceased as the subsequent regression progressed, at around 6,500 BP at the head of the estuary and at around 3,000 BP in the Bothkennar area.

2.4.2.3 The relationship of shorelines and sediments to sea-level events

The sea-level curves published by Sissons and Brooks (1971) and Robinson (1993) for the western Forth area are shown in Figure 2.9. The sea-level events shown on this curve were accompanied by formation of the above shorelines and sometimes by the deposition of associated sedimentary units. For clarity, on Figure 2.9 the events and the corresponding morphological or
Figure 2.9 Sea-level curves for the western Forth valley illustrating the inter-correlation with the shorelines and sea-level history shown in Figures 2.7 and 2.8. The curves are based on Sissons & Brooks (1971) and Robinson (1993). Black roman numerals [4] to [9] refer to the morphological features and letters [C] to [F] to sedimentary units.
### Table 2.2  Index points relating to Sissons and Brooks (1971) sea-level curve (cf. Figure 2.9)

<table>
<thead>
<tr>
<th>Index Point</th>
<th>Event or Geomorphological feature</th>
<th>O.S. Grid Ref.</th>
<th>m OD</th>
<th>Date BP</th>
<th>Position of sample /sediments</th>
<th>Interpretation/explanation</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>High Buried Beach (HBB) / Abandonment of High Buried Shoreline (HBS).</td>
<td>1</td>
<td>12.2</td>
<td>10300-&lt;br&gt;10100&lt;sup&gt;2&lt;/sup&gt;</td>
<td>As the beach does not occur within the Menteith moraine and its deposits lie partly on buried outwash deposits from the moraine, it is inferred that the beach ceased to be formed after the moraine had built up but before significant glacier decay. It probably ceased to accumulate at or shortly after the end of pollen zone III&lt;sup&gt;3&lt;/sup&gt;. The complete absence below the HBS of pink/brown highly micaceous deposits of which the HBB is composed marks this as a transgressive maximum.</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Main Buried Shoreline (MBS) / Abandonment of Main Buried Beach (MBB)</td>
<td>4</td>
<td>9.9</td>
<td>~9600</td>
<td>The uppermost MBB deposits and overlying peat was analysed at 5 sites&lt;sup&gt;6&lt;/sup&gt;.&lt;sup&gt;6&lt;/sup&gt;</td>
<td>Regression of sea from MBB at the end pollen zone IV or IVV junction&lt;sup&gt;3,5,8&lt;/sup&gt; is indicated.</td>
</tr>
<tr>
<td>3a</td>
<td>Low Buried Beach (LBB) / Abandonment of Low Buried Shoreline (LBS)</td>
<td>NS630 960</td>
<td>7.2</td>
<td>8690±140</td>
<td>Basal 2cm of peat overlying LBB</td>
<td>Pollen and macro-plant remains indicate salt marsh conditions during the final stages of accumulation of the LBB, followed by further regression marked by the development of freshwater swamps and woodland&lt;sup&gt;5&lt;/sup&gt;. Regression of sea-level from LBB very shortly before this date.</td>
</tr>
<tr>
<td>Minimum of regression</td>
<td>-</td>
<td>-</td>
<td>~8500</td>
<td></td>
<td>Sea restricted to well-marked buried channel of the Forth&lt;sup&gt;7&lt;/sup&gt;. Evidence from index points 3a &amp; 3b suggests a minimum sea-level at this date.</td>
<td></td>
</tr>
<tr>
<td>3b</td>
<td></td>
<td>NS630 960</td>
<td>7.68</td>
<td>8270±160</td>
<td>Top 1cm of peat overlying LBB</td>
<td>Transgression due to melting of N. American ice cap. Scotland undergoing isostatic uplift, but not keeping up with sea-level rise.</td>
</tr>
<tr>
<td>4</td>
<td>Main Postglacial Transgression (MPT)</td>
<td>NS631 961</td>
<td>9.5</td>
<td>8010±130</td>
<td>Top 2cm of thin peat layer overlying LBB</td>
<td>Associated with transgression: peat = Phragmites &amp; Carex; pollen = high values of Cyperaceae + limited Chenopodiaceae</td>
</tr>
<tr>
<td>5</td>
<td></td>
<td>NS582 952</td>
<td>13.3</td>
<td>7490±125</td>
<td>Peat/clay transition</td>
<td>Associated with transgression: pollen = high values of Cyperaceae, Gramineae &amp; Chenopodiaceae</td>
</tr>
<tr>
<td>6a</td>
<td>Main Postglacial Shoreline (MPS) / Raised Beach (MPRB)</td>
<td>&gt;6400±125</td>
<td></td>
<td></td>
<td>The main postglacial maximum was slightly before 6400±125. See index point 6a for evidence.</td>
<td></td>
</tr>
<tr>
<td>6b</td>
<td>Regression following culmination of the MPT</td>
<td>NS560 955</td>
<td>14.4</td>
<td>6490±125</td>
<td>Basal 10cm of peat overlying carse clay</td>
<td>Sea-level had begun to fall from its postglacial maximum. Peat changes from remains of Phragmites and Carex to Sphagnum at top of peat as freshwater conditions prevailed.</td>
</tr>
<tr>
<td>7</td>
<td></td>
<td>NS560 955</td>
<td>14.6</td>
<td>6135±105</td>
<td>Wood from 10cm thick branch of birch.</td>
<td>Wood sample dated to confirm date of point 6a in order to rule out contamination during sampling.</td>
</tr>
</tbody>
</table>

---

1 The altitude for index point 1 was taken from an average of twenty-three measurements on the horizontal shoreline of the High Buried Beach, immediately east of the Menteith moraine (Sissons & Brooks, 1971).

2 The date of the High Buried Beach/abandonment of the High Buried Shoreline, was not obtained directly from radiocarbon dating. See interpretation in Table 2.2 (Sissons & Brooks, 1971).

3 This refers to the standard system of pollen zones (Godwin, 1975). The quoted dates assume that pollen zones III, and IV or junction of IVV assumes that these zones in the western Forth valley are approximately synchronous with the same boundaries at Scaleby Moss, northern England (Godwin et al., 1957; Sissons & Brooks, 1971).

4 The altitude of index point 2 was taken from an average of sixteen measurements on the Main Buried Shoreline where it is horizontal immediately east of the Menteith moraine (Sissons & Brooks, 1971).

5 Newey (1966).

6 Brooks (1972).

7 Sissons (1966).

8 The location of index point 7 lies about 12km distant from the other sampling points. The up-arrow on the sea-level curve represents the probable amount of correction necessary to take isostatic tilting into account (Sissons & Brooks, 1971).

9 This date probably does not date the marine regression. Trees found lying under the peat were probably felled by human beings. This probably resulted in the formation of blanket bog (represented by the overlying peat). Marine regression would have occurred sufficiently long before the quoted date for large trees to grow and, therefore, sea-level at this date is plotted lower than the 10.7m altitude of the dated sample (Sissons & Brooks, 1971).
Table 2.3 Index points relating sea-level curve for western Forth valley (West Flanders Moss) (cf. Figure 2.9). All data Robinson (1993) (except where referenced otherwise).

<table>
<thead>
<tr>
<th>Index Point</th>
<th>Event or Geomorphological feature</th>
<th>O.S. Grid Ref. NS:</th>
<th>Elevation m OD</th>
<th>SURRC Radiocarbon Laboratory</th>
<th>Interpretation using pollen analyses</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Abandonment of the Main Buried Beach (MBB)</td>
<td>5500 9640</td>
<td>11.13 - 11.14</td>
<td>SRR-1250 9625±100</td>
<td>Basal subcarse peat overlying MBB deposits. Saltmarsh was succeeded by grasses and herbs: the shoreline receded and a damp freshwater environment became established.</td>
</tr>
<tr>
<td>2</td>
<td>Abandonment of the Low Buried Beach (LBB)</td>
<td>5816 9601</td>
<td>9.54 - 9.53</td>
<td>SRR-1428 8960±50</td>
<td>Basal subcarse organic deposits of peat layer overlying MBB deposits. Relates to saltmarsh peat that formed prior to the freshwater Corylus/Myricastrata rise in this locality. Sample about 1.5m below the MBS: accounts for the younger date than that at point 1. Using index points 1 &amp; 2, a fall in relative sea level of 20 to 33cm 100^1 years was estimated (assuming correct dates and no delay in the onset of organic deposition).</td>
</tr>
<tr>
<td>3</td>
<td>Abandonment of the Low Buried Beach (LBB)</td>
<td>6351 9611</td>
<td>7.10 - 7.11</td>
<td>SRR-1429 8680±50</td>
<td>Basal 1cm of subcarse peat overlying LBB deposits. Refers to early phase of transition from beach sediments to freshwater terrestrial peat conditions: initiated by falling sea-level as it abandoned the LBB. (Estimated parameters as the sampling site was about 6km east of West Flanders Moss).</td>
</tr>
<tr>
<td>4</td>
<td>Main Postglacial Transgression (MPT)</td>
<td>5939 9646</td>
<td>11.35 - 11.36</td>
<td>SRR-1423 7850±60</td>
<td>Transitional gyttja^2 at the base of the carse clay. Indications of a change from freshwater conditions to brackish conditions consequent upon the rising sea-level of the Main Postglacial Transgression.</td>
</tr>
<tr>
<td>5</td>
<td>Main Postglacial Transgression (MPT)</td>
<td>5749 9509</td>
<td>13.55 - 13.57</td>
<td>SRR-1812 7060±50</td>
<td>The uppermost 2cm of peat at the base of the carse clay. This sampling position was near the former shoreline of the Main Postglacial Transgression where the carse clay thins to a narrow wedge at the edge of the deep peat overlying Moss.</td>
</tr>
<tr>
<td>6</td>
<td>Culmination of the MPT</td>
<td>-</td>
<td>14.9</td>
<td>-</td>
<td>at or before 6900</td>
</tr>
<tr>
<td>7</td>
<td>Culmination of the MPT</td>
<td>577 962</td>
<td>13.35 - 13.36</td>
<td>SRR-1427 6850±50</td>
<td>Base of gyttja overlying carse clays. This site was probably freely drained permitting relatively early organic accumulation as the sea-level began to recede.</td>
</tr>
<tr>
<td>8</td>
<td>Culmination of the MPT</td>
<td>5597 9655</td>
<td>14.310 - 14.325</td>
<td>SRR-1426 6590±60</td>
<td>Gyttja containing some wood fragments overlying carse clay. At this sampling site, pollen evidence suggests that organic deposition was probably delayed by waterlogging.</td>
</tr>
<tr>
<td>9</td>
<td>Regression following culmination of the MPT</td>
<td>5749 9509</td>
<td>13.90 - 13.92</td>
<td>SRR-1811 6520±50</td>
<td>2cm of gyttja at upper minerogenic/organic transition (above carse clay). This sampling position was near the former shoreline of the Main Postglacial Transgression where the carse clay thins to a narrow wedge at the edge of the deep peat overlying Moss.</td>
</tr>
<tr>
<td>10</td>
<td>Regression following culmination of the MPT</td>
<td>14.2</td>
<td>5492±130</td>
<td>Q-533</td>
<td>10-12cm above carse clay surface. This sample was included to give an estimate of the trend of the sea-level curve at this time and gives only a minimal age estimate to this part of the curve.</td>
</tr>
</tbody>
</table>

---

1 Radiocarbon dates reported unaltered in conventional radiocarbon years BP, the standard error being ±1 Standard Deviation.
2 Gyttja - this refers to fine-grained detritus sediment comprising organic and minerogenic materials. It usually occurs as a transitional element between peat and minerogenic sediment, notably the lower (and upper) contacts of the carse clay.
3 Estimates of peat compaction had been made by Cullingford et al., (1980) for the Tayside region which ranged from 40 to 68%. The altitude at the time of sampling was, therefore, unlikely to reflect the original position. Robinson's sea-level curve for the western Forth valley shows a vertical arrow (index points 5, 7, 8 and 9) which represents compaction of 50% at the sampling sites.
4 Carse clay at this sampling point was underlain by about 1.5m of peat which was probably subject to some degree of compaction. Robinson (1993) included this point in the sea-level curve, but the up-arrow demonstrates the uncertainty of it's altitude.
5 This dated sample (Godwin & Willis, 1962) from the East Flanders Moss (6km NE of the West Flanders Moss) was included to give an estimate of the trend of the sea-level curve at this time. The dated sample was taken from 10-12cm above the carse clay surface which was reported by D.E. Smith to lie between 13.8 and 14.9m. An approximate level of 14.2m was used for index point 10 on the sea-level curve (D.E. Smith Pers. Comm. to M. Robinson (Robinson, 1993)).
6 Sissons & Smith (1965a).
sedimentological features shown on Figures 2.7 and 2.8 have again been numbered using the common system of numerals [1] – [11] and letters [A] – [F] already introduced. The sea-level curves represent the time interval from ~10,500 to ~4,000 BP only and therefore not all these units are shown on this figure. The details of the index points used on these curves are given in Tables 2.2 and 2.3.

There was a fall in relative sea level as the land rose following the main deglaciation. During this time the highest Late-glacial shorelines [1] and [2] were formed and the sediments of the Loanhead, Kinneil Kerse and Abbotsgrange beds [A] were deposited. This fall was halted by the growth of the Loch Lomond ice sheet which caused a rise in relative sea-level, probably due to a retardation of uplift or a redepression of the crust (Firth et al., 1993). Around this time the widespread planation surface and the corresponding Main Late-glacial shoreline [3] were formed. The sediments of the Bothkennar Gravel Formation [B] which overlie this surface were also deposited at this time or slightly later. Sea-level then rose to the height of the High Buried Shoreline [4]. There was then a punctuated fall in relative sea-level, during which the Letham beds were deposited [C] and the Main and Low Buried Beaches were formed [5], [6]. On these beach surfaces was developed the sub-carse peat [D]. From around 8,000 BP there was a substantial eustatic rise due to the melting of the Laurentian ice sheet, which overtook the isostatic recovery and caused a rise in sea-level (the main Holocene transgression) which culminated with the formation of the Main Post-Glacial Shoreline PG1 [7] at around 6,500 BP. Continuing isostatic recovery then caused a (possibly punctuated) fall in relative sea-level which continues to the present day. Both the transgression and regression were accompanied by the deposition of the Claret Formation [E]. During the general regression minor fluctuations or stillstands in sea-level have allowed the series of beach levels PG2-PG4 to be formed [8] - [10].

During this period the beach veneers and channel deposits of the Grangemouth Formation [F] were deposited. The sequence is completed by the modern shoreline [11].

2.5 HISTORICAL LAND RECLAMATION IN THE FORTH VALLEY

Although the shortage of land is perceived to be a recent problem, particularly in Scotland with its wide open spaces, a programme of land reclamation was initiated along the Forth River valley in Scotland as early as 1766, since at that time prime sites in close proximity to navigable waterways were a very important resource. Prior to 1850, all the land reclaimed in this area was used for agriculture; after this date reanimations were intended primarily for industrial purposes. A number of these land reclamation projects have been described by Udny (1831) and by Cadell (1913; 1929a) from which the following account has been drawn.

Land in the Vale of Menteith to the west of Stirling had been denuded of mixed forests by the Romans in the second century AD, resulting in a swampy, alluvial plain covered by peat, which continued to accumulate over the next fifteen centuries to a final depth of 3m to 4.5m. Above
Stirling the resulting flat, low-lying (carse) deposits extend westwards for about 25km and are 4km to 5km broad, covering an area of about 105 km², about three-quarters of which has been produced by reclamation. West of Stirling, the reclamation was undertaken by individual landowners, although the physical work was carried out by tenants who were each allocated about 4 hectares (10 acres). For example, work on the Blairdrummond Estate commenced in 1766 and continued until 1839; a total of nearly 1400 acres was reclaimed on this estate alone.

Table 2.4 gives a summary of the reclamation areas in the Forth Valley above Stirling and Figure 2.10 shows their locations (Cadell, 1929a: figure 6).

The reclamation procedure in the non-tidal areas consisted of stripping off the layer of peat and draining the soil to restore the land to a productive condition. Channels, which drained into the Forth, were cut into the peat down to the level of the clay and the pulpy part of the peat that was not suitable for use as fuel was thrown into them to float away downstream. This discarded peat caused considerable nuisance to both riparian inhabitants and to local fishermen, since it was usually redeposited in intertidal areas, being found as far away as Inverkeithing. Considerable pollution was thus caused and fishing nets, oyster and mussel beds were damaged (McLusky, 1978; 1987). The floating process continued until 1865 when it was finally stopped. It should be noted that the location of the final resting-place of this not inconsiderable volume of peat could well pose a problem for future stratigraphers.

2.5.1 Land Reclamation in the Tidal Stretch of the Forth Valley

Between Alloa and Bo'ness there is a strip of low-lying land which slopes gradually down to the estuary. The tidal range below Alloa is 5.5m at spring tides, and up to 6m at equinoctial tides (Hydrographer of the Navy, 1999). Due to this high tidal range, in some areas, the foreshore includes a wide fringe of salt marsh which is inundated at high water spring tides but is exposed at low water spring tides and during neap tides.

After the middle of 18th century, when the industrial revolution began in Scotland, riparian owners of these saltings decided to increase the area of their arable land. Due to the tidal influence, the procedure for reclamation needed to be markedly different from that carried out further west. Landowners added to their estates by using the Dutch principle of building sea-dykes in order to exclude the tide from fields along the foreshore.

Table 2.5 gives a summary of the dates and extent of the reclamation work to the east of Stirling and Figure 2.11 shows their locations on a map produced by Cadell (1929a: figure 10). These reclamations are visible on the 1:24,000 aerial photographs for the area (Royal Commission on the Ancient and Historical Monuments of Scotland (RCAHMS), 1971) and are delineated on a mosaic of these on Figure 2.12.
### Table 2.4 Summary of Land Reclamation in the Forth Valley in non-tidal areas West of Stirling (data source: Cadell, 1929a)

<table>
<thead>
<tr>
<th>Area Identification on Figure 2.10</th>
<th>Location</th>
<th>Area Reclaimed (acres)</th>
<th>Date of Reclamation</th>
<th>Landuse</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Vale of Menteith - Blairdrummond estate</td>
<td>1400</td>
<td>1766 – 1839</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Meikelwood estate</td>
<td>130</td>
<td>1828 – 1832</td>
<td>Agricultural</td>
</tr>
<tr>
<td>3</td>
<td>Flanders Moss (below Thornhill)</td>
<td>170</td>
<td>1832 – 1841</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>Goodie Water</td>
<td>75</td>
<td>before 1841</td>
<td></td>
</tr>
</tbody>
</table>

### Table 2.5 Summary of Land Reclamation in the Forth Valley to the East of Stirling (data source: Cadell, 1929a; 1929b)

<table>
<thead>
<tr>
<th>Area Identification on Figure 2.11</th>
<th>Location</th>
<th>Area Reclaimed (acres)</th>
<th>Date of Reclamation</th>
<th>Landuse</th>
</tr>
</thead>
<tbody>
<tr>
<td>1a and 1b</td>
<td>Dunmore &amp; Letham Moss (inland area near Airth)</td>
<td>100</td>
<td>1767 - 1792</td>
<td>Agricultural</td>
</tr>
<tr>
<td>2</td>
<td>Kinneil Kerse (west of Bo'ness)</td>
<td>450</td>
<td>1774</td>
<td>Agricultural originally, now mostly industrialised</td>
</tr>
<tr>
<td>3</td>
<td>West Kerse (at the mouth of the Carron River and including the area of the Bothkennar Soft Clay Research Site)</td>
<td>200</td>
<td>1784</td>
<td>Agricultural</td>
</tr>
<tr>
<td>4</td>
<td>Airth - Dunmore estate foreshore</td>
<td>130</td>
<td>1790 - 1841</td>
<td>Agricultural</td>
</tr>
<tr>
<td>5</td>
<td>- Airth Estate foreshore</td>
<td>185</td>
<td>1790 - 1841</td>
<td>Agricultural</td>
</tr>
<tr>
<td>6</td>
<td>Higgins Neuk (at old ferry pier, Kincardine Bridge, south side)</td>
<td>30</td>
<td>1827</td>
<td>Agricultural and pier construction (sea-wall was never built and area remains a salt marsh)</td>
</tr>
<tr>
<td>7</td>
<td>Allca to Kennetpans</td>
<td>220</td>
<td>before 1840</td>
<td>Agriculture</td>
</tr>
<tr>
<td>8</td>
<td>Kennetpans to Kincardine - Tulliallan estate</td>
<td>152</td>
<td>1821 - 1822</td>
<td>Agriculture</td>
</tr>
<tr>
<td>9</td>
<td>Kincardine to Longannet Point - Tulliallan estate</td>
<td>214</td>
<td>1829 - 1838</td>
<td>Agriculture</td>
</tr>
<tr>
<td>10</td>
<td>Grangemouth</td>
<td>260</td>
<td>1904</td>
<td>Industrial</td>
</tr>
</tbody>
</table>

1 1841 is the date of the New Statistical Account of the parishes of the Forth. Where the date of completion of a reclamation project is uncertain, 1841 is quoted as the completion date, even though this may have been some time prior to 1841.
Figure 2.10  The Forth Valley above Stirling, showing the reclaimed land and existing Peat Mosses, (after Cadell, 1929a: figure 6). Circled roman numerals refer to Table 2.4.

Figure 2.11  Land reclamation in the Forth valley to the east of Stirling (after Cadell, 1929a: figure 10). Circled roman numerals refer to Table 2.5.
Joseph Udney (1831) gave an account of how this reclamation was carried out. Firstly, a line of fir stakes 7 feet (about 2m) in length and 4 feet (about 1.2m) apart were driven about 4 feet into the mud, leaving 3 feet (nearly 1m) protruding above the mud. These stakes were then interlaced with fir brushwood, creating a backwater which enabled sediment-laden tidal waters to flow through, but was sheltered enough to allow the sediment to settle. It has been reported that, behind this first line of stakes, the sediment built up to a thickness of 3 feet up to the top of the stakes, within a year (Udney, 1831). A further line of stakes was driven in after about a year, 4 feet (1.2m) inside the first line, and so on, until the embankment was brought up to the height of the land upon the shore.

A number of reclamation projects carried out in this manner have been reported along the foreshore within the tidal section of the Forth Valley on both the north and south banks (Cadell, 1929a; 1929b).

On the southern foreshore, in the parish of Airth, between 1790 and 1841, an area of 130 acres and 185 acres, were reclaimed on the Dunmore and Airth estates respectively (Figure 2.11: localities 4 and 5). At Higgins Neuk (Figure 2.11: locality 6) at the site of the present Kincardine Bridge, an area of about 30 acres was reclaimed behind a line of stakes in 1827. However, the salt marsh appears to have subsequently prograded further into the estuary, possibly due to the later construction of the Kincardine Bridge in 1936 (Figure 2.13). Although this area is shown as reclaimed by Gostelow & Browne (1981: sheet no. 6), according to Cadell (1929a) the reclamation work was in fact abandoned. This is illustrated by Figure 2.13 which shows that although the saltmarsh is raised above the intertidal mudflats that extend out into the present estuary, the seawall was never constructed along this section. Photographs taken by the author show that this area is in fact inundated by estuarine waters at high water spring tides (Figure 2.14(a)) and remains flooded for about one to two hours at these times. Figure 2.14(b) shows the corresponding spring tide low water mark. At neap tides the salt marsh is exposed during the whole of the tidal cycle (Figure 2.14(c)). Presumably, with the construction of a sea-wall, this whole area could have been reclaimed.

2.5.2 Land Reclamation at the Bothkennar Soft Clay Research Site

It appears from historical reports (Udney, 1831) that an area including the Bothkennar Soft Clay Research Site was reclaimed about the year 1784, when 200 acres (about 81 hectares) was reclaimed on the West Kerse foreshore in the parish of Bothkennar (Figure 2.12: locality 3). Here the embankment was raised to a height of 9 feet (2.7m) over a three-year period, the whole area put to rapeseed during the fourth year and farm-houses built the following year. This gives an indication of the speed of formation of a hardened crust with an increased strength, as without this soil ripening process the building of dwelling-houses would have been impossible.
At the Bothkennar site the sea-wall, or dyke, is shown in Figure 2.15. A line of flotsam (Figure 2.16) which appeared above the HWST mark was probably deposited as a result of raised equinoctial tides or storm action. Figure 2.17 shows the contour levels at ground surface at the BSCRS and, since the mean high water ordinary spring tide (MHWST) level is at 2.95m OD (Hydrographer to the Navy, 1999), it can be clearly seen that most of the site would be inundated if the embankment were to be breached. Indeed, tides only about 0.15m above MHWST would cover the whole site.

In this respect there is concern over a recent application by Scottish Coal to extract coal (seam K2) which extends under the northern half of the Bothkennar site from about 550m west of the site at about 525m depth. This application is intended to safeguard the future supply to Longannet Power Station but is in direct contravention of the understanding by the former SERC when the site was first purchased in 1986, part of the original specification being that the site chosen should not be affected by future subsidence. Recent calculations by International Mining Consultants Limited (D.M.Wood, pers. comm.) estimate possible subsidence to be up to 0.85m offshore and up to about 0.55m onshore. They predicted a bowl of settlement across the Bothkennar site with a maximum of about 0.5m towards the north of the site and decreasing to less than 0.05m at the southern end of the site. They note, however, that these settlements have been calculated using procedures that are standard in the mining industry but that settlements at Longannet have generally been lower than those predicted.

On the basis of the predicted settlements certain remedial measures are proposed that would maintain the drainage of the area and reduce the likelihood of flooding which has been calculated for a 1 year return period rainfall event. It is stated that at present the flood embankments provide protection against 10 year return period events.

It is noted in the report that many of the drainage ditches in the area are not maintained in good condition. This is entirely correct. As example, the writer has observed one such ditch that has been dug around the inside of the embankment at the site in order to facilitate drainage. Although this drainage ditch was thought to be cut off from any tidal water ingress, it appears that the one-way tideflap valve which should allow drainage out of the site only, is faulty. Figure 2.15 shows the drainage ditch in the south-east corner of the site at about one hour before low water neap tide whilst Figure 2.18 shows the same position about two hours after high spring tide, when water had obviously entered the drainage ditch. Observation of the ditch confirmed that water appeared to be entering at the valve, rather than from any other source, such as under or through the embankment. Figure 2.19 shows the valve on the outside of the embankment over an hour after high spring tide, with water still gushing out of it. Although there appears to have been quite a large volume of water entering the drainage ditch, this water
Figure 2.12  Aerial photo mosaic showing the line of reclamation in the Kincardine-Grangemouth area (Royal Commission on the Ancient and Historical Monuments of Scotland, 1971).

Figure 2.13  Oblique aerial photo over the mudflats of the Forth estuary (facing south-west). Note the wide expanse of salt marsh adjacent to the Kincardine Bridge and position of the BSCRS. (Photograph: BKS Surveys Limited, 1974)
Figure 2.15  Sea-wall at the BSCRS showing the drainage ditch at ~1 hour before low water neap tide (facing -NNW).

Figure 2.16  Sea-wall at the BSCRS indicating the line of flotsam on the sea-wall above the high water spring tide mark (facing -NNW).
Figure 2.17 Map of the BSCRS showing the ground surface level contours with respect to OD. (after Nash & Lloyd, 1988b: figure 2). That part of the site below 2.95m OD (i.e. east of the red contour) would be inundated at HWOST if the flood bank were to be breached.
Figure 2.18 Flooded drainage ditch in south-eastern corner of BSCRS at ~2 hours after high water spring tide (facing ~north).

Figure 2.19 'One-way' tide-flap on the estuary side of the embankment draining water into the channel ~2 hours after high water spring tide.
does not actually flood the field due to the depth of the ditch and it is probable that any effect would be felt only at the western edge of the site.

2.6 SUMMARY

The geological background to the Bothkennar Soft Clay Research Site has four aspects: the bedrock geology of the hinterland from which the Holocene sediments originated; the glacial events which brought the precursor glaciogenic sediments into the Forth basin; the changes in relative sea-level that created the estuarine fill of which the sediments at Bothkennar are a part; and the artificial reclamation which created the site in its present condition.

The rocks of the hinterland comprise the Dalradian metamorphics of the southern Highlands; the Devonian Old Red Sandstones plus associated lavas; and a suite of mixed sediments and lavas of Carboniferous age. In colour and mineralogy these rock types can be matched both to elements of the glaciogenic sediments and to the estuarine fill.

The glacial sediments of greatest significance are the lodgement tills of the Forth valley, which are believed to mark the route of a major ice stream that discharged from the southern Highlands. These tills provided the immediate source materials for the estuarine fill and from them the Claret Formation acquired its mineralogy and certain aspects of its grading.

The sedimentology and architecture of the estuarine fill were controlled by a relatively complex pattern of sea-level change in the Forth. This in turn was controlled by four principal events (in chronological order): the unloading and subsequent rebound of the crust due to the retreat of the main British ice sheet from around 13,500 BP; the reloading and subsequent unloading of the crust due to the growth/decay of the Scottish Loch Lomond Stadial ice sheet between about 11,000 and 10,000 BP; a eustatic rise due to the melting of the Laurentian ice sheet that commenced around 8,500 BP; the continued slow isostatic recovery following the completion of the Laurentide eustatic event around 6,500 BP.

In the late 18th Century mudflats were reclaimed for agricultural use along much of the tidal and non-tidal stretches of the Forth. At Bothkennar this occurred around 1784 and is recorded in the contemporary literature. The present-day condition of the BSCRS is a direct consequence of this reclamation, which caused the deposition of artificially deposited lagoonal deposits above the Claret Formation and subsequently allowed the development of a subaerial weathering profile.
CHAPTER THREE

GEOTECHNICAL BACKGROUND TO
THE BOTHKENNAR SOFT CLAY RESEARCH SITE

3.1 INTRODUCTION

3.2 SUBSURFACE DATA FROM THE GRANGEMOUTH AREA

3.3 THE BRITISH GEOLOGICAL SURVEY UPPER FORTH ESTUARY PROJECT

3.4 INITIAL INVESTIGATIONS AT THE BOTHKENNAR SOFT CLAY RESEARCH SITE

3.5 THE BOTHKENNAR GEOTECHNICAL CHARACTERISATION STUDY

3.6 SUMMARY
CHAPTER THREE

GEOTEchnical BACKGROUND TO
THE HOLOCENE DEPOSITS AT BOTHKENNAR

3.1 INTRODUCTION

The carse clays of the Forth estuary have been the subject of geotechnical investigations for the past fifty years. One of the earliest publications on the geotechnical properties of these deposits was that by Skempton (1948) which arose from problems experienced during commercial investigations associated with developments at the Grangemouth petrochemical complex. Following this and other early work by Skempton (1944; 1950; 1953; 1970) and Skempton and Northev (1953), the study of clay soils in their engineering geological setting was later pursued at Imperial College (University of London) and at Laval and Sherbrooke universities in Quebec, Canada, by a number of other workers (Chandler, 1972; Leroueil et al., 1979; Locat & Lefebvre, 1986; Leroueil & Vaughan, 1990). A large volume of this work was reviewed by Burland (1990) in his 1990 Rankine Lecture and more recently by Hight (unpublished) in his Rankine Lecture of 1998. The approach is very much in line with the central philosophy of this Thesis: that a knowledge of the composition, fabric and geological architecture of a clay sediment is basic to understanding its geotechnical character.

Almost two decades after Skempton’s initial work at Grangemouth, the investigations of J.B. Sissons and his group at the University of Edinburgh led to the compilation of an extensive borehole database from which they were able to generalise the subsurface geometry over the Grangemouth - Bothkennar - Airth area and propose this as an aid to site investigation (Sissons, 1971). The work followed on from the investigation of the buried Late-glacial gravels and early Holocene morphology of the Forth (described in Chapter Two), which was ongoing throughout the late 1960s, and was largely a re-statement of these results for the benefit of a geotechnical audience.

Between 1978 and 1981 the British Geological Survey (BGS) undertook a major engineering geology study of the ‘upper Forth Estuary’ (Gostelow & Browne, 1986), a part of which was a more detailed pilot study in the Bothkennar1-Carron area (Gostelow & Lambert, 1978; 1979;

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1 This work, undertaken in 1978, was a ‘calibration study’ for the methods to be used by the BGS Upper Forth study and was conducted at some distance from the site which later became the BSCR.
Chapter 3  Geotechnical Background to the Holocene Deposits at Bothkennar

Lambert & Gostelow, 1978). This programme was carried out under contract to the Department of the Environment and derived from the Scottish Development Department planning programme for industrial development across various regions of Scotland. Hence the study was broadly based (including foundation conditions, mining subsidence and groundwater) and was not focussed on the Holocene geology per se, although this was obviously central to some parts of the exercise. Since the project was undertaken to satisfy a non-specialist user base, and was intended for broad planning purposes, the results are necessarily of a generalised nature.

In 1984 the SERC, on the recommendations of an expert panel\(^2\), commissioned a search for a UK national soft clay test site, as described in Chapter One (section 1.2). At the time this was seen as the first stage in a larger strategy to establish a range of test sites on engineering soils of varying character. The contract was obtained by A.B.Hawkins, W.J.Larnach and D.F.T.Nash of the University of Bristol, who, between 1986 and 1987, co-ordinated a range of desk studies, subsurface borings, sample collection and laboratory analyses from a number of sites in the UK, including the coastal site at Bothkennar that was subsequently to become the BSCR S. On the basis of these results and its compliance with the other, non-geotechnical criteria described in Chapter One, a recommendation was made to the SERC in 1987 that the Bothkennar site should be purchased.

Following the initial investigations at Bothkennar, a new programme of highly detailed, more specialised studies was conducted in 1988-90 to characterise the site from both the geotechnical and the engineering geological standpoints. The geotechnical characterisation study was led by D.W.Hight of the Geotechnical Consulting Group partnership (in association with Imperial College, University of London) and involved eight university soil mechanics groups. The engineering geological study, led by M.A.Paul of Heriot-Watt University, was of longer duration (1988-94) than the geotechnical characterisation and was independent of it, although the early phases overlapped in time and were carried out in parallel. The present author was employed by this programme as the Research Associate responsible for the conduct of the work and the results so obtained form the basis of this Thesis.

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\(^2\) This cause was championed by the late Professor C.P.Wroth, then one of two programme co-ordinators for the SERC geotechnical engineering steering group, and was prompted by the recommendations of the 1981 SERC task force chaired by Sir Alan Muir Wood. The panel included representatives from the Institution of Civil Engineers and the Construction Industry Research and Information Association, in addition to those from the SERC itself (Hawkins et al., 1989).
3.2 SUBSURFACE DATA FROM THE GRANGEMOUTH AREA

3.2.1 Profiles of Geotechnical Properties

Skempton (1948) carried out in situ vane tests on the reclaimed flats near the then Anglo-Iranian Oil Company works (now part of British Petroleum) in the summer of 1948 in collaboration with Soil Mechanics Ltd, with the immediate object of reconciling conflicting shear strength results, obtained by conventional sampling and testing, with those from vane testing. The investigations also provided the first deep (28m) geotechnical profiles to be published from this area and showed a simple stratigraphy of two clay units separated by a gravel layer, with the lower unit resting on till (Figure 3.1). If it is assumed that the gravel layer represents the Bothkennar Gravel Formation then the two units can be reasonably correlated with the Claret and/or Grangemouth formations (upper unit) and the Loanhead Beds/Abbotsgrange Beds (lower unit). It is difficult to be certain of more than this from the published description (Skempton, 1948), although the colour of the lower unit (purple-grey) and the mention of thin sand layers might suggest the Loanhead Beds and with less confidence the lack of reference to laminations in the upper unit might indicate the Claret Formation in preference to the Grangemouth Formation.

The profiles (Figure 3.1) show the two units to be in some respects geotechnically continuous across the Bothkennar Gravel Formation. Although the sampling interval is rather large, it appears that the water content increases to a depth of about 15ft (4.5m) and then falls slowly with depth in an irregular manner throughout most of the profile. Examination shows that this irregularity largely follows fluctuations in the liquid limit. There is a noticeable fall in the liquid limit from ~50% to ~35% between depths of ~35ft and ~55ft (~10.7m – 16.8m) which may indicate some stratigraphic unit not shown on the published profile. There is also a very noticeable rise in liquid limit in the upper unit to 60% at ~15ft (4.5m) which anticipates that seen in many profiles at the BSCRS (below). In his Rankine Lecture 42 years after the profiles were first published, Burland (1990) showed that, based on these results, the void index of the lower unit can be seen to follow the standard sedimentation compression line in common with very many other clays and thus is normally consolidated. The upper unit is more complex, being overconsolidated in its upper part and normally consolidated below a depth of ~20ft (6m).

The (intact) undrained shear strength shows a related pattern in which, from a maximum value of ~800 lb ft² at a depth of ~5ft (37 kPa at 1.5m), there is a fall to a minimum of ~300 lb ft² at ~23ft (14 kPa at 7m). This closely mirrors the liquidity index profile as would be expected from the later work of Skempton and Northey (1953). Thereafter the strength rises at a constant rate.

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It should be noted that this figure shows a composite profile formed by the amalgamation of two separate boreholes.
Figure 3.1 Composite geotechnical profiles from a site [unidentified] at Grangemouth, Scotland (source: Skempton, 1948: figure 2).

Figure 3.2 Schematic overburden pressure and shear strength profiles commonly observed in a typical 'normally consolidated' Late-glacial or Holocene estuarine clay (source: Skempton, 1948: figure 4).
which is proportional to the plasticity index (Skempton 1957). The two low values (~200 lb ft\(^2\) at ~47 ft and 300 lb ft\(^2\) at ~92 ft) which do not follow the profile correspond to disturbed sections of the core. Based on these results and those from other estuarine clays, Skempton (1948) proposed a schematic shear strength profile for such clays (Figure 3.2). This suggestion remains in agreement with later results from Bothkennar and has a sound genetic basis in the isostatic uplift that has affected the Claret Formation following its deposition. The work of the present writer (Barras & Paul, 1999) would suggest, by analogy with the sequences studied at the BSCRS, that the normally consolidated zone I was formed in a subtidal environment and the overconsolidated zone II was formed under intertidal conditions. The status of this model for other estuarine clays (alluded to by Skempton (1948)) but not identified explicitly) is uncertain in view of possible differences in sea level history.

### 3.2.2 Three Dimensional Geometry in the Grangemouth – Kincardine Area

In the Grangemouth-Falkirk-Kincardine area the thickness of the deposits above the Bothkennar Gravel Formation was established by Sissons (1969; 1971) on the basis of around 2000 commercial boreholes which formed a part of a much larger database (14,000 in total). Sissons (1971) states that around 1500 were used to delimit the thickness of the carse clay ([sic] = Claret/Grangemouth formations) and around 900 to delimit the extent and depth of the buried gravel layer ([sic] = Bothkennar Gravel Formation). The results reported by Sissons (1971) are very similar to those reported in his earlier paper (Sissons, 1969), although augmented by a number of generalisations about likely constraints on foundation design. The main thrust of the paper is an exhortation to civil engineers of the value of existing data when combined with a more sophisticated approach to geomorphological interpretation; a message that has now largely been heeded.

The results of this work have led to a contoured surface for the top of the Bothkennar Gravel Formation (Figure 3.3). The data are rather sparse and so the surface profile published in 1971 is quite generalised; the depths to the Bothkennar Gravel Formation at the BSCRS are not shown on this map, although some extrapolation indicates a depth around 20m - 25m, in line with that subsequently discovered (Hawkins et al., 1989: cf. section 3.4.2 below).

### 3.3 THE BRITISH GEOLOGICAL SURVEY UPPER FORTH ESTUARY PROJECT

#### 3.3.1 Spatial Distribution of Geotechnical Units and Profiles

The ‘upper Forth Estuary’ was taken by this project to be a corridor that extended on both sides of the Forth estuary from Queensferry-Rosyth westwards to about six kilometres beyond Alloa (Figure 3.4). The BSCRS lies within the western part of this area, although much of its hinterland in the Forth valley around Stirling and beyond does not. The results were presented as a set of
Figure 3.3 Generalised configuration of the 'buried gravel layer' [Bothkennar Gravel Formation] in the Grangemouth area (source: Sissons, 1971: figure 4).

Key:
1. Buried gravel layer resting on planated Late-glacial marine deposits;
2. Till planated in association with buried gravel layer; 3. Rock planated in association with buried gravel layer; 4. Carse clay less than 12m thick; 5. Edge of carse; 6. Well-defined limit of marine planation; 7. Contours on surface of buried gravel at 1.5m intervals [elevations shown are in feet].
Figure 3.4 Location and extent of the British Geological Survey "Engineering Geology of the Upper Forth Estuary" project. The map also shows the principal surface materials. The box indicates the location of the Bothkennar calibration area (source: Gostelow & Browne, 1986; figure 15b).
eight engineering geological maps and an accompanying report (Gostelow & Browne, 1981; 1986).

The deposits in the study area were first categorised into eight geotechnical units (A to G, including subunits A(i)/A(ii)), which are defined in Table 3.1 (from Gostelow & Browne, 1986). The overall classification was based largely on lithology and so is closely related to the stratigraphical units defined by Browne et al. (1984), the correlation with which was shown in Table 2.1. The approach to the Quaternary sediments centred on the definition of an engineering stratigraphy based around nine standard geotechnical profiles (termed types 1-7, but including subdivisions 3(i)/3(ii) and 7(i)/7(ii)), whose distribution could then be plotted over the area. The definition and distribution of these geotechnical profiles is shown in Figure 3.5 and Figure 3.6 respectively.

The concept of a standard profile of value for geotechnical work in the Forth area was originally stated by Sissons (1969), who proposed what he termed profile types A, B and C, defined by their relationship to the Bothkennar Gravel Formation and the sub-carse peat. Sissons' profiles essentially lie in sequence seawards from the basin margin: type A samples only older beds landward of the subcrop of the Bothkennar Gravel Formation, type B lies seaward of the Main Postglacial Shoreline and so penetrates the Claret Formation and the buried beaches above the Bothkennar Gravel Formation, whereas type C lies seaward of the limit of the buried beaches and so passes into the Bothkennar Gravel Formation immediately beneath the Claret Formation. As shown in Figure 3.5 this is essentially the approach also taken by the BGS study: the differences lie in the greater complexity of the stratigraphic cross section used to define the profiles and thus in their greater number, which no longer follow a simple spatial sequence in their numbering. However, their areal distribution (Figure 3.6) still follows a pattern analogous to that of the surface drift deposits, since both are controlled by the same palaeogeographical model. That the methodology works in this way relies on the Late-glacial/Holocene infill of the estuary having the regular architecture described in Chapter Two. This suggests the approach is valid for sequences elsewhere whose architecture was imposed by some regular fluctuation, such as changes in relative sea-level.

3.3.2 Detailed Results from the BGS Bothkennar Calibration Area

The main Upper Forth Estuary project was preceded by a special calibration study in what was then termed the ‘Bothkennar area’. This is shown in Figure 3.4. It is important to realise that this is not the BSCR and is in fact an area of some 20 km² centred about 3km south of the EPSRC site. It also extends some distance inland. It will therefore be termed the ‘Bothkennar calibration area’ here, to avoid possible confusion with the BSCR. The BGS study involved the sinking of 20 boreholes and 17 cone penetration tests, plus subsequent sedimentological and geotechnical
Table 3.1 Definitions of geotechnical units employed in the BGS Upper Forth Project
(Source: Hawkins et al., 1989).

<table>
<thead>
<tr>
<th>Geotechnical Units/Groups</th>
<th>Soil Description of Groups</th>
<th>Geological Stratigraphic Classification</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ai</td>
<td>Top 0.5 m of clayey topsoil or brick/cinder made ground overlying firm or stiff dark yellowish brown clayey SILT or silty CLAY with moderate brown mottling. Shell bands and pods are common.</td>
<td>Desiccated layer</td>
</tr>
<tr>
<td>Aii</td>
<td>Very soft-firm dark yellowish brown to olive grey silty CLAY. In north of area this becomes closely laminated with fine/medium SAND.</td>
<td>Weathered Carse Clay</td>
</tr>
<tr>
<td>B</td>
<td>Very soft-soft olive black silty CLAY with laminations of silty fine SAND, silty animal burrows, small shell fragments and occasional carbonaceous pods are also present.</td>
<td>POST GLACIAL</td>
</tr>
<tr>
<td></td>
<td>Soft olive grey clayey SILT with laminations or lenses of fine SAND or medium/fine SAND with laminations of silty CLAY or fine/coarse GRAVEL.</td>
<td>Unweathered Carse Clay</td>
</tr>
<tr>
<td>C</td>
<td>Compact sandy coarse GRAVEL with COBBLES and BOULDERS sometimes with a clay matrix. Shells are present in south east of area.</td>
<td>LATE GLACIAL</td>
</tr>
<tr>
<td>D</td>
<td>Soft-firm greyish brown or dusky yellowish brown silty CLAY closely laminated with silty fine SAND, to loose to dense medium SAND with clayey laminations.</td>
<td>Gravel</td>
</tr>
<tr>
<td>E</td>
<td>Soft greyish brown silty CLAY with occasional laminations or dustings of silt or silty fine SAND. Sand laminations become more numerous near the base of this layer.</td>
<td>Laminated Clay</td>
</tr>
<tr>
<td>F</td>
<td>Firm/stiff to hard dusky yellowish brown silty sandy gravelly CLAY (BOULDER CLAY). This often passes down into compact SAND especially in the southern part of the area.</td>
<td>Glacial Deposits</td>
</tr>
<tr>
<td>G</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
analyses (Lambert & Gostelow, 1978; Gostelow & Lambert, 1978; 1979). These borings were made by the shell and auger technique and the percentage core recovery was poor from many of the holes. It is likely that the samples were therefore badly disturbed in some cases and thus that the geotechnical results on 'intact' samples must be treated with caution.

Figure 3.7 shows the geotechnical profile (Type 6) found in the Bothkennar calibration area. If it is assumed that the gravel layer (unit D) represents the Bothkennar Gravel Formation then it appears that in this area the Claret Formation (units A,B,C) is ~5m in thickness and that most of the borehole lies in the Late-glacial Kinneil Kerse and Loanhead beds (units E and F) to a depth of ~23m. (The Abbotsgrange Beds (Browne et al., 1984) are not represented here since they only occur seawards of the Kinneil Kerse Beds). The relatively thin expression of the Claret Formation, as compared with the BSCR5 (Hawkins et al., 1989; Nash et al., 1992a), is due to the rising surface of the gravel layer as it approaches the buried Late-glacial shoreline, as shown in Figures 3.3 and 2.8.

Table 3.2 shows the geotechnical characteristics of units A to F as they appear in the calibration area and the stratigraphical units with which they correlate. The data show that there is considerable variation both within and between the units in terms of their plasticity and their grading. Within the Claret Formation (units A-C) the material becomes distinctly sandier towards the surface. The minimum sand content is seen in unit C and the proportion increases downwards toward the till basement. The plastic limit is similar in all the units (a feature also seen in Skempton's (1948) borings) whereas the liquid limit and plasticity index are quite variable between the units, being greatest in the upper, unweathered Claret Formation (unit B) and in the Loanhead beds (unit F). The reasons, however, appear to be quite different. The Loanhead beds have a clay fraction activity in the range 0.41-0.71 (mean: 0.50) and a clay-sized percentage 12% - 64%: their high plasticity appears to be the consequence of a relatively inactive mineral suite but a sometimes high clay content. The upper, unweathered Claret Formation has a clay-sized percentage of 14% - 47% and a clay fraction activity up to 2.26 (mean: 0.83). On mineralogical grounds an activity value of ~0.4, similar to that from the other units, would be expected: in this case the plasticity seems to be the consequence of an unusually high activity in the clay-sized fraction.

The mineralogy of the clay-sized fraction is similar in units B, E and unit F (Gostelow & Lambert, 1978: table 3; reproduced here in modified form as Table 3.3). The principal active minerals are said to be mica (probably illite), chlorite and kaolinite which together comprise 50% - 70% + of the clay-size fraction in both units. Illite is generally around 40%, the other two 10% to 20% each. Using the mean values for the units and taking illite to have an activity of 0.9, kaolinite 0.4 and chlorite 0.2 (based on Skempton, 1953) leads to estimates of activity of 0.42 for unit B and 0.41 for
Table 3.2  Characteristics of geotechnical units A-F in the BGS Bothkennar Study Area¹.

<table>
<thead>
<tr>
<th>Geotechnical unit</th>
<th>Stratigraphical Unit ²</th>
<th>Liquid limit (%)</th>
<th>Plastic limit (%)</th>
<th>Plasticity index (%)</th>
<th>Activity</th>
<th>Organic (%)</th>
<th>Clay (%)</th>
<th>Silt (%)</th>
<th>Sand (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Weathered Claret Formation</td>
<td>41.9 (36.1 - 48.1)</td>
<td>22.3 (20.9 - 23.5)</td>
<td>19.5 (13.5 - 27.1)</td>
<td>0.89 (0.61 - 1.06)</td>
<td>2.04 (1.84 - 2.23)</td>
<td>20 (14 - 27)</td>
<td>61 (13 - 73)</td>
<td>19 (5 - 51)</td>
</tr>
<tr>
<td>B</td>
<td>Unweathered Claret Formation (silty clay)</td>
<td>45.9 (28.5 - 70.5)</td>
<td>23.2 (17.2 - 33.2)</td>
<td>22.7 (3.4 - 42.9)</td>
<td>0.83 (0.23 - 2.26)</td>
<td>0.97 (0.13 - 2.52)</td>
<td>28 (14 - 47)</td>
<td>64 (26 - 78)</td>
<td>7 (0 - 59)</td>
</tr>
<tr>
<td>C</td>
<td>Unweathered Claret Formation (clayey silt)</td>
<td>32.5 (17.9 - 45.2)</td>
<td>19.6 (13.2 - 26.9)</td>
<td>14.1 (4.7 - 29.8)</td>
<td>0.74 (0.25 - 1.14)</td>
<td>0.82 (0.13 - 2.23)</td>
<td>25 (13 - 32)</td>
<td>71 (55 - 81)</td>
<td>4 (0 - 10)</td>
</tr>
<tr>
<td>D</td>
<td>Bothkennar Gravel Formation</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
<td>not given</td>
<td>not given</td>
<td>not given</td>
<td>not given</td>
</tr>
<tr>
<td>E</td>
<td>Kinnell Kerse Beds</td>
<td>29.5 (24.2 - 35.3)</td>
<td>18.7 (13.2 - 22.5)</td>
<td>10.7 (6.0 - 16.6)</td>
<td>0.34 (0.22 - 0.47)</td>
<td>0.52 (0.10 - 0.73)</td>
<td>32 (17 - 45)</td>
<td>61 (50 - 67)</td>
<td>7 (0 - 33)</td>
</tr>
<tr>
<td>F</td>
<td>Loanhead Beds</td>
<td>48 (36.7 - 61.1)</td>
<td>22.7 (17.7 - 37.6)</td>
<td>26.3 (18.9 - 35.3)</td>
<td>0.50 (0.41 - 0.71)</td>
<td>1.17 (0.10 - 2.42)</td>
<td>49 (12 - 64)</td>
<td>45 (23 - 64)</td>
<td>10 (0 - 13)</td>
</tr>
</tbody>
</table>

¹Values are reported as mean and (range). Data from Gostelow & Lambert (1978): table 2.
Table 3.3  XRD Analysis of samples from the BGS Bothkennar Calibration Area.

<table>
<thead>
<tr>
<th>BH/Depth (m)</th>
<th>Geotechnical Unit</th>
<th>Stratigraphical unit</th>
<th>Quartz (%)</th>
<th>Calcite (%)</th>
<th>Dolomite (%)</th>
<th>Mica (%)</th>
<th>Chlorite (%)</th>
<th>Kaolinite (%)</th>
<th>Other minerals detected</th>
</tr>
</thead>
<tbody>
<tr>
<td>13/ 3.5</td>
<td>B</td>
<td>Claret Beds</td>
<td>31</td>
<td>5</td>
<td>4</td>
<td>38</td>
<td>11</td>
<td>11</td>
<td>feldspar, pyrite, amphibole</td>
</tr>
<tr>
<td>24/ 4.0</td>
<td>B</td>
<td>Claret Beds</td>
<td>21</td>
<td>6</td>
<td>4</td>
<td>41</td>
<td>14</td>
<td>14</td>
<td>feldspar, pyrite, amphibole</td>
</tr>
<tr>
<td>13/ 6.3</td>
<td>B</td>
<td>Claret Beds</td>
<td>30</td>
<td>8</td>
<td>4</td>
<td>37</td>
<td>13</td>
<td>8</td>
<td>feldspar, pyrite, amphibole</td>
</tr>
<tr>
<td>8/ 9.8</td>
<td>E/F</td>
<td>Kinneil Kerse Beds/ Loanhead Beds</td>
<td>35</td>
<td>10</td>
<td>2</td>
<td>26</td>
<td>16</td>
<td>11</td>
<td>feldspar, pyrite</td>
</tr>
<tr>
<td>19/ 14.8</td>
<td>E</td>
<td>Kinneil Kerse Beds</td>
<td>24</td>
<td>5</td>
<td>2</td>
<td>40</td>
<td>17</td>
<td>12</td>
<td>feldspar, pyrite</td>
</tr>
<tr>
<td>19/ 18.8</td>
<td>F</td>
<td>Loanhead Beds</td>
<td>19</td>
<td>2</td>
<td>&lt;1</td>
<td>39</td>
<td>20</td>
<td>20</td>
<td>feldspar, pyrite</td>
</tr>
</tbody>
</table>

Source: Gostelow & Lambert (1978).
units E/F. This value for E/F agrees well with the mean values given in Table 3.2 (0.34 [E], 0.50 [F]) whereas the reported mean value of 0.83 for unit B is clearly in excess of that calculated from the mineralogy. A similar result has been obtained at the BSCRS and here it has been shown (Paul & Barras, 1999) that it is due in part to the effect of the amorphous organic component that arose from the estuarine fauna. This point is discussed in detail in Chapter Seven (section 7.3). Gostelow and Lambert (1978) dismiss the sediments in the Bothkennar calibration area as ‘virtually inorganic, in most cases less than 1% was recorded’ (section 4.4.4, p.7) which is not only misleading about their own data for unit B (max value 2.52%; mean 0.97%) but also overlooks the importance of even small quantities of the appropriate organic materials in plasticising an otherwise inorganic sediment. It is very likely that this is one cause of the anomalous activity in the Claret Formation at the calibration site in the same way that it is in the Claret Formation at the BSCRS.

Figure 3.7 shows the depth profiles of a number of geotechnical properties from borehole 25 (Grid square NS 9182) in the calibration area near the River Carron. They are based on rather few data points (typically one every two metres) and so outlying values have a disproportionate effect and must be treated with caution. In unit E (other than in its uppermost metre) and in unit F they show the pattern expected in a normally consolidated clay: liquidity index falls consistently and undrained shear strength rises at a uniform gradient (S_u/P_d) of 0.23. The sensitivity is around 2-4 over much of the profile. The unexpectedly low value (<0.4, w ~20%) of liquidity index in the uppermost metre of unit E, which corresponds with a high value of undrained shear strength (40 kPa - 80 kPa) has been suggested by Gostelow and Lambert (1978) to be a former desiccated zone indicative of subaerial exposure and thus a lowered sea-level. This has implications for the Late-glacial sea-level (section 2.4.2) and agrees with the view that the Bothkennar Gravel Formation represents a former beach. This has been challenged by Peacock (1998) who argues that the BGF was ice-rafted and that the evidence of this one data point is uncertain. This is undoubtedly so: there are other reasons why the water content might be lowered in this sample, notably its evident association with a coarser, laminated zone within unit E, as indicated by both the cone profile and the large spread of shear strength values at this point. There is simply insufficient evidence from these data to confirm or deny the existence of a ‘buried crust’, although the suggestion itself is intriguing.

In units A-C, which represent the Claret Formation, the water content appears to show an initial rise over a depth of around 3m to a local maximum of ~40% (I_c ~1.0), followed by a reduction below this depth. This pattern is based on only two data points, from which it is difficult to draw many conclusions. However, examination of the detailed boreholes logs from the calibration area (Lambert & Gostelow, 1978) suggests that the maximum occurs within or just below the weathered zone of the Claret Formation and not at the local base of the desiccated crust. If this is
Figure 3.7 Illustrative geotechnical property profiles (Type Profile 6) from borehole 25 [grid square NS 9182] within the Bothkennar calibration area. (source: Gostelow & Browne, 1986; figure 24).
the case in borehole 25 (for which no detailed log appears to be published) it indicates that the lowered water content in the uppermost part of the profile may follow the general pattern proposed by Skempton (1948) and illustrated by Figure 3.2. This pattern is certainly seen in the boreholes at the BSCRS (below) and appears to be a consequence of intertidal exposure during the closing stages of deposition (as discussed in Chapter Eight).

Figure 3.7 also shows the chlorninity profile of the pore water. This reveals that above about 5m depth the pore water is virtually chloride-free and that there is an increase below this depth to between 4% to 6% over much of the profile, with a maximum value of about 6.5% (=11.7% salinity) at around 12m depth. This agrees with the findings from the BSCRS (Paul et al., 1992a), that showed that the pore water had a peak salinity of around 19% at 9.5m depth and reduced steadily from ~11% at ~8m to ~4% at ~2m. This suggests that, firstly, there is a well-defined freshwater layer above the somewhat saline body of the sediment and, secondly, that at the BSCRS the porewater below about 2m is more saline than that in borehole 25. This is explicable if it is noted that the BSCRS lies immediately adjacent to the coast, whereas the grid square that contains borehole 25 lies about 2km inland (although in part close to the tidal River Carron). This may indicate saline penetration into the Claret Formation which decreases in strength away from the coastal area.

3.4 INITIAL INVESTIGATIONS AT THE BOTHKENNAR SOFT CLAY RESEARCH SITE

3.4.1 Objectives of the Initial Investigations

Prior to the purchase of the site at Bothkennar, a number of geotechnical investigations were undertaken during 1986-87 in order to investigate its suitability as a permanent soft clay test bed site. These comprised an initial site selection phase to determine the overall suitability of the site and, when Bothkennar had become the leading choice, a second phase of more extensive investigations to obtain a full geotechnical profile of the soft clay deposits and to establish possible variability across the site. The results were in due course reported in a series of publications and various internal documents (Nash & Lloyd, 1988a; 1988b; 1989; Lloyd, 1989; Hawkins et al., 1989; 1991; Nash et al., 1992a).

These investigations had three principal objectives:

- To describe the sediments and to establish their thickness and continuity;
- To investigate possible lateral variation and the presence of sand or peat;

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4 Salinity = 1.80655 x chlorinity (Riley & Chester, 1971).
To undertake a preliminary geotechnical characterisation.

The site plan for the various subsurface investigations is shown in Figure 3.8. Over the two phases, the investigations comprised a total of six conventional (piston core) boreholes, six Delft stocking boreholes and twenty-six cone/piezocone profiles, together with in situ shear vane tests and earth pressure measurements (Table 3.4). The piston coring and continuous sampling were carried out by Soil Mechanics Ltd; earth pressure measurements and vane tests by the Building Research Establishment; the cone penetration tests by Messrs Fugro and the Delft stocking sampling by Delft Geotechnics UK.

3.4.2 Description of the Sediments

The boreholes showed that the Claret Formation was present across the entire site and was everywhere underlain by the Bothkennar Gravel Formation at depths between -13m and -19m OD (Figure 3.8) (16m - 22m bgl). It was also found that in the south-eastern corner of the site the Claret Formation had been locally eroded and replaced by finely laminated clayey silts to a depth of about 7m below ground level. This structure appeared to be the flank of a major channel, the main body of which lay to the east of the site boundary.

The desiccated crust was correctly stated to be around 1m - 2m in thickness and so to be thinner than is normal in this area. Despite the relatively sheltered setting this was attributed to erosion by wave action (Hawkins et al., 1989) and a more likely explanation, that the site is situated on reclaimed land, was for some reason not identified, despite the reclaimed area being clearly shown on map sheet 5 (Drift geology) from the 1981 engineering geology study (Gostelow & Browne, 1981), although map sheet 6 (Engineering classification of surface sediments) from the same study shows (erroneously) that the desiccated crust is absent from the reclaimed areas and this has obviously caused confusion.

The appearance and macrofabric of the sediments has been described in detailed logs of boreholes BH1 and D1-D6 (Lloyd, 1989; Hawkins et al., 1989;1991), supplemented by a photographic record (Nash & Lloyd, 1989). Unfortunately the cheese-wire/hand splitting technique used by the authors to prepare the sediments for description and photography was not entirely effective in these soft, cohesive sediments and resulted in some damage and loss of detail. In addition, the photographic log was reproduced at a small scale (about 15% of natural size) with further loss of detail and no X-ray photographs were taken. As a result, some fabric features subsequently known to occur in these sediments were either overlooked or under-reported. In the later work by the present writer, these deficiencies were largely remedied by the use of the osmotic knife technique for surface preparation (cf. section 4.3.1.2) and photographic reproduction at full natural size.
Figure 3.8 Plan of the Bothkennar Soft Clay Research Site showing the locations of the boreholes and in situ tests conducted during the initial investigation phase (1986-87) and the geotechnical characterisation study (1989). Also shown is the approximate elevation of the surface of the Bothkennar Gravel Formation below the site. (Source: Hight et al., 1992a: figure 1 with additions from Nash et al., 1992a).
Table 3.4 Boreholes and in situ tests made at the BSCRS during the initial site investigation (Source: Hight et al., 1992a; table 1).

<table>
<thead>
<tr>
<th>Phase</th>
<th>Date</th>
<th>Notation</th>
<th>Borehole/in situ test details</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1986</td>
<td>F1–F6</td>
<td>6 cone penetration tests</td>
</tr>
<tr>
<td></td>
<td></td>
<td>F7–F8</td>
<td>2 piezocone tests</td>
</tr>
<tr>
<td></td>
<td></td>
<td>BH1</td>
<td>Profile of overlapping piston samples</td>
</tr>
<tr>
<td></td>
<td></td>
<td>BH1–2</td>
<td>Profile of overlapping piston samples</td>
</tr>
<tr>
<td>2</td>
<td>1986–87</td>
<td>V1</td>
<td>In situ vane profile</td>
</tr>
<tr>
<td></td>
<td></td>
<td>PR1</td>
<td>Pressuremeter profile</td>
</tr>
<tr>
<td></td>
<td></td>
<td>P1–P18</td>
<td>18 piezocone tests</td>
</tr>
<tr>
<td></td>
<td></td>
<td>D1–D6</td>
<td>6 Delft continuous sampling holes</td>
</tr>
<tr>
<td></td>
<td></td>
<td>BH2/1</td>
<td>Additional piston sampling hole</td>
</tr>
<tr>
<td></td>
<td></td>
<td>BH2/3</td>
<td>Additional piston sampling hole</td>
</tr>
<tr>
<td></td>
<td></td>
<td>BH2/3/2</td>
<td>Additional piston sampling hole</td>
</tr>
<tr>
<td></td>
<td></td>
<td>BH2/1 + DBM</td>
<td>Deep borehole to 37 m</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Piezometers</td>
</tr>
</tbody>
</table>

Table 3.5 Universities involved in the Bothkennar Characterisation Study (Source: Hight et al., 1992a; table 2).

<table>
<thead>
<tr>
<th>Investigation</th>
<th>Apparatus/test</th>
<th>Institution</th>
</tr>
</thead>
<tbody>
<tr>
<td>Permeability</td>
<td>Variable head oedometer tests</td>
<td>Laval University</td>
</tr>
<tr>
<td></td>
<td>Constant head triaxial tests</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Radial flow cell tests</td>
<td></td>
</tr>
<tr>
<td></td>
<td>75 mm and 150 mm Rowe cell tests with vertical and radial flow, using flow</td>
<td>Glasgow University</td>
</tr>
<tr>
<td></td>
<td>pump techniques</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Constant vertical flow tests in oedometer cells</td>
<td>Oxford University</td>
</tr>
<tr>
<td>One-dimensional, compressibility,</td>
<td>Incremental load (IL) oedometer tests</td>
<td>Bristol University</td>
</tr>
<tr>
<td>expansibility and yielding</td>
<td>Restricted flow (RF) oedometer tests</td>
<td>Oxford University</td>
</tr>
<tr>
<td></td>
<td>Continuous load (CL) oedometer tests</td>
<td>Bristol Polytechnic</td>
</tr>
<tr>
<td>Undrained strength</td>
<td>Triaxial compression and extension tests on 38 mm dia. × 76 mm specimens at</td>
<td>City University</td>
</tr>
<tr>
<td></td>
<td>5% strain/day</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Simple shear tests in a Geonor device on 80 mm dia. × 20 mm specimens at 2.5%</td>
<td></td>
</tr>
<tr>
<td></td>
<td>strain/hour</td>
<td></td>
</tr>
<tr>
<td>Yielding characteristics</td>
<td>Triaxial tests on 38 mm dia. × 76 mm specimens, with local measurement of</td>
<td>Imperial College</td>
</tr>
<tr>
<td></td>
<td>axial and radial displacement and mid-height pore pressure</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(Specimens trimmed from Laval and Sherbrooke samples from similar depths)</td>
<td></td>
</tr>
<tr>
<td>Destructuring of clay and effects of</td>
<td>Strain path tests on 100 mm dia. × 200 mm specimens with local measurement</td>
<td>Surrey University</td>
</tr>
<tr>
<td>sampling</td>
<td>of axial and radial displacement and mid-height pore pressure</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(Specimens subjected to idealized tube sampling strains before subsequent</td>
<td></td>
</tr>
<tr>
<td></td>
<td>shear)</td>
<td></td>
</tr>
</tbody>
</table>
The macrofabric descriptions obtained from the initial programme are similar to those presented in Chapter Five, although less detailed, and so will only be discussed briefly here. The present writer has subsequently described a more formal facies model for these sediments (Barras & Paul, 1999; and Chapters Five and Eight following) and these facies names are given below in parenthesis whenever a direct comparison has been made with the earlier descriptions.

The Bristol group showed (Lloyd, 1989; Hawkins et al., 1991) that the sediments were divisible into three broad categories and three colour zones. In the first category are sediments with some visible bedding, sometimes with silty partings or laminae (bedded facies). In the second are mottled sediments in which any bedding is not easily visible (mottled facies). The third category comprises finely laminated sediments with very numerous silt partings (laminated facies). The sediments were found to vary in colour with depth: from brown at surface in the desiccated crust (probably unit A(i) of Gostelow & Browne, 1986) through light grey immediately below the water table to a depth of 3m to 4m (probably unit A(ii)) and then dark grey to black, (probably unit B) although this colour lightened to light grey after about 30 minutes exposure to the atmosphere.

3.4.3 Lateral Variability

The lateral variation was mainly investigated by inter-comparison of the cone penetration profiles from 26 positions across the site. These showed that, other than in the laminated area, the lateral variability is very limited and thus that any discussion must centre on what are very minor variations by comparison with those found in other estuarine clays (Figure 3.9) (cf. Hawkins et al., 1989: figure 33).

Below about 2m depth (the base of the crust and the shell bed), the raw cone resistance profiles (Figure 3.10) are confined to a generally narrow envelope. There is no clear evidence of any systematic lateral change between the profiles. Derived parameters from the piezocone tests have also been described by Nash et al. (1992a) and are shown in Figure 3.11. These confirm the general results obtained from the raw data, although there is some indication that the friction ratio and pore pressure ratio are more sensitive to minor lithological variations than are the other parameters. Of particular interest is the normalised net cone resistance $q_c/\sigma'_{v,0}$ which shows that below about 7m the net resistance is directly proportional to vertical in situ stress: a characteristic of a normally consolidated soil. Above this depth the normalised resistance is greater than in a normally consolidated soil and increases upwards until the shell bed is reached. This is indicative of light overconsolidation and thus agrees with other data, such as the profile of water content, that shows the sediment to be increasingly overconsolidated above about 7m depth.
Figure 3.9 Comparative envelopes of Delft piezocone profiles from Brean, Newport, Swale and Bothkennar. (Source: Hawkins et al., 1989; figure 28).

Figure 3.10 Envelope of Delft piezocone profiles [total cone resistance] from the Bothkennar Soft Clay Research Site (P series: locations shown on Figure 3.8). (source: Hawkins et al., 1989; figure 29).
Figure 3.11 Envelopes of piezocone profiles from the cone tests illustrated in Figure 3.10
(a) normalised net resistance; (b) normalised pore pressure change; (c) friction ratio.
(source: Nash et al., 1992a: figure 18).

Figure 3.12 Comparative profiles from Delft stockpiling cores D1 to D6 (locations shown on
Figure 3.9) to illustrate the variability of the extent, style and depth of mottling
across the Bothkennar Soft Clay Research Site
(Source: Hawkins et al., 1991; figure 3).
The nature and extent of the visible mottling (Hawkins et al., 1989; 1991) also gives some indication of lateral variability, although again no clear picture emerges. Inter-comparison of the uppermost 14m of the six Delft stocking profiles (Figure 3.12) (hereafter referred to as D1, D2 and so on), shows that in D1 (in the south of the site) the upper part of the profile is the more mottled, whereas in D3-D6 (the middle and north of the site) the converse is the case. Examination of the detailed logs (Nash & Lloyd, 1989) indicates that in the latter cores the uppermost 3m-5m below the crust is largely bedded and contains numerous coarser laminae, whereas in D1 (and also in borehole HW3, to be described later in this Thesis) the uppermost sediments are somewhat finer and thus mottling has developed. It seems reasonable to presume that these variations reflect the normal, medium scale, variability of processes on tidal flats, although at the scale of the site any consistent lateral pattern cannot be distinguished.

3.4.4 Preliminary geotechnical characterisation

Hawkins et al. (1989) and Nash et al. (1992a) have described profiles of water content and bulk density, Atterberg limits, particle size distribution, undrained shear strength, vertical effective stress and yield stress. These will only be considered in outline, since the geotechnical character of the soil has been established in much greater detail by the main characterisation study (section 3.5 below) and by the work of the present writer (cf. Chapter Seven).

3.4.4.1 Particle size distribution and composition

Outline particle size analyses have been reported by Hawkins et al. (1989) for the Delft boreholes (Figure 3.13). These analyses were carried out using the hydrometer method (BS 1377, 1975; Lloyd, 1989). It should be noted that there was probably some underestimation of the clay content using this method due to incomplete dispersion of clay aggregates (Hawkins et al., 1989). This implies that there was a corresponding overestimation of the silt fraction. This would probably affect the entire silt range, since aggregates spanning sizes throughout this range have been seen under the scanning electron microscope (cf. Chapter Five).

3.4.4.2 Plasticity and water content

Atterberg limit tests (Nash et al., 1992a) have shown the soil to be of medium plasticity (I_p ~25% to ~55%, exceptionally ~20% and 60%). The liquid limit was found to be particularly sensitive to the method of test: air drying in accordance with BS1377 (1975) reduced these values by up to

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5 It should be noted that the Bristol investigations did not recognise mottling below about 14m in any of the Delft stocking cores (D.F.T.Nash, pers. comm. Figure 3.12). The present writer has identified mottling below this depth (Chapter Five).
Figure 3.13  Individual particle size and organic content profiles for Delft series boreholes (a) D1; (b) D2; (c) D5; (d) D6. (Source: Hawkins et al., 1989; figures 22 and 23.)
20\% in some cases. The plasticity chart (Figure 3.14(a)) shows that the material generally plots above the A-line and can be classified as an inorganic clay of high plasticity (Nash et al., 1992a). Figure 3.14(b) shows that the activity averages 1.34. Knowledge of the mineralogy gained from this present work (Paul et al., 1992a) indicates that this figure is somewhat excessive, probably as a result of the presence of organic materials (cf. Chapter Seven). The organic content (measured by loss on ignition) was found to be around 5\%-10\% at all depths throughout the boreholes (Figure 3.13). It may in part also be a reflection of the low percentage values for the clay size fraction obtained by the Bristol University investigation by the use of the hydrometer method without ultrasonic agitation.

Figure 3.15(a) shows the water content profiles for boreholes D1 to D6 plotted together with the average Atterberg limits for the whole site (Nash et al., 1992a). The profiles for the six boreholes generally follow a similar pattern, with fluctuations, and (except for D2, D3 and D5) are contained within the average plasticity index envelope. In all of the boreholes, below the water table, there is a general increase in water content to between about 50\% and 70\% down to a depth of about 7.0m, below which depth the values decrease and tail off to between 30\% and 50\% at the bottom of each borehole at the Bothkennar Gravel layer.

Figure 3.15(b) shows that in each of boreholes D1 to D6, the liquidity index is greater than one in the upper part of the profile. This is not readily apparent in Figure 3.15(a) owing to the scatter of results and it is therefore more convenient to consider the values for each borehole individually. Figure 3.16 (a) to (e) shows the water content and the Atterberg limits at each of the boreholes D1, D2, D4, D5 and D6 (detailed results for D3 are not available in the literature). In the figure they are presented in an order that facilitates comparisons in the north-south and east-west directions, although at the scale of sampling no clear spatial pattern emerges.

3.4.4.3 Undrained shear strength and sensitivity

The undrained shear strength was measured during the initial site investigation at the site (Hawkins et al., 1989; Nash et al., 1992a). These were made in situ by both direct measurement using field vanes, and indirectly by using the results of piezocone and Marchetti dilatometer tests. The laboratory measurement of shear strength was carried out using undrained triaxial tests. For the purpose of this study only the results of the in situ field vane measurements will be reported.

A number of sets of the test were carried out close to borehole D1 and the results of these are shown in Figure 3.17 although subsequent work indicates that these values could be low (Nash et al., 1992a). Below the desiccated crust there is a general increase in shear strength with depth: values steadily increased from about 15kPa immediately below the shell-bed layer to 55kPa at
Figure 3.14  (a) Plasticity chart for the samples collected during the initial investigations at the Bothkennar Soft Clay Research Site. The samples were tested from the natural water content without prior drying. (b) Activity chart for the samples shown in Figure 3.14(a). (Source: Nash et al., 1992a; figure 7.)
Figure 3.15 Composite profiles of (a) water content and (b) liquidity index for Bothkennar boreholes D1 to D6. (Source: Hawkins et al., 1992a; figure 6).
Figure 3.16 (a)-(e) Individual water content profiles for Delft series boreholes (source: Hawkins et al., 1989).
the base of the borehole. In the surface crust there is a decrease in the vane shear strength from \( \sim 30 \text{kPa} \) at the ground surface to \( \sim 22 \text{kPa} \) just above the shell-bed layer.

Field vane measurements suggest the average sensitivity to be \( \sim 5 \) (Figure 3.17). Hight et al., (1992a) reported somewhat higher sensitivities obtained by using the fall-cone method: these authors obtained values between 5 to 8 between depths of 2.7m and 7.0m and between 7 to 14 from a depth of 7.0m to the base of the borehole. The reasons for these rather higher values are unclear but probably relate to the smaller size of the fall cone test, which is thus affected by the local structure, itself variable at the small scale.

3.4.4.4 Vertical yield stress and yield stress ratio

The compression behaviour of the Bothkennar sediments was obtained by means of conventional, incremental load oedometer tests and the yield stress (overconsolidation pressure) was estimated using the graphical construction technique proposed by Casagrande (1936). The vertical profile of yield stress so obtained is shown in Figure 3.18(a). The corresponding profile of yield stress ratio (overconsolidation ratio) is shown in Figure 3.18(b). The results indicate that below about 5m the yield stress ratio (YSR) is relatively constant and exceeds 1.5 over most of this depth. Later work (Nash et al., 1992a; this Thesis, Chapter Seven) has shown that above 5m the YSR rises to values of 2-3. Below about 5m this ‘overconsolidation’ is a reflection of the level of structure in the soil and probably does not signify any history of previous loading (see section 3.5.4 below). Above this depth the overconsolidation is believed to be a real effect that is due to exposure in the intertidal zone, as discussed in Chapters Seven and Eight.

3.5 THE BOTHKENNAR GEOTECHNICAL CHARACTERISATION STUDY

3.5.1 Aims of the Geotechnical Characterisation Study

The geotechnical characterisation study had several aims and at the time was seen as more than just the collection of geotechnical data, albeit of unusually high quality, from the ‘Bothkennar clay’ (as it was by then invariably called \(^6\) ). The principal of these aims were:

- To employ and evaluate state-of-the-art methods for the collection of samples in soft clay;
- To establish a framework for the mechanical behaviour of the Bothkennar clay;

\(^6\) The term “Bothkennar clay” is the name used by the geotechnical community for the material that comprises the soft clay sediments of the Clare Formation at the BSCR. It is convenient to use it here since it has been widely used in the publications that have arisen from the various characterisation studies. However, it is not a geological name and has no stratigraphical significance, nor is it a general synonym for ‘clay’, itself an informal term.
Figure 3.17 Profile of undrained shear strength and sensitivity measured adjacent to borehole D1 at the Bothkennar Soft Clay Research Site (source: Hawkins et al., 1989: figure 31).

Figure 3.18 (a) Profile of yield stress and (b) yield stress ratio measured at the characterisation study area (Figure 3.8) at the Bothkennar Soft Clay Research Site (source: Nash et al., 1992a: figure 8).
• To investigate the extent of geotechnical structure in the Bothkennar clay;

• To provide additional stratigraphical information based on profiles of geotechnical parameters.

The work was undertaken by a consortium of universities and made use of a range of geotechnical techniques (Table 3.5). It was also envisaged as an opportunity to evaluate state-of-the-art sampling and testing procedures, some new in the UK, and to revise existing models of soft clay behaviour in both the intact and remoulded states. In this respect it provided a test of many current concepts, using data of much higher quality than had been available previously. The results of this large body of work were published in 1992 in the Eighth Géotechnique Symposium in Print (Géotechnique 42 (2)), a key paper within which was the overview of the results by D.W.Hight and his colleagues (Hight et al., 1992a).

It is beyond the scope of this Thesis to review all the results and so attention is concentrated on those which are thought to have the greatest relevance to the engineering geological investigations on which the Thesis is based since, as mentioned earlier, these were ongoing in parallel with the characterisation study and so were to some extent shaped by its findings.

3.5.2 Sample Quality

The results of comparative tests on samples collected by different techniques (Hight et al., 1992a; 1992b) showed clearly that the highest quality, undisturbed samples were obtained by the use of the Sherbrooke sampler (Lefebvre & Poulin, 1979) which carves a block sample from the base of a borehole. By contrast, the conventional piston sampler (in common with other tube samplers) imposes an estimated centre-line strain cycle of around 1% to 2% when using 100mm diameter tubes of 2mm wall thickness (Baligh et al., 1987). This was found to be sufficient to take the Bothkennar clay beyond its peak strength (reached at around 0.5% strain: Hight et al., 1992a; 1992b) and so results in some irreversible loss of structure. The effect appears greatest in samples from the mottled facies.

3.5.3 The Framework for Geotechnical Behaviour

The framework was established by triaxial tests on remoulded samples, the results of which acted as a baseline for comparison with results on intact samples; any differences were attributed
to the level of structure\(^7\) within the soil. Work by Allman and Atkinson (1992) showed that the frictional characteristics of the soil were incompatible with its natural plasticity as measured by the index tests and that many of the standard correlations involving these parameters were therefore invalid for the Bothkennar clay. This was an important conclusion. As discussed later in this Thesis the plasticity in the natural state is a consequence of the pore water salinity and organic component of the soil: complementary work by this writer (Paul et al., 1992a; Paul & Barras, 1999) has shown that the plasticity index of the pure mineral soil is more commensurate with its frictional character and that on using this value, the normal correlations are obeyed. Importantly also, in the remoulded state the soil initially showed no evidence of apparent overconsolidation but, after about 200 hours at constant stress, some increased resistance was seen in subsequent compression tests. This indicated that some form of geotechnical structure was developed at constant stress by a process that can be loosely described as ageing (Bjerrum, 1967; Burland, 1990).

3.5.4 Evidence for a Structured Clay

The majority of the characterisation study was devoted to tests intended to establish the level of structure in the soil (illustrated schematically in Figure 3.19), based on comparisons of the behaviour in the intact and remoulded states. There were four main lines of evidence from the characterisation study that lead to the conclusion that the Bothkennar clay is structured:

1. At any given depth (i.e. stress level) the range in water content values was found to be at least 5% and could be as great as 25% in some cases (Hight et al., 1992a). Comparable variability was seen in the normalised parameters of liquidity index and void index. This indicates that at a given level of in situ stress the soil can exist at a range of void ratios which is commensurate with its being a structured material.

2. The void index varies systematically down the profile and reached a maximum with respect to the ICL at about 7m depth (Figure 3.20: Hight et al., 1992a). This indicates that the level of structure varies down the profile: since it does not systematically decrease with increasing depth there is an indication that conditions of deposition have varied during the accumulation of the sediments.

\(^{7}\) "Level of structure" is an inexact but useful term that indicates the size of the difference between the behaviour of a soil in the intact and remoulded states under the same stress regime. In the geotechnical context "structure" is defined as any component of resistance that is independent of the current void ratio and stress history (Leroueil & Vaughan, 1990). It is commonly assumed to be the result of some particular, three-dimensional arrangement of the soil particles but this is not automatically implied by the use of the term.
Figure 3.19 Schematic representation of the behaviour of intact and reconstituted samples during the destructuration process, during compression (a) and shear (b).
Figure 3.20 Profile of *in situ* void index for the Bothkennar clay.  
(Source: Hight *et al.*, 1992a; figure 13a).
3. One dimensional compression tests indicated the existence of a well-defined peak yield stress (Nash et al., 1992b). This varies with depth and corresponds to a yield stress ratio (OCR) of around 1.5 – 1.6 over most of the profile.

4. Those tests which took the soil past its initial failure condition caused a change in behaviour such that at large strains the measured parameters approached those of the soil in its remoulded state (Smith et al., 1992), indicating that a progressive process termed 'destruction' had occurred.

Hight et al. (1992a) related the level of structure exhibited by various samples to their visible soil fabric, using as a framework the sedimentary facies defined by the writer and her colleagues (Paul et al., 1992a; Barras & Paul, 1999; see also Chapter Five). Two caveats are required here: the number of individual specimens tested from each facies was very small (one or two only) and the samples from the laminated facies may have suffered considerable destructuring on sampling due to water loss via the laminaions, as acknowledged by Hight et al. (1992a).

However, from their review of the characterisation study, Hight et al. (1992a) concluded that, over a range of different tests, the level of geotechnical structure (however expressed in any particular test) was always in the sequence mottled facies > bedded facies > laminated facies. They found in particular:

1. That the mottled facies was stiffer in compression (before failure) than either the bedded or the laminated facies.

2. That the mottled facies was stronger than the bedded facies which, in turn, was stronger than the laminated facies.

3. That the mottled facies was softer (after failure) than the bedded facies.

The loss of geotechnical structure by a soil sample is known as destruction. This process was observed in all tests that involved straining an intact sample: as the strain increased so the level of structure declined and the response of the originally intact sample approached that of a remoulded sample (cf. Figure 3.19). Clayton et al. (1992) found that destructuring occurred progressively in both compression and in undrained shear. The results also showed that destructuring is not only progressive but also that it occurred more rapidly (in terms of strain) in the bedded facies than in the mottled facies: i.e. that it takes more strain to achieve a given loss of structure in the mottled facies. This implies that whatever is the aspect of fabric responsible for the level of geotechnical structure, it is more resistant to destructuring in the mottled facies than in the bedded.
3.6 SUMMARY

Taken as a whole, the results of the various investigations at Bothkennar have shown the following to be the case:

The sediments are silty clays and clayey silts which show considerable uniformity over the area of the Bothkennar research site. They are of medium to high plasticity in their natural state, although this is greatly influenced by sample preparation and does not appear consistent with their mineralogy. The earlier deposits in the area (known from the BGS studies) are of similar grading and mineralogy but appear to be much less plastic.

In their natural state the sediments are highly structured in the geotechnical sense, the mottled facies more so than the bedded. This structure is easily lost during sampling and in the early stages of many routine tests. There is some evidence that the level of structure varies with the visible fabric of the sediment, although this is based on rather few samples. Destructuration is progressive and is achieved more rapidly in the bedded facies than in the mottled. This suggests some form of bonding which in the mottled facies is stronger but also more flexible.

There is a distinct geotechnical stratigraphy at the site that can be detected in many property profiles, particularly those normalised for the effects of in situ stress, such as void index, normalised undrained shear strength and yield stress ratio. It is usually found that several distinctive zones exist below the desiccated crust: a zone of downwards-increasing water content and void index, elevated yield stress ratio and elevated shear strength, which extends to about 5m - 7m depth; a zone of high water content, low normalised shear strength and higher sensitivity, below which water content reduces with depth, shear strength increases and the soil shows the expected characteristics of a normally consolidated clay; a narrow zone of lowered water content, raised strength and irregularities in some property profiles. This zone terminates downwards against the Bothkennar Gravel Formation.
CHAPTER FOUR

PROGRAMME OF WORK

4.1 INTRODUCTION

4.2 SAMPLE COLLECTION AND FIELD PHOTOGRAPHY

4.3 INVESTIGATION OF SEDIMENT FABRIC

4.4 INVESTIGATION OF SEDIMENT COMPOSITION

4.5 INVESTIGATION OF GEOTECHNICAL PROPERTIES

4.6 SUMMARY
CHAPTER FOUR

PROGRAMME OF WORK

4.1 INTRODUCTION

This chapter details the practical programme carried out by the author between 1988 and 1995 at the Bothkennar research site. This includes all the field sampling and laboratory testing carried out between 1988 and 1994 during the three SERC-funded grants previously mentioned and also some subsequent work. The programme was extensive in its scale and made use of a wide range of techniques, some novel in this type of research.

The aims of the present study arose as a consequence of the geotechnical results outlined in Chapter Three and from a desire to understand the geotechnical character of the sediments in their geological context. They were:

1. To describe the sediments of the Claret Formation at Bothkennar in terms of their macrofabric and microfabric, with an emphasis on those aspects that might explain the evidence of `geotechnical structure` exhibited by samples when tested in the intact state.

2. To describe the sediments of the Claret Formation at Bothkennar in terms of their composition, with an emphasis on those aspects that might explain the geotechnical behaviour exhibited by samples when tested in the intact and remoulded states.

3. To relate these aspects of composition and fabric to the source material of the sediments, the geological processes by which they had accumulated and to their subsequent post-depositional history.

4. To explain the vertical profiles of selected geotechnical properties by reference to variations in the depositional conditions, to relate these to a palaeoenvironmental model based on the Holocene evolution of the Bothkennar area and thus erect a local stratigraphy with engineering geological relevance.

5. To compare the geotechnical character of the Bothkennar clay sediment with that of other soft clays of Holocene age from analogous areas, to identify similarities or differences and to explain the reasons for these.

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Chapter 4  Programme of work

The programme of work fell into four parts: sample collection, fabric studies, compositional studies and geotechnical studies. The general programme was as shown below, although inevitably there was some overlap and the work did not proceed in a neat, linear fashion.

4.2  SAMPLE COLLECTION AND FIELD PHOTOGRAPHY

4.2.1  Design of the Field Sampling Programme

The results of the initial site investigations carried out at Bothkennar by Bristol University were reported in Chapter Three of this thesis. With the preliminary Bristol reports to hand (Nash & Lloyd, 1988a, 1988b, 1989) it was possible, for this work, to design the engineering geological sampling against a number of objectives:

- Samples should be as little disturbed as possible and taken with high quality equipment, using the best available practice within the resources available;

- For the initial engineering geology study at least one continuous borehole was required to sample to the maximum thickness of the Holocene soft clay deposit (Claret Formation) at the site;

- Other boreholes were used to investigate the deposits in greater detail. These investigations were restricted to the upper part of the sequence for practical reasons.

Figure 4.1 shows a plan of the site illustrating the depth to the Bothkennar Gravel Formation (i.e. the local thickness of the Claret plus Grangemouth formations above the buried gravel layer) and the lateral extent of the laminated unit in the south-east corner of the site. These two factors had a major impact on the choice of sampling positions across the site.

A total of eight borings and one trial pit were made between 1988 and 1999 during the Heriot-Watt University investigations. Figure 4.1 shows the positions of all these excavations: piston sampler boreholes are prefixed HW and TP1 refers to the trial pit. Table 4.1 is a summary of the locations, elevations, depths and excavation/sampling methods used in each case.

This Thesis is based on the results from boreholes HW3, HW7, HW8 and HW9. Borehole HW4 and trial pit TP1 were used for other investigations that do not form part of those reported here. Boreholes HW1, HW2 and HW6 are also shown on Figure 4.1 for completeness but did not yield useful data since sampling problems caused unacceptable disturbance. Finally, borehole HW5 in the south-east corner of the site terminated within the laminated channel infill and did not sample the underlying Claret Formation. No use is therefore made of this borehole other than to provide an X-radiograph of the laminated facies, which typifies that present in channels within the Claret Formation more generally.
Figure 4.1 Sampling positions at the Bothkennar site, Heriot-Watt boreholes (HW) and trial pit (TP) are marked. (Source: gravel contours: Nash et al., 1992a; laminated unit: Nash & Lloyd, 1988a; aerial photograph base map: Royal Commission on the Ancient and Historical Monuments of Scotland, 1988.)
Table 4.1 Location and sampling details for the Heriot-Watt boreholes and trench.

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<tr>
<th>No.</th>
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<th>Depth sampled (m)</th>
<th>Sampling method</th>
<th>Date</th>
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HW = Heriot-Watt borehole.
TP = Trial pit/trench.
1 Boreholes shown in bold text are the subject of this study.
2 Local co-ordinates refers to a grid set up at the BSCRS (shown on Figure 3.8).
3 The trial pit was originally excavated for a project not under the writer's control and the exact date is not known.
Laboratory work on the samples so collected was carried out between October 1988 and September 1990, and between October 1991 and February 1996 and is described below. The majority of this work was carried out in the Department of Civil and Offshore Engineering, Heriot-Watt University (the Department), although some use was made of facilities outside the Department, both on the Heriot-Watt campus and at other Universities. These are detailed in Table 4.2.

4.2.2 Collection of Undisturbed Samples

It is recognised that no sampling method gives completely undisturbed samples. Studies of sample disturbance from earlier sampling campaigns at Bothkennar (Clayton et al., 1992; Hight et al., 1992b; see also Chapter Three) showed that sampling procedures, sample transport, storage and specimen preparation were all found to cause disturbance of varying degrees, in particular leading to a reduction in the mean effective stress of the sample (i.e. the introduction of a degree of apparent overconsolidation) (Hight et al. 1992b). This has particular significance for triaxial and oedometer tests for which the Sherbrooke sampler (Lefebvre & Poulin, 1979) and the Laval sampler (La Rochelle et al., 1986) were found to give less disturbed samples than did piston sampling (Hight et al., 1992b).

Despite its shortcomings, continuous flight thin-walled piston coring was considered to be the most appropriate method for geological fabric analysis within the resources available to this study. The use of high quality Shelby seamless tube with a wall thickness less than 2mm and a sharp cutting taper was expected on theoretical grounds (Baligh et al., 1987) to limit the disturbance due to stress relief to around 1% strain on the centre-line. This proved to be optimistic, since on examination of the core samples visible disturbance extended to a maximum of 150mm from the top of each 900mm section. It is thought this disturbance was the consequence of water escape during the sampling process rather than due to stress relief. For this reason in later phases of the work two adjacent boreholes (termed 'paired boreholes') were sunk at each location and a staggered coring interval used to ensure that every part of the profile was sampled in an undisturbed state. Although recorded, for completeness, by means of drawings and photographs these disturbed sections were not used for any electron microscopy or geotechnical tests. Disturbance in these sections was not always complete and on most occasions it was possible to distinguish details for facies description purposes.

4.2.3 Field Sampling Programme

In order to sample a maximum thickness of soft clay the first set of boreholes (HW1, HW2 and HW3) were sunk (~1m apart) in the south-west corner of the site where the soft clay extends from about 2.0m to a depth of about 20m below ground surface before the gravel layer is encountered
<table>
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(Figure 4.1). For these boreholes it was necessary to avoid the laminated clayey silt deposits which lie immediately under the crust in the south-east corner of the site.

Additional boreholes were used to sample the uppermost deposits. Two adjacent boreholes were sunk about 1m apart for boreholes HW5/6 and HW7/8 in order to provide overlapping, staggered samples so that the disturbed section at the top of each tube did not coincide. The boreholes HW5/6 (in the south-east corner of the site), HW7/8 (north-west corner) and the trial pit (TP1) were intended principally to investigate the nature of the surface crust and underlying transition zone and so were not intended to sample the complete soft clay sequence. In this Thesis samples from the borehole pair HW7/8 are used to supplement the data from HW3 since the techniques employed by the writer and colleagues had developed and improved by the time of their collection in 1991. In addition, a final borehole (HW9) provided some overlap with the uppermost part of the Claret Formation in borehole HW3 itself. This made available valuable information which was not collected from borehole HW3. The HW5/6 pair did not provide any data on the Claret Formation since both failed to penetrate beyond the Grangemouth Formation.

In all cases data from the existing Delft stocking boreholes D1 through to D6 and cone penetration profiles from the Bristol University initial site investigation (Nash & Lloyd, 1988a; 1988b; 1989; Hawkins et al., 1989) were used to build up a complete 3D structure of the site. This was supplemented by a personal examination and sampling of the curated material at the University of Bristol by kind arrangement of Mr D.F.T. Nash.

Continuous flight piston cores were taken by the Building Research Establishment (BRE) under contract to Heriot-Watt University. Following the problems encountered in recovery and disturbance in boreholes HW1 and HW2 where thin-walled acrylic tubes were used, further sampling was conducted using thin-wall aluminium alloy tubes (refer Table 4.1 for details).

Borehole HW3 was sunk in April 1989 to a depth of 19.90m bgl. It should be noted that this borehole was the only Heriot-Watt borehole to sample the complete Holocene sequence.

Boreholes HW7 and HW8 were sunk within about 1m of each other near to the north-west corner of the site (Figure 4.1) in October 1991 and an (almost) continuous overlapping flight of piston cores was obtained down to a depth of 6.7m bgl.

In November 1993 the final borehole (HW9) was sunk using Heriot-Watt University's own piston sampler, specifically to obtain fresh samples for geochemical analyses, which were required to complete the investigation into the organic content in the topmost 5m of the sequence. It was positioned as close to the former site of HW3 as possible (within about a metre), in order to obviate the need for a complete new logging and description exercise.
4.2.4 Field Photography

Photography was carried out in order to record most aspects of the fieldwork. A Canon EOS 500 SLR camera was used with 200 ASA Kodachrome Gold print and Ektachrome slide film. More recently, since 1996, 24bit colour images have also been recorded at 1280 x 960 resolution using a Kodak DC120 digital camera.

Routine photographs were taken during various stages of the coring in order to record the operation of the piston sampler and the recovery of the samples. General views of the site and sea defences were also recorded on a number of occasions and photographs were taken of the drainage ditches which bound the site on the north, south and east sides. These were taken since it was noticed that, in the south-east corner of the site on a flooding tide, there was an ingress of tidal water into the ditch from the drainage valve which was intended only to allow drainage of water out of the site (cf. section 2.5.2: Figures 2.15 and 2.18).

A limited study of tidal inundation over the mudflats and saltmarsh adjacent to the site was also undertaken in order to observe the extent, rate and pattern of tidal incursion at the present day. The study was made over twelve-hour tidal cycles for both spring and neap tides during the summer of 1996. This was felt to be a useful adjunct to inferences about tidal exposure that arose from the sedimentological facies model (see Chapter Eight) and for future work on the development of the crust.

For this study, photographs were taken both from the Kincardine bridge, facing south to southwest, taking in the area between the bridge and the Bothkennar site, and immediately adjacent to the site. From the bridge the tidal inundation was recorded at hourly intervals until 3 hours before high tide, and then half an hour after high tide, in order to record the water levels in relation to the main tidal channel in the Forth estuary. The extent of inundation of the saltmarsh immediately adjacent to the bridge where sea defences were lacking was also recorded.

Adjacent to the site, photographs were taken initially at about 4.5 hours before high tide (when the mudflats were completely exposed) and then at close intervals between 2.5 hours before and 1 hour after high tide in order to record the inundation of the mudflats and saltmarsh and the time taken for the water to subside again. Thus comparisons were made between the relative levels at spring and neap tides.

4.2.5 Curation of Piston Core Samples

Immediately after sampling, the aluminium tubes (with the exception of borehole HW9, see below) were cleaned, labelled and sealed at both ends. Samples must be preserved in order to maintain, as far as possible, the in situ physical conditions of the deposits, such as moisture content, oxidation state and soil structure and wax is used for this purpose. For this work a
50% beeswax:50% paraffin wax mixture was used as this has proved to be less brittle than paraffin wax on its own. The wax was warmed to 40°C: warm enough to melt the wax but not hot enough to bake the sampled material when it was applied. Wax was poured into the space left by the piston at the top of the tube and into the bottom of the tube where the bottom 1cm of material had been cut out. Successive layers of wax and cling film, followed by end caps and tape, were applied. This method was very efficient in excluding air and preventing oxidation, loss of moisture and shrinkage of the core material.

The cores were transported to Heriot-Watt University, supported on a foam rubber packing, and kept at a temperature between 5°C and 10°C until the material was extruded.

Piston cores from borehole HW9 were used for organic content determinations and were extruded on site immediately after sampling. After testing for Eh (section 4.4.2.3) they were immediately wrapped in clingfilm in an environmental chamber in an inert atmosphere (to excluded oxygen) taking care to ensure that any gas bubbles were exclude from between the core and the clingfilm. The samples were further wrapped in plastic and sealed with tape. These sections were transported to Heriot-Watt University and immediately frozen at -20°C until required.

4.2.6 Subsampling and Preservation of Subsamples for later use

Sub-samples were taken from the core material for subsequent SEM, mineralogical, geochemical and geotechnical analyses. The subsampling methodology varied for the different boreholes depending on the overall purpose of the project investigation. The suite of tests also grew steadily more extensive as experience was gained and so HW7/8 was investigated by a greater range of techniques than was HW3. The subsampling strategy in its final form is shown schematically in Figure 4.2(a) (HW7) and Figure 4.2(b) (HW8), and Table 4.3 shows the analysis intervals and techniques used for boreholes HW3, HW7/8 and HW9.

4.3 INVESTIGATION OF SEDIMENT FABRIC

A detailed analysis and description of the sediments at the site included a study of the macroscopic and microscopic fabric and features. Sedimentary structures such as the external form of the bedding, (shape, thickness, continuity of sedimentary units), internal structure of beds (graded bedding, cross-bedding) and bedding plane, sole markings and surface irregularities on the top and bottom of beds were examined. Biogenic traces and evidence of sediment reworking were also identified. The microfabric was studied in terms of the arrangement and composition of the minerogenic particles and biogenic material, and the form of biogenic structures.
Figure 4.2  Schematic diagram of subsampling strategy for extruded core sections: (a) borehole HW7; (b) borehole HW8.
Table 4.3  Analyses intervals for parameters: HW3, HW7/8 and HW9.

<table>
<thead>
<tr>
<th>Parameter measured/analysed</th>
<th>Analysis interval (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>HW3</td>
</tr>
<tr>
<td>Eh</td>
<td>–</td>
</tr>
<tr>
<td>pH</td>
<td>–</td>
</tr>
<tr>
<td>X-ray densimetry</td>
<td>Whole core</td>
</tr>
<tr>
<td>Photography of split core</td>
<td>Whole core</td>
</tr>
<tr>
<td>X-ray photography</td>
<td>–</td>
</tr>
<tr>
<td>Moisture content (w%)</td>
<td>0.05</td>
</tr>
<tr>
<td>XRD</td>
<td>0.90</td>
</tr>
<tr>
<td>SEM</td>
<td>0.05 – 1.0</td>
</tr>
<tr>
<td>Cations – Lithium acetate</td>
<td>Na⁺</td>
</tr>
<tr>
<td></td>
<td>Mg²⁺</td>
</tr>
<tr>
<td></td>
<td>K⁺</td>
</tr>
<tr>
<td></td>
<td>Ca²⁺</td>
</tr>
<tr>
<td>Cations - acid extractable</td>
<td>Fe</td>
</tr>
<tr>
<td></td>
<td>Mn</td>
</tr>
<tr>
<td></td>
<td>Ti</td>
</tr>
<tr>
<td>Cations - DCB extractable</td>
<td>Fe</td>
</tr>
<tr>
<td></td>
<td>Mn</td>
</tr>
<tr>
<td>Organic material</td>
<td>Total organic material</td>
</tr>
<tr>
<td></td>
<td>Polysaccharides</td>
</tr>
<tr>
<td></td>
<td>Methanol-toluene extract</td>
</tr>
<tr>
<td></td>
<td>Kjeldahl nitrogen</td>
</tr>
<tr>
<td>Particle size distribution</td>
<td>0.15</td>
</tr>
<tr>
<td>Atterberg limits</td>
<td>From natural w%</td>
</tr>
<tr>
<td></td>
<td>Air dried</td>
</tr>
<tr>
<td></td>
<td>Treated (H₂O₂)</td>
</tr>
<tr>
<td>Bulk Density + w%</td>
<td>0.15</td>
</tr>
<tr>
<td>Shear Strength + w%</td>
<td>0.15</td>
</tr>
<tr>
<td>Oedometer</td>
<td>–</td>
</tr>
</tbody>
</table>
4.3.1 Investigation of Macrofabric

This section describes the methods employed in the investigation of the macrofabric. In the first instance, non-destructive X-ray densimetry was used to produce a high resolution profile of changing density down the core. Only once these profiles had been obtained was the sample material extruded and systematically logged in detail.

4.3.1.1 Fabric investigations using X-ray densimetry

This work was carried out at the Department of Engineering Science, Oxford University by kind agreement with Dr G.C.Sills. The X-ray densimeter is a high-resolution scanning device normally used to make non-destructive measurements of bulk density in very soft clays (Been, 1981; Been & Sills, 1981; Edge & Sills, 1989). However, if it is used to scan a core of naturally structured sediment, the output trace reveals densimetric features which arise from the sedimentary fabric. Subsequent dissection of the scanned core, followed by careful comparison with the cleaned surface and X-radiographs, shows in many cases how the densimetric features seen on the output have arisen. The method was used to investigate the fabric of the cores from boreholes HW3, HW5, HW7, HW8 and TP 1 (Paul et al., 1996).

The bulk density is measured by counting the scintillations induced by a narrow X-ray beam that is passed through the sediment sample while still in its original sampling tube. The X-ray beam is attenuated by a varying amount due to different mineral densities and by different packing densities within the sediment column. Variations in mineral specific gravity are usually unimportant at the beam energies used in this instrument and so comparison of the attenuation with that caused by materials of known packing density allows the scintillation count rate to be calibrated for packing variations (i.e. fabric) alone.

This method relies on two components: a source of X-rays and a radiation detector. The source is a 160kV X-ray tube. This relatively high power (cf. X-ray photography below) is used in order to overcome the effect of differential absorption by minerals within the sediments (Been, 1981). A finely collimated horizontal beam of X-rays is traversed down the vertical core at a constant speed of 1.6mm per second. As this method relies on the core being held exactly vertical, precise measurements were made to ensure this.

The radiation detector consists of a sodium iodide scintillation crystal and photomultiplier which measures the attenuated X-ray count rate. The resultant trace is calibrated by means of at least two samples of known density (of similar material to that being measured), plus one of water, made up in the same core liner material as that which contains the core being analysed. The vertical position of the X-ray beam in relation to the core material was referenced by means of a measured datum on the core tube. The equipment was housed in a lead-lined room and
operated remotely once the X-ray beam was activated. The position of the beam in relation to the core material was noted in order to mark reference points on the chart recorder and relate this back to exact elevations, or depths, of the sample material. As the chart recorder speed (1mm s⁻¹ and 2mm s⁻¹ were used on different occasions) differed from the traverse speed of the X-ray head (1.6mm s⁻¹) these measurements were necessary in order to scale the plotter trace.

Where no X-radiographs of the core had been taken (as was the case with borehole HW3) and the visible features on the surface of the split core were not sufficient to detail all of the fabric and facies changes, this method of investigation proved to be very useful. By studying the photographs and drawings of the split core from these sections and making a comparison between their densimetry traces and traces from cores with a known internal fabric (where photographs, drawings, densimetry traces and X-ray photographs were available), it was possible to deduce the fabric and facies type of those cores were no X-radiographs had been taken. This method could possibly be further extended to make some degree of an estimation of the fabric and facies type of a core of soft clay material before it was extruded from its sampling tube.

4.3.1.2 Preparation of surface for core logging
Material from boreholes HW3 and HW7 was extruded by means of a hand-operated piston. Each 900mm length of sample core was extruded into a piece of half-round plastic guttering 300mm at a time at a rate of one or two per day. This length of material was chosen as a manageable length to handle in order to minimise sample disturbance and the handling time. Each section was covered with clingfilm in between procedures in order to prevent drying out of the material and oxidation of the surface.

The sections were split longitudinally by means of a fine cheesewire. Another half-round piece of guttering was placed on top of the section and the two halves were prised apart. One half was immediately covered with clingfilm; water content subsamples were taken from the other half which was then also covered with clingfilm.

The split surface was cleaned with an osmotic knife. This method of preparing a clean surface in a clay sediment relies on the temporary re-alignment of the clay platelets at the immediate surface when subject to an electric field, so preventing sticking or smearing when the knife is drawn through the clay. In the opinion of the writer its use was a major factor in allowing the sedimentary facies to be seen clearly.

4.3.1.3 Visual record and description of the cores
The cleaned surface was photographed in order to obtain a permanent visual record. Borehole HW3 was photographed using colour only, while borehole HW7 was recorded using both black
and white and colour photography. A scale bar, a Munsell colour chart (Munsell Color, 1975) and an identification label were included in each frame. Photography was carried out immediately after cleaning the surface with the osmotic knife: the time of exposure of the surface of the split core was minimised in order to preserve the freshly-exposed, unaltered colour of the sediments. This was particularly necessary where mottles were present as they became less distinct when the surrounding darker material became oxidised after exposure to the atmosphere. A4-size prints of each 300mm section photographed were obtained in order to have a close to life size record of the core.

A representational drawing of all visible features to be seen on one of the exposed faces of each 300mm core section was made onto an acetate sheet. These drawings are a useful backup as photographs viewed at a later date have a lower resolution than the naked eye. Observations such as colour (using the Munsell colour chart); the presence and nature of boundaries between features; sand and silt laminae; mottling size, shape and frequency; shells and shell fragments; cracks and sampling disturbance were noted.

The sediments were classified into two main facies types: the bedded and the mottled facies. The criteria for this classification were based on the presence or absence of mottling which was generally directly related to the level of disturbance of primary sedimentary structures by bioturbation. The degree of bioturbation was classified according to the bioturbation index (BI) of Taylor and Goldring (1993) (Table 4.4). Mottling was present up to about 90% coverage of the vertical surface and the complete reworking of the sediments and loss of sedimentary structure was normally obtained at levels of ~5% mottling coverage and greater. Where mottling was sparse (<5%) the sediments were classified as bedded. These criteria were arrived at by the comparisons of the visual description, the X-ray densitometry trace and, where available, X-ray photographs (below) although the latter were not essential for this classification.

4.3.1.4  X-ray Photography of thin slabs
The macrostructure of the cores from borehole HW7 was recorded by means of X-ray photography using an 55kV X-ray set made by Scandinavian X-ray UK Ltd. This instrument was kindly made available by the Edinburgh office of the British Geological Survey.

After photographing and drawing the 300mm sections they were cut in half and slabs of material were cut (~17mm thick x 150mm long x 80mm wide) from the surface. A cheesewire was used in the first instance, and the slabs were then smoothed off with the osmotic knife in order to minimise any thickness differences which could have been misinterpreted as variations in density of the material. The slabs were kept at natural moisture content by wrapping in clingfilm.
Table 4.4 Bioturbation Index: each grade is described in terms of the sharpness of the primary sedimentary fabric, burrow abundance and amount of burrow overlap. (Source: Taylor & Goldring, 1993).

<table>
<thead>
<tr>
<th>Grade</th>
<th>Percent bioturbated *</th>
<th>Classification</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>0</td>
<td>No bioturbation</td>
</tr>
<tr>
<td>1</td>
<td>1-4</td>
<td>Sparse bioturbation, bedding distinct, few discrete traces and/or escape structures</td>
</tr>
<tr>
<td>2</td>
<td>5-30</td>
<td>Low bioturbation, bedding distinct, low trace density, escape structures often common</td>
</tr>
<tr>
<td>3</td>
<td>31-60</td>
<td>Moderate bioturbation, bedding boundaries sharp, traces discrete, overlap rare</td>
</tr>
<tr>
<td>4</td>
<td>61-90</td>
<td>High bioturbation, bedding boundaries indistinct, high trace density with overlap common</td>
</tr>
<tr>
<td>5</td>
<td>91-99</td>
<td>Intense bioturbation, bedding completely disturbed (just visible), limited reworking, later burrows discrete</td>
</tr>
<tr>
<td>6</td>
<td>100</td>
<td>Complete bioturbation, sediment reworking due to repeated overprinting</td>
</tr>
</tbody>
</table>

* The percentage bioturbation values are not an absolute class division and should be used as a guide only.
and keeping under humid conditions in sealed containers until all slabs were ready for processing.

The slabs were unwrapped and exposed to X-rays at 55 kV, 4mA for 1.5 minutes. The X-radiographs were developed and later printed onto photographic paper as positive contact prints. It should be noted that the radiographs were exposed and developed in conditions aimed at optimising the show of detail, so that the apparent contrast on photographs does not necessarily represent equivalent density changes. There is, nevertheless, a broad correlation between the positions of the dark bands on the photograph and the higher density parts of the density profile. It has been found that densimetric features are often characteristic of particular structures in the sedimentary fabric and are thus diagnostic: both for the individual structures themselves and, as an assemblage, for the facies that contain them (Paul et al., 1996).

X-ray photography of the split cores proved to be very useful in the investigation of sedimentary structures. Primary bedding and associated boundary structures could be resolved and the traces of burrows left by macrofauna were often visible. Disturbance of primary sedimentary structures by bioturbation and sampling procedures could also be detected. Unfortunately, this technique was not available when borehole HW3 was investigated, and X-ray photographs are therefore only available for borehole HW7 (i.e. to a depth of about 6.4m bgl).

4.3.2 Investigation of the SEM Microfabric

The microfabric of the sediments was investigated by means of scanning electron microscopy (SEM\(^1\)). A total of 30 samples from boreholes HW3 and HW7 was prepared for examination for a qualitative study (at varying magnifications) and for a quantitative study at a fixed magnification (below). In total a library of over 2000 micrographs has been obtained. These were processed and printed personally by the author using facilities in the Department of Civil and Offshore Engineering, Heriot-Watt University.

4.3.2.1 Sample selection

Samples were selected both to provide systematic coverage of the sedimentary facies down the full length of the Bothkennar boreholes and to investigate specific features of the sedimentary fabric. For the quantitative study in particular the samples were selected in order to ensure that a variety of aspects of the varying sedimentology were covered. The selection of these samples and details of their setting within the profile are discussed in Chapter Five.

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\(^1\) SEM is used as an abbreviation for both scanning electron microscopy and the scanning electron microscope where its context in unambiguous.
4.3.2.2 Sample preparation

Preparation of the SEM samples was carried out at the School of Environmental Science, University of East Anglia by kind arrangement with Dr N.K.Tovey. The samples were prepared by freeze-drying using the technique of Smart and Tovey (1982). A small core of about 12.5mm diameter was cut from a subsample which had been preserved in wax under refrigerated conditions (above 4°C). The sample was first frozen in liquid nitrogen buffered by liquid propane and was then placed under vacuum in a temperature-controlled chamber, held at -50°C for about 24 hours, and then raised in a controlled manner to -20°C while the vacuum was monitored. The sample was returned to room temperature. In this way the production of interstitial ice was avoided, so preserving the delicate structural features of the clay fabric. Samples were carbon and/or gold coated before observation under the SEM.

A number of samples were air dried during the initial stages of the project. Although these were not suitable for fabric studies due to a partial collapse of the clay structure when drying by this method, micrographs from these samples have been used to illustrate some minerogenic and biogenic components (Chapter Five).

4.3.2.3 Preliminary observations

SEM images for preliminary, qualitative interpretation were collected at 20kV using a Cambridge Instruments S360 in the Department of Applied Geology, University of Glasgow. Around 2000 photo-micrographs were acquired at varying magnifications from about 7 times (x) which encompassed the whole sample, and then routinely at 100, 200, 500, 1,000, 2,000, 5,000 and 10,000x. At 100x, a micrograph width encompassed about 1mm of sample whilst at 10,000x (the maximum resolution of the SEM) a micrograph width was about 10µm: This allowed features to be seen at scales from about 2µm up to 200µm.

These micrographs are referred to as 'uncontrolled': although the chosen magnifications were systematically used on each sample, the position at which the micrograph was captured was chosen subjectively, although every effort was made to obtain micrographs which were representative of 'normal' structures as well as unusual structural features and biogenic material. Where a particular feature of interest was encountered, the magnification was adjusted to optimise the image of that feature and so was not confined to the routine set of magnifications. A selection of these micrographs has been published by Paul et al. (1992b). The range of features seen during these observations are described and discussed in Chapter Five (section 5.5.2).

4.3.2.4 Quantitative characterisation of SEM microfabrics

A series of micrographs on nine of the freeze-dried samples was obtained using a Cambridge Stereoscan 250 Mk2 at 20kV in the Department of Computing and Electrical Engineering.
Heriot-Watt University. This set of micrographs is termed 'controlled' and was used for the quantitative characterisation of the SEM microfabric. In all the sets every micrograph was taken under identical conditions (750x magnification, zero tilt). Sets of 24 micrographs were captured from each sample at randomly chosen positions and each individual micrograph was then subdivided into a grid of 24 contiguous squares (each 25μm by 25μm) using an acetate overlay. The features visible in both the grid squares and on the micrograph as a whole were classified according to a tabulated set of criteria. The discussion of these criteria and the interpretation of the microfabrics so classified are provided in Chapter Five (section 5.5).

4.4 INVESTIGATION OF SEDIMENT COMPOSITION

4.4.1 Mineralogy

The qualitative mineral content of the Bothkennar deposits was analysed using X-ray Diffraction (XRD), and an energy dispersive X-ray system (EDX) in conjunction with the scanning electron microscope. A binocular microscope was used to study sand- and coarse silt-size grains.

4.4.1.1 X-ray diffraction analysis

The bulk mineral composition at Bothkennar was determined by X-ray powder diffraction using a Philips PW1800 X-ray diffractometer with Cu Kα radiation. Whole sample splits at a number of positions down the soil profile from boreholes HW3 and HW7, and at two positions in HW9 were analysed. In addition, tests were carried out on selected coarse silt and clay-size fractions in order to determine the distribution of the minerals between the size fractions.

Bulk mineral composition was determined on powdered oven dried samples (105°C for 24 hours) in a presumed random orientation. Mineral identification was made by matching d-spacing values, calculated from 2θ diffraction angles measured, against tables of natural minerals (Brown & Brindley, 1980), and the principal reflections identified. Peak numbers, 2θ values and d-spacings were produced in the form of a table by the software associated with the XRD instrument. The methodology for these analyses is detailed in Chapter Six.

Samples of the clay-size fraction were analysed on a commercial basis by Dr. C.V. Jeans in the Department of Earth Sciences, University of Cambridge. They were analysed under three conditions: after air drying at room temperature; after furnace drying at 550°C and after glycerolation. The sample dried at room temperature served as the standard. The purpose of these treatments and their effect on the minerals in question is discussed in Chapter Six (section 6.2.5).
4.4.1.2 Energy dispersive X-ray system

Energy dispersive X-ray (EDX) analyses and automatic element identifications were obtained using a Link 10000 analyser fitted to the SEM, together with the manufacturer's software at The Department of Applied Geology, Glasgow University. The EDX system involves the collection of the characteristic X-rays which are generated from the sample while it is scanned by the primary electron beam of the SEM. This electron beam ionises the surface atoms of a mineral by ejecting electrons from inner orbital shells causing electrons from the outer shells to fill the inner shell vacancies in order to maintain stability. X-rays, whose energies are specific to the elements and orbital transitions involved, are released in the process.

The X-rays so generated are collected by an energy sensitive detector. Any major element in the sample will yield a peak on a graph (the EDX spectrum) whose axes are the energy level (X-axis) and the X-ray count detected at that energy level (Y-axis). The set of peaks is identified by a pre-programmed marker system within the analyser software.

The particle in question is normally isolated by using magnifications to 20,000x to 50,000x or by adjusting the electron beam to scan a reduced area. It can be difficult to isolate very thin particles (e.g. clay minerals) as the X-ray beam can completely penetrate the particle and a combined signal is obtained from both the mineral being scanned and its substrate. If this substrate can be identified, then its signature can be subtracted from the spectrum and the identification of the target particle may be resolved. In this respect the use of polished resin impregnated blocks can sometimes give a better signature from clay particles, since the surface topography of an unpolished sample is usually dominated by quartz grains which create high points and cause the smaller clay particles to be hidden from the X-ray beam (Welton, 1984).

Mineral identification of the target particle was checked by comparing relative peak heights using spectrograms published in the AAPG SEM petrology atlas (Welton, 1984) combined with direct SEM observation of the particle morphology. Elements which yield low energy levels (e.g. Na and Mg) result in peak heights which are reduced relative to their concentration, because these X-rays are partly absorbed by the Beryllium window of the detector. Thus, for elements with a low atomic number, comparison between peak heights and concentrations are not reliable and identifications which rely solely on these elements are suspect. Elements which have atomic number <11 (sodium) are not detected by EDX.

4.4.1.3 Sand fraction analysis using the binocular microscope

The coarse silt- and sand-sized particles (retained on the 38μm and 63μm sieves respectively after wet-sieving) were air dried and examined at 50x magnification using a Zeiss Axioshot optical
microscope. Determination of mineral content and morphology were the main objectives of this part of the investigation.

4.4.2 Inorganic Geochemistry

4.4.2.1 Determination of exchangeable cations (Na\(^+\), K\(^+\), Mg\(^{2+}\), Ca\(^{2+}\))

Exchangeable cations were extracted from the soil by leaching with ammonium acetate (Na\(^+\), K\(^+\), Mg\(^{2+}\)) and lithium chloride - lithium acetate solution (Ca\(^{2+}\)) in accordance with the Department of the Environment/National Water Council (1983). This process removes those cations which are adsorbed onto clay minerals by displacing them with competing ions. The leachate was analysed by means of atomic absorption spectrophotometry (AAS) using a Pye Unicam SP9 atomic absorption spectrophotometer.

4.4.2.2 Determination of iron, manganese and titanium

These cations were assayed only in boreholes HW7 and HW9 in order to identify possible cementing agents within the soil fabric. Two different extraction methods were used:

1. the relatively mild sodium dithionite-sodium citrate-bicarbonate (DCB) digestion method, which principally digests those amorphous and crystalline iron oxides and hydroxides that act as particle coatings. It has a negligible effect on iron within the clay structure itself;

2. the more aggressive nitric acid-peroxide method which also digests particulate material.

Both of the methods are described in detail in the literature (Mehra & Jackson, 1960 and Krishnamurty et al., 1976, respectively) and are therefore not detailed here.

4.4.2.3 Eh measurements

The accurate measurement of Eh (the oxidation state of the sediments) required that it be measured either in situ or immediately after the material was brought above ground, with as little exposure to the air as possible as Eh values change rapidly once a soil has been exposed to the atmosphere or as a result of diffusion. The cores from borehole HW9 were extruded immediately upon recovery and the Eh measured using Corning 250 Ion Analyser with an Orion platinum redox electrode. The area to be measured was first scraped with a palette knife in order to expose a fresh surface and the electrode was immediately pushed directly into the core. The value was recorded as soon as the reading had stabilised.

4.4.2.4 pH measurements

pH measurements were made using a Corning 250 Ion Analyser and Corning pH combination electrode and automatic temperature control probe. Three readings were taken for each sample.
and the average calculated. The samples were prepared in accordance with Smith & Atkinson (1975) by mixing dry soil with distilled water in a ratio of 1:2.5 and agitated using a magnetic stirrer for 15 minutes. The electrodes (above) were introduced into the slurry and the value recorded when the reading had stabilised. The ion analyser was calibrated by means of standard buffer solutions at pH 4 and 7.

4.4.3 Organic Geochemistry

Organic material occurs in estuarine sediments in a number of forms, which originate in the water column or through the activities of benthic organisms (McLusky, 1987; Moore, 1987; Paterson et al., 1990; Syvitski, 1991; Paterson, 1992). Following post-depositional degradation sedimentary organic material consists of a mixture of identifiable structural components, amorphous materials and secondary degradation products. In the present work this complex mixture has been characterised on the basis of total organic material, monosaccharide residues, organic nitrogen and methanol-toluene extractable material, a relatively simple system that has been shown to have geotechnical relevance (section 7.3.5; see also Paul & Barras, 1999).

The nature of sample collection for organic geochemical analyses developed over the course of this work. Samples from boreholes HW3 and HW7 were sealed in clingfilm and wax and refrigerated until required for analysis whereas those from HW9 were frozen and kept at -20°C until required. Freezing of the samples was to prevent the growth of bacteria and the post-sampling digestion of organic material. The subsamples (HW9 only) were removed in an inert (argon) atmosphere in order to inhibit the oxidation of organic materials present.

4.4.3.1 Total organic material

The content of total organic matter was initially determined by means of the hydrogen peroxide oxidation method for samples from borehole HW3. Although this method is recommended as a pre-treatment procedure, it is reported to eliminate organic material that does not consist of plant remains (Head, 1984) and by calculating weight difference before and after treatment an estimate of organic content can be obtained. However, potassium dichromate oxidation is the recommended standard method for soils (BS1377, 1990; Head, 1984: pg 249, 1992a) and was implemented for boreholes HW7 and HW9. The alternative method of loss on ignition is not recommended for clay soils due to dehydration of the clay minerals themselves and was therefore not regarded as suitable for this work.

4.4.3.2 Polysaccharides

Many intertidal and subtidal infaunal species are known to produce mucal polysaccharides with which they line their burrows and epipelagic diatom flora exude mucopolysaccharides onto the
sediment surface and in the upper layers of the mudflat deposits (Holland et al., 1974; Frostick & McCave, 1979; Paterson et al., 1990; Paterson, 1992).

Polysaccharides were hydrolysed to monosaccharides using 1N sulphuric acid in a water bath at 30°C for 1 hour and the Dubois phenol - sulphuric acid method (Dubois et al., 1956). The monosaccharide yield was then assayed by visible light colorimetry at a wavelength of 485nm. This method simply gives the total monosaccharides present in the sample and does not distinguish between the different types of sugar residues, nor does it identify separately the more complex amino sugars that commonly form such mucal materials.

4.4.3.3 Organic (Kjeldahl) nitrogen
The measurement of organic nitrogen is a proxy measure for protein, peptide and peptidoglycan residues in the soil profile. Most of the organic nitrogen present in organisms is incorporated into proteins: polymers of amino acids which have an amino group (NH$_2$) and carboxylic acid (COOH) group attached to the same carbon atom.

Organic nitrogen content in the sediments was determined by means of Kjeldahl reduction. These determinations were undertaken in the Department of Biological Sciences at Heriot-Watt University who had a well-established methodology for this procedure that is similar to that detailed in Smith and Atkinson (1975). The method involved the digestion of a known dry weight of sample with the addition of concentrated sulphuric acid into purpose-made digestion tubes which were then heated on a thermostatically controlled block. A subsequent distillation with sodium hydroxide and titration against boric acid with methyl-red/bromocresol green indicator completed the procedure.

4.4.3.4 Methanol-Toluene extractable organic material
Organic material which is extractable by solvents such as chloroform, hexane, toluene, methanol or acetone (Killops & Killops, 1993) includes the group 'lipids': a general term for compounds such as oils, fats and phospholipids (glycerides), waxes, steroids, pigments (e.g. carotene and chlorophyll) and a large variety of other ring compounds that are insoluble in water. It was not within the scope of this work to identify specific lipid compounds and extractions were of a general nature using soxhlet extraction apparatus and an azeotropic methanol/toluene mixture (40ml toluene:100ml methanol). Extraction proceeded over a period of 8 hours on air-dried material. The weight loss was taken to be the weight of material extracted and was calculated as a percentage of the original dry weight of soil.
4.5 INVESTIGATION OF GEOTECHNICAL PROPERTIES

4.5.1 Geotechnical Tests on Disturbed Samples

4.5.1.1 Particle size analysis

Particle size analyses (PSA) were carried out using a Micrometrics SediGraph 5100 particle size analyser in the Department of Geography and Geography, University of St. Andrews. This instrument measures particles between 0.1μm and 250μm by a settling technique based on the application of Stokes law.

Samples were air dried, carefully ground with a mortar and rubber pestle and passed through a 250μm sieve. 3.0g of sample were mixed with 40ml of 1% calgon solution (a dispersing agent) and agitated in an ultrasonic bath for 25 minutes. The sample was then stirred with a magnetic stirrer for 25 minutes and introduced into the SediGraph for analysis. This procedure gave reliable, reproducible results, although the strength of the Calgon solution was increased to 2% for those samples which exhibited persistent flocculation (evidenced by the cumulative frequency graph dipping sharply at about the 2μm size fraction). The procedure was later modified to include treatment with 20 Vol. hydrogen peroxide before preparation for PSA (borehole HW7/8 Atterberg limit samples only) and the use of 2% calgon in place of 1% as the dispersing agent for all samples from boreholes HW7/8 and HW9.

This procedure and the SediGraph method appears to be a reproducible and effective method since it gives consistent and higher values of clay content from the Bothkennar sediments than those obtained by the conventional pipette or hydrometer methods (BS1377, 1975).

4.5.1.2 Water content

Detailed water content profiles were obtained for all of the boreholes. Values for the water content of each borehole were obtained at the time of opening of each core section whereas water content measurements for geotechnical tests or geochemical analyses were made at the time of the test, when weighed subsamples were taken or when material was first exposed for drying (as appropriate for each method). Determinations were made by oven drying weighed samples (to 0.0001g) at 105°C for 24 hours. Water content was then calculated by taking the weight loss as a percentage (or proportion where appropriate) of the dry weight.

4.5.1.3 Atterberg limits

Atterberg limit tests (liquid limit (LL) and plastic limit (PL)) were carried out using the cone penetrometer in accordance with BS 1377 (1975; 1990) although with varying treatments. These were: testing from natural water content, after air drying and after treatment with hydrogen peroxide to remove organic material. For these tests a bulk, homogenised sample was split and
the treatments and tests done on separate subsamples in order to eliminate any effect of overworking by repeated testing of the same material.

4.5.2 Geotechnical Tests on Undisturbed Samples

4.5.2.1 Bulk density
Bulk density determinations were made on intact samples by filling a cutting ring of known volume (12.25cm³). After sampling, the cell was cleaned off, weighed and the bulk density calculated.

4.5.2.2 Compression and yield stress
The uniaxial compression index and yield stress were determined for samples from borehole HW7 by standard (incremental load) oedometer tests in accordance with Head (1984). The loading interval was 24 hours (sufficient for 95% dissipation of excess pore pressure) and the loading range was 22kPa – 880 kPa in doubling increments. The approximate yield stress was obtained by locating the turning point of the compression curve using the Casagrande method (Casagrande, 1936).

4.5.2.3 Undrained shear strength
The laboratory shear vane was used to carry out shear strength measurements on the sediments. For measurements on samples from HW3, the sample was put into a cylindrical cell approximately 40mm in diameter and 21mm in height (measured to 0.1mm) with a cutting edge angled at 30°. The cell was carefully pushed into the material to be sampled and cut at both ends with a cheesewire. The cell was weighed for bulk density determinations before being tested for intact and remoulded shear strength using the 12.5mm x 12.5mm vane. Water content determinations were made on each sample immediately after the shear vane test was complete. An alternative procedure was carried out for boreholes HW7 and HW8 where the vane was introduced directly into the whole diameter of the core in accordance with BS 1377 (1990). The results (cf. Chapter Seven) are comparable from both procedures.

4.6 SUMMARY
The programme of work carried out during the sampling, geological description and characterisation of the Bothkennar sediments involved a wide variety of techniques. Some of these were standard or accepted methods for the description and testing of soil properties, others were devised at Heriot-Watt University and improved upon as the projects developed. Nevertheless, it is always possible, on hindsight, to come up with improved techniques and these will be discussed later with suggestions for further work on the Bothkennar sediments. The results of these investigations are detailed in the next three chapters.
CHAPTER FIVE

PROPERTIES OF THE CLARET FORMATION AT BOTHKENNAR
PART 1: SEDIMENT FABRIC

5.1 INTRODUCTION

5.2 SEDIMENTARY FACIES

5.3 FINE-SCALE DENSIMETRIC FEATURES

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CHAPTER FIVE

PROPERTIES OF THE CLARET FORMATION AT BOTHKENNAR

PART 1: SEDIMENT FABRIC

5.1 INTRODUCTION

This chapter presents in detail the sedimentological results from the investigations of the Claret Formation at Bothkennar. Together with the compositional data to be presented in Chapter Six, these provide the framework within which the new geotechnical results, presented in Chapter Seven, are placed. The central concept is that the sediments can be divided into a limited number of facies which occur repetitively throughout the sequence, although there is a systematic change in their relative proportions. The facies have been defined initially on strictly lithological criteria (including the style of bioturbation) and their genetic interpretation is taken as a separate issue. This approach is considered essential in order to avoid any circular argument in the palaeoenvironmental interpretation which is undertaken in Chapter Eight.

The results are based principally on the very detailed study of the core from borehole HW3, which penetrated the full sedimentary sequence at Bothkennar. Additional fabric information has been obtained, where required, using cores from boreholes HW5 and HW7.

As described in Chapter Four, a range of techniques has been used to obtain these results. Prior to opening, the cores have been scanned using high-resolution X-ray densimetry as a tool to examine the sedimentary fabric: this non-destructive test is normally used to track the progress of early consolidation in engineering soils. They have been split, cleaned by osmotic knife, and logged in detail at the millimetre to centimetre scale. Using these logs as a guide, the cores have been subsampled for examination under the scanning electron microscope. This has led to a large volume of data which is presented both descriptively, indicative of the range of features to be seen, and as a systematic statistical study of the distribution of these features within a defined microfabric classification.

5.2 SEDIMENTARY FACIES

5.2.1 Overview

The sediments of the Claret Formation can be divided principally into a bedded facies and a mottled facies on the basis of their primary fabric and the extent and nature of subsequent bioturbation. In the bedded facies three subfacies have been identified: [normal] bedded,
burrowed bedded and basal bedded. Within the mottled facies it is also possible to recognise three distinct subfacies, in this case based on the style of mottling: types I, II and III mottling, described below. In certain parts of the BSCRS there occurs a laminated facies of more restricted extent, apparently confined to channels of varying size, which it is possible to recognise as discrete units within an otherwise mottled/bedded sequence. The characteristics of these facies are summarised in Table 5.1. The bedded and mottled facies replace one another repetitively throughout the sequence, as illustrated by the detailed lithofacies logs from boreholes HW3 and HW7 (Figures 5.1(a) and (b)). The key to these logs is shown on Figure 5.1(b). The log from borehole HW3 also shows how the lower part of the succession is dominated by the bedded facies and the upper part by the mottled facies: a similar pattern is seen in the borehole logs from the earlier investigations (Hawkins et al., 1989) (cf. Figure 3.12). Figures 5.2(a) and (b) show the logs for boreholes HW3 and HW7 respectively in a more generalised form. The key to these logs is shown on Figure 5.2(b). These have been used in Chapters Six, Seven and Eight where an illustrative log is shown beside profiles of sediment properties.

In the following section examples of different subfacies are shown in Figures 5.3 to 5.12. It should be noted that the X-ray densimetry traces included in these figures are not discussed in detail until section 5.3.

5.2.2 Bedded Facies

This facies (Figure 5.3) comprises black to dark grey (Munsell colour 5Y2.5/1 to 5Y4/1) silty clay sediments whose individual beds are well defined on X-radiographs and range in thickness from a few millimetres to about ten centimetres. They generally exhibit a strong fining-upward character although this is not always apparent on examination of the vertical section but can be established by X-radiography and X-ray densimetry.

The beds are often separated by silty partings on which small scale ripples are occasionally visible. These are often symmetrical in cross-section, which may indicate formation by oscillating (probably tidal) currents. The upper surfaces of the ripples are frequently coated by a micaceous silt drapes which terminates in a sharp upper boundary. The lower contacts are usually sharp, with limited evidence of erosion of the underlying unit. There are occasional thicker laminae (3mm to 10mm) of medium to coarse silt which may have been deposited under temporarily more energetic conditions. Included in this facies are those horizons where limited bioturbation (evidenced by mottling) has occurred but primary sedimentary structures remain clearly visible. It has been found empirically that this usually corresponds to a relatively low density of mottling (perhaps around 5% or less areal coverage on a vertical surface).
Table 5.1  General characteristics of the facies of the Claret Formation

<table>
<thead>
<tr>
<th>Bedded Facies</th>
<th>Mottled Facies</th>
<th>Laminated Facies</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>General Appearance</strong></td>
<td>Black to very dark gray (Munsell colour 5Y2.5/1 to 5Y3/1)</td>
<td>Very numerous black to dark gray (Munsell colour 5Y2.5/1 to 5Y4/1) clayey silt layers, usually around 5mm to 10mm thickness, separated by thin laminae (typically 1mm to 4mm thickness) of clean, medium to fine silt. The silt laminae have sharp bases with evidence of erosion of the underlying clay layer. Individual laminae are often lenticular in form and may contain minor, symmetrical ripples of height ~3mm and length ~30mm. Each succeeding silt layer usually has a graded base and fines upwards to be terminated by an undulating, eroded upper contact. The thicker silt clay layers occasionally show local evidence of bioturbation, usually in the form of limited Type I or II mottling essentially similar to that described above.</td>
</tr>
<tr>
<td><strong>Bedded burrowed subfacies:</strong> Near the top of the Claret Formation the bedded facies contains rip-up clay flakes and pellets a few millimetres in size and locally abundant, occurring along distinct horizons above sharp, cross-cutting, erosive boundaries. Beds often separated by silty partings and burrow infills of coarse silt occur.</td>
<td><strong>Type I subfacies:</strong> The mottles are relatively large (10mm to 15mm) and usually appear irregular or elliptical in vertical section.</td>
<td><strong>Densimetric Features</strong> X-ray radiographs reveal that mottling is associated with internal reworking of the sediments and consequently with the partial or complete loss of any primary sedimentary bedding structures. The extent of reworking is indicated by the densimeter trace, which may be almost uniform.</td>
</tr>
<tr>
<td><strong>Basal subfacies:</strong> Near the base of the Claret Formation individual beds are frequently separated by laminae of coarse silt or fine sand. Lenses and pockets of coarse silt and fine sand also found.</td>
<td><strong>Type II subfacies:</strong> The mottling is finer (2mm to 4mm) and individual mottles are mainly elliptical or round with some curved, hooked or irregular shapes present 3mm to 10mm.</td>
<td><strong>Densimetric Features</strong> On the densimeter profile the thicker silt laminae normally appear as strong peaks, whereas the thinner laminae cannot be resolved. The magnitude of the peaks are generally higher than in the bedded facies and majority of the larger peaks are more nearly symmetrical. The intervening silt clay beds show the low relief saw-tooth pattern typical of graded units like that seen within the bedded facies.</td>
</tr>
<tr>
<td><strong>Densimetric Features</strong> Within individual beds the density structure can be resolved by X-ray densimetry. Individual inter-bed contacts and silt laminae are resolved and there is a 'saw tooth' pattern that arises from an upward gradation from higher density clayey silt to lower density silty clay within each sedimentary unit.</td>
<td><strong>Type III subfacies:</strong> The mottling is elongated (5mm to 10mm) but narrow (1mm to 2mm) and forms a fine, threadlike, subhorizontal pattern which is usually densely packed.</td>
<td></td>
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</tbody>
</table>
Figure 5.1(a) Schematic log for borehole HW3. The whole sequence of sediments from ground level to the base of the borehole are included for completeness. Refer to Figure 5.1(b) for the key to this diagram.
Figure 5.1(b) Schematic log of borehole HW7. The whole sequence of sediments from ground level to the base of the borehole are included for completeness. The key applies to Figures 5.1(a) and 5.1(b).
Figure 5.2 Generalised facies logs: (a) borehole HW3 and (b) borehole HW7.
Figure 5.3 shows a section of bedded sediments from borehole HW7. From the photograph of the split core [a] the thickness of the individual beds can only be made out with difficulty although the drawing does highlight further features recorded at the time of opening [b]. However, when the X-radiograph is studied [c], much more detail of the internal structure can be clarified. Silt, which appears as darker material on the X-radiograph, can be seen to occur in four ways: generally as part of the silty clay beds [A], as lenses [B], as fine partings [C] overlying fine grained (pale, less dense) beds [D], and as thicker (up to 15mm) beds of denser more silty material at [E]. Symmetrical ripples are visible at [F]. Some sparse mottling [G] is also visible and can be seen on the X-radiograph (near the top of the section) to be associated with burrowing.

The basal subfacies (Figure 5.4) is recognised only near the base of the Claret Formation. It is distinguished by the presence of frequent laminae and lenses of coarse silt and fine sand, which separate the individual beds, and by the presence of both whole valves and fragments of species such as *Spisula subtruncata* and *Turretella communis* (J.D.Peacock, pers. comm.) disseminated throughout the sediment and possibly derived from the underlying Bothkennar Gravel Formation (although these species tend to favour sandy and muddy habitats respectively (Schäfer, 1972)). Figure 5.4(a) shows a section from near the top of the basal bedded facies. Here disseminated shell fragments can be seen along with sandy silt laminae [A] and lenses [B]. While the shell fragments are better represented on the photograph of the vertical section [a], the laminae and lenses are better resolved on the drawing/sediment log [b]. Cracks which are probably due to post-sampling handling have formed along laminae [D].

Figure 5.4(b) shows the lowermost section of the basal bedded subfacies. Here, the silt laminae [A] and lenses [B] are much thicker and whole disarticulated valves [C] more common than those in upper section of this subfacies (cf. Figure 5.4(a)). Also, the laminae are seen to have rippled upper surfaces [D]. In other respects the subfacies is similar to the normal bedded facies. The grading, sedimentary structures and shell detritus suggest deposition in a more energetic environment than the overlying sediments, perhaps associated with tidal currents. It is recognised that as there was no X-ray photograph available for borehole HW3, it is possible that sedimentological features which were present may have been overlooked (cf. Figure 5.3).

The burrowed subfacies (Figure 5.5) contains distinctive primary structures which include symmetrical ripples developed on silt-clay surfaces, rip-up flakes and mudballs, all of which suggest higher energy conditions that have caused re-erosion of the sediments, possibly associated with subaerial exposure. The ripples are about 10mm in scale and symmetrical in cross-section. Their upper surfaces are frequently coated by a micaceous silt drape which terminates in a sharp upper boundary. The mudflakes and mudballs are a few millimetres in
Figure 5.3  Section of core from borehole HW7, 4.95 - 5.17 m bgl showing an example of the bedded facies. See text for explanation.
Figure 5.4(a) Section of core from borehole HW3, 19.36-19.60 m bgl, showing an example of the basal bedded facies. See text for explanation.
Figure 5.4(b) Section of core from borehole HW3, 19.72-19.90 m bgl showing the base of the basal bedded facies. See text for explanation.
Figure 5.5  An example of the bedded burrowed subfacies from borehole HW7, 2.17-2.47 m bgl. See text for explanation.
size and are locally abundant, occurring along distinct horizons above sharp, cross-cutting, erosive boundaries.

The biogenic structures from which the subfacies is named comprise large (10mm to >50mm) burrows of at least two types: U-shaped dwelling structures [A] possibly produced by the amphipod *Corophium* and funnel-shaped structures [B] possibly produced by the lugworm *Arenicola* (Kingston, pers. comm.; cf. also sections 5.3 and 5.4 following). These are not always visible on the split core: the U-shaped burrow [A] was evident and drawn on the sediment log but the funnel-shaped burrow [B] was identified from the X-radiograph. Bedding is often greatly disrupted around these burrow structures [C]. Smaller funnel-shaped burrows (5mm to 10mm) [D] were also found and these are thought to have been produced by the small gastropod *Hydrobia*. The burrows usually contain an infilling of fine to medium silt which shows little internal structure and are terminated by a bedding surface marked by a silty lamina.

5.2.3 Mottled Facies

Bioturbation, the reworking of the sediment by burrowing organisms, is one of the most significant postdepositional processes to have occurred in the soft clay sequence. Sediments which exhibit a bioturbated fabric can be considered to be a modified version of the bedded fabric: they are dominantly silty clays in which primary bedding and laminations are poorly defined or absent due to the activities of burrowing organisms. Where biological activity has been intense this usually results in the total reworking of sediments and the loss of any primary sedimentary bedding structures. Bioturbation appears at a wide range of scales and different forms and is most significant in the upper and middle part of the Claret Formation.

The mottled facies (Figures 5.6 to 5.10) is composed of black to very dark gray (Munsell colour 5Y2.5/1 to 5Y3/1) silty clay in which primary bedding and laminations are poorly defined or absent due to bioturbation, which appears to be locally suppressed by siltier beds or laminae. The expression of this bioturbation is a lighter mottling (gray to olive gray, Munsell colour 5Y5/1 to 5Y4/2), easily visible on a freshly cut surface. On exposure to the atmosphere, oxidation of monosulphide in the surrounding, darker material soon occurs and within about 30 minutes the mottling is no longer visible.

It is possible to define three subfacies (types I, II and III) on the basis of the size and style of the mottling.

In the Type I subfacies (Figure 5.6(a) and (b)) the mottles are relatively large (10mm to 15mm) and are usually irregular or elliptical in vertical section. Generally, Type I mottles occur singly or in small clusters which usually cover up to about 10% of the surface at most, occasionally up to
Figure 5.6 Type I mottling from borehole HW7.
(a) Sparse mottling: 2.91 to 2.96m bgl. Note arrowed burrow in the X-radiograph associated with 10 x 20mm mottle on the photograph and sediment log. Bedding remains visible on X-radiograph and densimetry trace.
(b) Denser mottling: 6.07 to 6.12m bgl. Some bedding still visible, but the X-radiograph and densimetry trace show more disruption than in (a). The smaller mottles mainly visible on the sediment log are Type II mottles.
~40%. The X-radiograph in Figure 5.6(a) shows some disruption to the bedding, particularly at the burrow marked by an arrow. This is marked by a faint mottled area in the photograph. Figure 5.6(b) is an example of a Type I mottled area with increased mottling (cf. Figure 5.6(a)) and although there are some density differences on the X-radiograph, disruption of the bedding is much greater.

In the Type II subfacies (Figure 5.7(a) and (b)) the mottling is finer and individual mottles are mainly elliptical or round (2mm to 4mm), with some curved, hooked or irregular shapes (3 to 10mm) present. The X-radiograph shows severe disruption of the bedding in this section of the core at [A] (Figure 5.7(a)), with less disruption at [B] where the mottling is less dense and there is evidence of burrowing. Figure 5.7(b) shows an area of Type II mottles where it can be seen that the level of disruption is much less than in Figure 5.7(a) at [A]. It will be shown later (section 5.3.2) that the degree of mottling and extent of disruption do not always correlate.

Careful serial sectioning of the core shows that the burrowing probably resembles a loosely rolled ball of wool rather than a regularly branched structure which possesses a central vertical shaft. The Type II mottles themselves usually occur in dense patterns that can cover up to about 70% of a vertical surface. The different shapes on the surface of the split core are probably due to the cross-section being cut through different orientations of the same type and size of burrow. Elliptical and round mottles are occasionally seen to surround small holes (~1mm diameter) on the split core surface (both vertical and horizontal section). For example, Figure 5.8(a) illustrates a branched burrow and Figure 5.8(b) and (c) a hook-shaped burrow, both from a mottled horizon. These burrows are about 1mm in cross section and were surrounded by an oxidised (mottled) area.

In the Type III subfacies the mottling is elongated (5mm to 10mm) but generally narrow (1mm to 2mm) forming a fine, threadlike, subhorizontal pattern which is usually densely packed, often covering more than ~70% of the vertical surface. Figure 5.9 (between the depths of 4.27m and 4.42m in borehole HW7), illustrates very dense Type III mottling in the upper section [A]. Some of the mottles are visible on the X-radiograph as black, sinuous, horizontal areas [A] which are likely to be due to the mineralisation of burrows. Pyrite (FeS₂) has commonly been observed

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1 It might be suggested that elliptical mottles could be indicative of the effect of compaction on what were originally round mottles and that the degree of ellipticity could be a simple measure of the degree of consolidation. It should be noted, however, that if a round tube-shaped structure is cut at an angle then the cross-section would be an ellipse. Elliptical mottles are also seen on the horizontal cut surface and these could not be a result of compaction.
Figure 5.7(a) Example of Type II mottling (borehole HW7, 4.90m - 4.96m bgl). The denser mottling at [A] shows disruption of primary sedimentary structures on the X-radiograph and uniform X-ray density trace. In contrast, at [B] where mottling is less dense, the X-radiograph shows persistent bedding as well as small-scale burrows, probably associated with the formation of the mottles. Fine anastomosing burrows are also seen at [A].

Figure 5.7(b) Example of Type II mottling (borehole HW7, 6.17m - 6.20m bgl). The X-radiograph and X-ray densimetry trace show that primary sedimentary structures have been partially retained although bedding is not visible on the photograph of the core or the sediment log drawing. Fine anastomosing burrows can also be seen.
Figure 5.8 Close-up view of burrows associated with Type II mottling.
(a) Horizontal burrow with vertical and horizontal branches.
   Arrow points to branch perpendicular to plane of X-radiograph.
(b) Hooked burrow.
(c) Enlarged view of hooked burrow (b) taken at position (h).
   Mucal lining to burrow can be seen coating pyrite framboids.
   Cracks in the mucous are probably due to the air-drying process.
Figure 5.9  Example of Type III mottling (borehole HW7, 4.27m-4.42m bgl). The top section of the core at [A] shows a dense section of fine, horizontal mottling. The X-radiograph shows some of the associated burrows to be black due to mineralisation (probably with pyrite) and the densimetry trace shows increased density in this region. In contrast, in the section below [B] the mottling is less dense and mineralisation is absent. Total disruption of bedding has occurred, apart from at [B] and the oblique feature [C] which truncates the mottles and was logged as a clay horizon.
using EDX during electron microscopy \(^2\). Although this mineralisation corresponds to an horizon of Type III mottles, other similarly mottled areas do not exhibit this phenomenon. In this part of the core sedimentary structures have been totally reworked, apart from a more sparsely mottled area [B], and the oblique feature [C], which appears to be a clay horizon along which mineralisation has taken place.

The mottled subfacies frequently succeed one another in a regular, cyclic pattern (Figure 5.10). Complete cycles may occupy up to 500mm, although many examples of shorter, incomplete cycles are seen. Within a cycle, the transitions between subfacies are gradual and combinations of mottle types I/II and types II/III have been observed. All types also occur together with larger (5mm to 150mm) burrows and fine (0.5mm to 1mm) networks of burrows described below. A cycle usually commences with a short (20mm to 50mm) section of sparse types I, I/II or II mottles, followed by a gradational transition to dense Type II mottling usually over an interval of a couple of centimetres, although the change can occur within a few millimetres in some cases. The cycle is usually dominated by Type II mottling; Type III mottles usually occupy the uppermost few centimetres and may be absent entirely; such mottles appear to be due to horizontal burrowing immediately under a silty bedding surface. Mottled sequences are often abruptly terminated by the bedding surfaces of a local siltier bed or lamina leading to an incomplete cycle, above which a new cycle may often commence.

5.2.4 Laminated Facies

In the south-east of the site (cf. Figure 4.1) the upper part of the Claret Formation is cut out by a unit largely comprising sediments of the laminated facies referred to the Grangemouth Docks Member of the Grangemouth Formation (Paul et al., 1995). This unit occupies a discrete channel and rests on the underlying sediments with a strongly erosive contact (Hawkins et al., 1989). Elsewhere at Bothkennar, similar sediments occur within the Claret Formation as a minor unit of limited thickness (~1m) in boreholes HW3 and D1 (cf. Figures 4.1 and 3.8). The lower boundary has a strongly erosive lower contact with the underlying sediments (mottled facies) but passes upwards into the sediments of the mottled facies with a gradational contact. The unit is absent from nearby boreholes D5 and D6 (cf. Figure 3.8) and is also believed to represent a channel fill of limited extent. In borehole HW3 the unit is found between 7.70m and 10.15m but is

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\(^2\) The burrow infill shown in Figure 5.9 was examined under the binocular microscope and was seen to be packed with spherical minerals which were reddish yellow in colour. An XRD analysis of the material identified gypsum (calcium sulphate: a breakdown product of pyrite (Deer et al., 1966)) and jarosite (hydroxide of potassium iron sulphate). As these two minerals would not have shown up as opaque on the X-ray photograph (Figure 5.9(b)) the author suggests that pyrite was originally present but has been broken down: the binocular microscope study and the XRD analyses on these samples were not carried out as part of the main study and the samples had probably deteriorated.
Figure 5.10  Example of cyclic succession of the mottled subfacies (borehole HW3). The arrow at top right points to oxidation associated with burrows near the edge of the core.
interrupted by units of the mottled facies (cf. Figure 5.1(a)) and is probably due to a temporary switching of the position of the channel.

The facies comprises a regular, repetitive sequence of black to dark gray (Munsell colour 5Y2.5/1 to 5Y4/1) silty clay layers (Figure 5.11), usually around 5mm to 10mm in thickness, separated by thin laminae (typically 1mm to 4mm thickness) of clean, medium to fine silt. The silt laminae have sharp bases with evidence of erosion of the underlying clay layer. Individual laminae are often lenticular in form [A] and may contain minor, symmetrical ripples of height ~3mm and length ~30mm. For the most part these beds are horizontal (or nearly so). Each succeeding silty clay bed usually has a graded base and fines upwards to be terminated by an undulating, eroded upper contact. Occasional cross cutting horizons [B] and minor funnel shaped burrow structures [C] are also visible. The X-radiograph reveals a wealth of such fabric information down to the limit of its resolution. Within the unit in HW3 there are two sporadic layers (~20mm) of fine sand which rest on the underlying beds with a strongly erosive contact.

The thicker silty clay horizons within the laminated facies occasionally show local evidence of bioturbation and exhibit intact primary sedimentary structures only where burrowing was relatively sparse (less than ~5% of the vertical split surface) and disturbance is minimal. The bioturbation is usually in the form of types I or II mottling essentially similar to that described above. The degree of disturbance to the primary sedimentary structure varies: where the mottling is relatively sparse disturbance is minimal; where silt partings between silty clay beds are just a fine dusting of silt, these can be disrupted; and where silt laminae are 2mm or more thick, only large-scale burrowing can be seen to disrupt the structure.

5.3 FINE-SCALE DENSIMETRIC FEATURES

The X-ray densimeter and its use in the high-resolution study of fabric features in soft clay material is described in section 4.3.1.1. The profiles obtained in this study highlight the fabric differences in each facies type and are reported below.

5.3.1 Bedded Facies

In the bedded facies, X-ray densimetry reveals the presence of primary depositional bedding and can resolve density variations within individual beds. A detailed comparison of the density profile with the X-radiograph (Figure 5.3) shows that individual bed contacts are visible [i], as are silt laminae [ii] and that a saw tooth pattern [iii] arises from the upward gradation from higher density clayey silt to lower density silty clay within each sedimentary unit. A silt lens can be seen at [iv] and a shell fragment registers a peak at [v].

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Figure 5.11 An example of the laminated facies from borehole HW5, 3.95 to 4.10m bgl. See text for explanation.
SEM observations (section 5.5 below; cf. Paul et al., 1992b) have shown that in the lower, denser zone many particles are individual mineral grains of fine to medium silt size, whereas in the upper, less dense, zone many ‘particles’ are silt-sized aggregations of clay platelets (flocs or biogenic aggregates) which are probably hydraulically equivalent to the silt grains of the lower zone. Similar features have been seen in artificially sedimented beds studied with this instrument (Edge & Sills, 1989) and it has been shown that the ability of a clay layer to support the next, overlying, silt layer is developed within a few hours, presumably by processes such as particle re-adjustment and chemical bonding.

In the basal bedded subfacies (Figure 5.4(a)) individual silt lenses and laminae are picked out as peaks on the densimetry trace. They exhibit localised coarsening upwards sequences at [i] and fining up sequences at [ii]. The densimetry trace in Figure 5.4(b) picks out sandy silt laminae and lenses [i] but also shows an overall increase in density with depth: the presence of an increased sand content in the lowest section of the borehole is reflected by this increase.

In the burrowed subfacies the X-ray densimetry profile (Figure 5.5(d)) shows that major burrows stand out as significant features, probably as a result of their silty infilling. Features visible include a major U-shaped burrow [i]; disturbed and broken laminae [ii]; minor burrows [iii]. The trace shows that the effect of this disturbance is to introduce small scale lateral variation which reduces the average density contrasts within the section and so smoothes the density trace in a manner analogous to, although less complete than, the bioturbation associated with the mottled facies (below). Overall, the effect of these sedimentological features is to introduce a corresponding variation into the densimeter trace with a resolution down to a few millimetres in some cases.

5.3.2 Mottled Facies

X-radiographs of sediments from the mottled facies reveal that the mottling is associated with internal reworking of the sediments and, consequently, with the partial or complete loss of any primary sedimentary bedding structures. Figure 5.12(a) (6.07m to 6.30m depth, borehole HW7) shows an example which straddles a bedded/mottled facies boundary [i]. It can be seen that after the transition from the bedded to the mottled facies, the densimeter profile becomes more uniform, due to mixing by bioturbation as indicated by a dense pattern of small mottling on the cut surface. The onset of this mixing occurs over a short distance which corresponds both to the sharp change from the bedded to the mottled density signature and to the visible increase in the density of mottling. It can be seen from the X-radiograph and the densimetry trace that, in this section, some remnants of original bedding remain even though mottling is quite dense.
Figure 5.12(a) Section of core from borehole HW7 showing the transition from bedded to mottled facies. The X-radiograph shows that bedding is not entirely disrupted in the mottled section. This is also reflected in the densimetry trace.
In some cases, the extent of reworking is indicated by the densimeter trace. In Figures 5.9 and 5.12(b) [i] the trace in the mottled facies is almost uniform. Sparse Type I mottling (Figure 5.6(a)) shows a density trace where the bedding is relatively undisturbed, whilst the section in Figure 5.6(b), which shows a denser coverage of mixed types I and II mottling, has been disrupted to a greater extent and this is reflected in the densimetry trace. In other examples (cf. Figure 5.12(a)) the loss of structure is less complete and the correlation between the intensity of surface mottling and the degree of internal reworking appears variable. However, a simple relationship might not necessarily be expected. The visible mottling is a chemical alteration of the surrounding sediment, whose extent depends on the local redox environment and rate of alteration and thus does not always provide a simple trace which can be used to determine the proportion of the actual biogenic disturbance. Also, some reworking appears restricted to individual layers and so, although it creates surface mottling, does not remove completely the density signatures of the different layers.

5.3.3 Laminated Facies

The X-ray density profile (Figure 5.11(d)) reflects the laminated structure: there is broad correlation between the higher densities and the darker features apparent on the radiograph, although not all the thin, high contrast features correspond to an equally thin layer on the densimeter profile. In general, however, the thicker silt laminae normally appear as strong peaks, whereas the thinner laminae are below the nominal resolution (~2mm) of the instrument.

Two aspects are noticeable in particular: that the magnitudes of the peaks are generally more uniform than in the bedded facies and that the majority of peaks are symmetrical [i] (Figure 5.11(d)). The former effect is clearly due to the repetitive nature of the silt laminae and the latter is due to their internal uniformity; only occasionally do gradations in density produce any discernible saw-tooth pattern. Conversely, the intervening silty clay beds show the low relief saw-tooth pattern [ii] typical of graded units like that seen within the bedded facies.

5.4 BIOGENIC FEATURES

5.4.1 Descriptions

The sediments in all the facies exhibit a variety of biogenic structures. These include not only the pervasive bioturbation which is associated with surface mottling, but also U-shaped burrows, funnel-shaped structures of varying size, large, irregular shaft-like structures and complex, anastomosing systems of fine burrows with multiple side branches and end chambers. Several of these structures tend to be associated with particular facies and to vary in frequency with depth in the succession. However, it must be pointed out that as these features are often only visible on
Figure 5.12(b) Section of core from borehole HW7 showing the transition from bedded to mottled facies. The X-radiograph shows that primary sedimentary structures in the mottled section have been entirely reworked. This is also reflected in the densimetry trace which shows a uniform trace in the mottled in stark contrast with the bedded section below.
the X-radiograph, their presence (or absence) below 6.37m bgl (the base of borehole HW7) is not always obvious.

*Macrofaunal burrows* occur on a much larger scale than those in the mottled facies (10mm to 30mm across, 25mm to 100mm in length) and were found in the upper part of the Claret Formation within the bedded burrowed facies described above. Bedding is often totally disrupted due to the size of the burrows in relation to the width of the core, although where the burrow is restricted to the edge of the core sedimentary structures may be visible in the relatively undisturbed areas. Burrows are often terminated by a bedding surface marked by a silty lamina.

These structures occur in a variety of shapes and sizes. Like types I to III mottles, they are produced by burrowing organisms but do not generally manifest themselves by lighter oxidised areas on the surface of the split core because the deposits in which they occur are generally lighter in colour (due to oxidation) than those where the mottles are found. They can be recognised by visual inspection of the split core where a burrow has been sectioned by the splitting of the core, particularly where the burrow infill is coarser grained than the surrounding material (Figure 5.5). Where the infill remains fine grained it is often seen to have been reworked by smaller burrowing organisms, displaying Type II mottles within the outline of the larger burrow. Very often these larger burrows are only distinguishable on X-radiographs.

Examples of these structures and their relationships to the host sediment are shown in Figure 5.13 from a section of borehole HW7 from 3.39m to 3.50m. *Funnel structures* are 5mm (b3 and b4) to ≥30mm (b1) in length and have been seen to show incremental internal layering which may be evidence of escape through an aggrading sediment bed. The funnel structure (b1) tapers downwards to a width of about 5mm and is truncated at the top by a fine (~1mm) silt lamina. U-shaped burrows (b2) are usually infilled with fine to medium silt which shows little internal structure and terminate against a bedding surface, often marked by a silty lamina. They often appear to penetrate other bedding surfaces with little disruption but this could be a three-dimensional effect.

*Finer feeding burrows:* Fine (0.5mm to 1.0mm diameter), occurring as narrow anastomosing (downward branching) burrows (Figure 5.14(a) and (b)) and broader anastomosing burrows (networks) (Figure 5.14(c) and (d)) have been noted at a number of depths between 3.00m and 6.24m in HW7. The former burrows are usually vertical, but also occur inclined at angles up to 45 degrees from vertical. Both of the burrow systems lack any infilling and appear to be a late development which cross-cuts other biogenic structures, particularly the pervasive bioturbation.

These fine burrows are most obvious as branching, light coloured (low density) features on the X-radiographs. The low density indicates that they do not have a silt infill but often terminate
Figure 5.13 An example of macrofaunal burrows at 3.39m-3.50m from borehole HW7. Note: b1: funnel-shaped burrow; b2: U-shaped burrow, both truncated by silt laminae. Smaller burrows (b3 and b4) can also be seen. R indicates ripple surface at top of clay-rich bed and overlain by silt. In (a) denser (silt) shows as dark areas and laminae. See Figure 5.6 for further examples of these burrows.
above at a silt lamination with a raised silt cast: probably a pile of faecal pellets (Schäfer, 1972: figure 159). Elliptical structures on branches of these burrows (Figure 5.14(a)-(d)) could be turning chambers where the burrowing organism physically turned in order to burrow in another direction, or they could be communal interconnections with larger burrows (Howard & Frey, 1975). The burrows are often seen to continue downwards through more silty horizons into finer material below. On photographs of the split core they can occur as barely discernible pale (oxidised), very fine (0.5mm across) markings following the directions seen on the X-radiographs. Small elliptical (1mm) Type II mottles also coincide with these burrows and therefore could be due to the oxidation of sediments in the vicinity of turning chambers. However, very few of the burrows seen on the X-radiographs appear on the photographs of the split core, probably due to the burrows being beneath the vertical cut surface of the X-rayed slab.

The networks of burrows (e.g. Figure 5.14 (c) and (d)) appear to be formed by organisms which can tolerate a more silty environment, being found in sediments from 24% clay content and above (up to 68% silt), whereas the downward branching form seems to favour clay contents of at least 35% (52-65% silt), although both styles were seen to burrow through siltier horizons.

The networks of burrows occur in conjunction with types II and III mottles, but do not appear to be responsible for the main bulk of this mottling. They only coincide with these mottles as very fine 1mm ellipses on the split core surface. The fine burrows appear to be more densely packed where the mottles are dense and correspondingly less dense where the mottles are sparse on the cut surface. Although it is possible that these two features have been caused by the same organism, most of the mottles do not appear to conform to the same shape and size as the fine burrows visible on the X-radiographs. It is feasible that the mottles are larger due to the fact that they have oxidised the material for a distance around the burrow and not just the burrow itself.

It seems more likely that these finer branched burrows are the result of burrowing by a different organism from that which caused the Type I and Type II mottles, and that they both inhabited the same environment and thrived under similar conditions.

Disturbance of primary sedimentary features by these fine burrows appears to be minimal. Most of the disturbance appears to have been caused by the organisms causing Type II mottles and where the finer burrows do breach silt laminae, disturbance is very localised due to the size of the burrows. These finer burrows have not been identified in HW3 as no X-radiographs were taken of this borehole and positive identification by means of photographs of the split core alone is not possible.

Biogenic structures were also identified under the scanning electron microscope and these are detailed in section 5.5.5.
Figure 5.14 Examples of small-scale anastomosing feeding burrows.
TC = turning chamber; SC = silt cast.
(a) X-radiograph of the narrower downward branching burrows with schematic drawing of the same feature (b).
(c) X-radiograph of a broader network burrow with schematic drawing (d).
See text for further explanation.
5.4.2 Interpretation of Biogenic Features

The larger-scale structures described above (cf. Figure 5.13) usually cannot be attributed unambiguously to particular organisms. However, comparison with the known infaunal ecology of the Forth estuary at the present day (Elliot & Kingston, 1987; McLusky, 1987; Moore, 1987; Kingston pers. comm.) enables some tentative deductions to be made, based simply on the most common candidates reported by the above authors. On this basis it seems likely that the U-shaped burrows are produced by the amphipod *Corophium*. Funnel-shaped structures occur over a range of sizes which suggests that more than one type of organism may be responsible: candidates include the small gastropod *Hydrobia*, and a number of the larger polychaete worms. The large, irregular shafts are believed to be sections of deep burrows produced by bivalves such as *Scrobicularia plana* or *Mya arenaria* which are deeper-burrowing organisms than those previously mentioned. The pervasive bioturbation is attributed both to polychaetes (e.g. *Nereis, Polydora*) and to meiofaunal oligochaetes (e.g. *Tubificoides*) and is associated with visible mottling as discussed earlier.

The fine burrow networks appear to be similar to the tiny feeding burrows of the polychaetes *Prionospio* (Howard & Frey, 1975), *Heteromastus filiformis* (Schäfer, 1972; Howard & Frey, 1975) or *Scoloplos armiger* (Kingston, pers. comm.). Although the latter organism prefers sandy rather than muddy substrates (Schäfer, 1972), the high silt content (above) could account for their presence. Schäfer (1972) states that *Heteromastus filiformis* inhabits anoxic deposits and that there is never an oxidation halo around its burrows; this could account for these burrows being visible on X-ray photographs but not on the split core and this supports the argument (above) that the mottles and the finer burrows were traces of different organisms. The anastomosing burrow systems (Figure 5.14 (c) and (d)) are also very similar to those of decapods such as *Upogebia* and *Callianassa* (Howard & Frey, 1973) although *Callianassa* are generally larger organisms and would probably leave traces on a different scale.

These structures show an inter-relationship that appears to be determined by depth and particle size. Figure 5.15 shows in detail, for borehole HW7, the nature and extent of biogenic disturbance in relation to the sedimentary profile, using the methodology of Taylor and Goldring (1993) (Table 4.4). It appears that discrete burrows are associated with slightly siltier layers and are most frequent in the uppermost 1m to 2m below the *Cerastoderma* bed: at greater depths discrete burrows are more sporadic. Pervasive bioturbation is confined to the more clayey layers, where it often obliterates all other structure except the anastomosing burrow systems. There appear to be two separate associations comprising the discrete funnel and U-shaped burrows in Figure 5.13, and the anastomosing burrows and pervasive bioturbation (Figure 5.14). This suggests either a temporal succession, in which the discrete-burrow association is replaced by
Figure 5.15  Relationship of biogenic traces to depth and facies in borehole HW7. The bioturbation index is that of Taylor & Goldring (1993) (Table 4.4). Two associations are seen: (A) discrete shafts, funnel and U-shaped burrows; (B) pervasive reworking with later, anastomosing burrows superimposed.
pervasive bioturbation, itself later to be overprinted by the anastomosing burrow systems, or that instead there has been mutual exclusion between the two associations, controlled by particle size or by inter-species interference such as that observed between Corophium and the polychaete Nereis (Olafsson & Persson, 1986).

It should be noted, however, that there can be a confusion with certain plant roots and these finer polychaete burrows (Howard & Frey, 1975). The investigation by Bristol University (Hawkins et al., 1989; 1991; Nash et al., 1992a) attributed mottling (= types I, II and III above) in the Claret Beds at Bothkennar to oxidation around rootholes (Hawkins et al., 1991) and, as no X-ray photographs of the cores were taken during their investigation, the finer networks of burrows described above were never identified. The writer found only occasional roots associated with oxidation and found no carbonaceous plant remains in association with either the types I to III mottling or the finer structures described above. The mottling, therefore, has been attributed to polychaete burrowing, although the possibility of a plant root system cast cannot be ruled out completely.

5.5 SEM MICROFABRIC

One aim of this research has been to describe the sediment microstructure and to link this to the description of the macrofacies. Although there are many descriptions of SEM microfabrics in the literature (Collins & McGown, 1974; Smart & Tovey, 1981; Bennett et al., 1991; Syvitski, 1991), these often focus on qualitative examples or on "interesting" features and no indication is given of how representative they are of the sediment in general. They also seldom attempt a quantitative assessment of the structures they describe. For this reason, a new methodology has been developed in order not only to describe the microfabric and its components, but also to assess the relative occurrence of different microfabric types within a given macrofacies.

5.5.1 Introduction and Methodology

It is possible to describe geological materials on a wide variety of scales. In the present case there are four scales under consideration: the core, the SEM sample, the overall micrograph and the features to be seen within it. The last two cases are clearly dependent on the magnification. Figure 5.16 shows these relationships in an idealised form: by coincidence there is approximately one order of magnitude between each scale. The features so described range from the macrofabric through the macrofabric elements (sometimes termed the mesofabric) to the microfabric proper.

The methodology adopted was as follows. Each SEM sample was categorised according to the local macrofacies from which it was collected (Tables 5.2(a) and (b) and Table 5.3). All the existing 'uncontrolled' micrographs already available (Paul et al., 1992b) from the Bothkennar micrograph
Figure 5.16 Schematic illustration of the hierarchy of fabric scales.

CORE
~100mm width

SEM SAMPLE
~13mm diameter

ELECTRON MICROGRAPH
750x magnification
~0.2mm width

MICROGRAPH GRID SQUARE
0.025 x 0.025mm
(25 x 25μm)

Macrofabric scale: the soil is described by a macrofacies

Microfabric scale: the soil is described by a microfabric type

Microfabric element: forms a component of the microfabric
Table 5.2(a) Summary of freeze dried samples for qualitative analysis.

<table>
<thead>
<tr>
<th>Borehole</th>
<th>Depth (m)</th>
<th>Macrofacies</th>
<th>Sample Setting</th>
</tr>
</thead>
<tbody>
<tr>
<td>HW3</td>
<td>2.30</td>
<td>Bedded (burrowed)</td>
<td>Disrupted bedded horizon</td>
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<tr>
<td>HW3</td>
<td>3.15</td>
<td>Mottled</td>
<td>Type II (dense)</td>
</tr>
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<td>HW3</td>
<td>3.80</td>
<td>Mottled</td>
<td>Type II (dense)</td>
</tr>
<tr>
<td>HW3</td>
<td>4.95</td>
<td>Mottled</td>
<td>Type II (dense)</td>
</tr>
<tr>
<td>HW3</td>
<td>5.70</td>
<td>Bedded</td>
<td>Mid-bed (fine)</td>
</tr>
<tr>
<td>HW3</td>
<td>5.76</td>
<td>Bedded</td>
<td>Mid-bed (fine)</td>
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<td>HW3</td>
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<td>Type I (sparse)</td>
</tr>
<tr>
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<td>7.05</td>
<td>Mottled</td>
<td>Type III (dense)</td>
</tr>
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<td>Type III (dense)</td>
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<td>Type III (dense)</td>
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<td>Across laminae</td>
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<td>8.85</td>
<td>Laminated</td>
<td>20mm above a 20mm thick silt/sand lamina</td>
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<td>9.76</td>
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<td>Type II/III (sparse)</td>
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<td>9.91</td>
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<td>Type II (dense)</td>
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<td>Type II/III (dense)</td>
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<td>Non-erosive contact</td>
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<td>2.26</td>
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<td>Major burrow structure</td>
</tr>
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<td>2.43</td>
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<td>Thick lamination, erosive contact siltlump horizon</td>
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<tr>
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<td>Type I (sparse)</td>
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<td>Mid-bed (silty)</td>
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<td>Non-erosive contact</td>
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### Table 5.2(b) Summary of air dried samples for qualitative analysis.

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<td>HW3</td>
<td>11.65*</td>
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<td>Mid-bed, possibly burrowed</td>
</tr>
<tr>
<td>HW3</td>
<td>13.00</td>
<td>Bedded</td>
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</tr>
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<td>HW3</td>
<td>14.95</td>
<td>Bedded</td>
<td>Very sparse Type I/II mottles</td>
</tr>
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<td>16.00</td>
<td>Bedded</td>
<td>Very sparse Type I mottles, possibly burrowed</td>
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<tr>
<td>HW3</td>
<td>17.05</td>
<td>Bedded</td>
<td>Very sparse Type I mottles, possibly burrowed</td>
</tr>
<tr>
<td>HW3</td>
<td>17.80*</td>
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<td>Mid-bed</td>
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<tr>
<td>HW3</td>
<td>17.80</td>
<td>Bedded</td>
<td>Mid-bed</td>
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<td>Mid-bed, shelly horizon</td>
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<td>Type III (dense)</td>
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<tr>
<td>HW3</td>
<td>14.06</td>
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<td>Type II/III (dense)</td>
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* Horizontal surface (parallel to bedding)

### Table 5.3 Summary of ‘controlled’ samples for qualitative and quantitative analysis.

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<td>Bedded</td>
<td>Penetrative contact</td>
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<td>Bedded</td>
<td>Thick lamination</td>
<td>SE2</td>
</tr>
<tr>
<td>HW7</td>
<td>5.79</td>
<td>Bedded</td>
<td>Mid-bed fine</td>
<td>SE6</td>
</tr>
<tr>
<td>HW7</td>
<td>6.30</td>
<td>Bedded</td>
<td>Mid-bed silty</td>
<td>SE3</td>
</tr>
<tr>
<td>HW7</td>
<td>6.32</td>
<td>Bedded</td>
<td>Non-erosive contact</td>
<td>SE4</td>
</tr>
<tr>
<td>HW7</td>
<td>4.30</td>
<td>Mottled</td>
<td>Type III dense</td>
<td>SE7</td>
</tr>
<tr>
<td>HW7</td>
<td>4.84</td>
<td>Mottled</td>
<td>Type II dense</td>
<td>SE9</td>
</tr>
<tr>
<td>HW3</td>
<td>7.21</td>
<td>Mottled</td>
<td>Type III dense</td>
<td>SE8</td>
</tr>
</tbody>
</table>
library (about 2000 in total, taken at varying magnifications (Table 5.4)) from each freeze-dried sample were examined and the recurrent features to be seen at scales from about 2μm up to about 200μm were described qualitatively. This has produced two sets of microfabric descriptors: one applies to the complete image at a scale around 100μm to 200μm (Table 5.5) and the other to image elements at a scale around 20μm (Table 5.6). It has been found that just three principal categories are sufficient to classify the majority of both images and image elements encountered, although additional categories (or sub-categories) have been used to distinguish significant minor features.

The above descriptors of the principal microfabrics were applied systematically to standardised sets of new ‘controlled’ micrographs (24 micrographs from each of the freeze-dried samples SE1 to SE9 taken at 750x magnification, zero tilt at randomly chosen positions). Each individual micrograph was assigned subjectively to one of the three categories given in Table 5.5 on the basis of the appearance of the micrograph as a whole. The micrograph was then subdivided into a grid of 24 contiguous squares (each 25μm by 25μm) and each grid square was assigned to one of the five sub-categories of the microfabric elements (Table 5.6). These procedures produced 24 classified images and nominally 570 classified microfabric elements per sample, although some squares could not be classified due to poor image quality. From these data were determined, for each sample, the relative frequency (1) of the microfabrics (section 5.5.3) and (2) of the microfabric element (section 5.5.4) 3.

5.5.2 Descriptions of Microfabrics

Examination of the micrograph library has shown that the three principal microfabrics described in Table 5.5 together account for over 90% of the images. These have been termed the aggregated, the granular and the cumulate microfabrics: Figures 5.17 to 5.19 show examples of micrographs from each type of fabric. The aggregated microfabric consists of clay platelets organised into flocs and in turn into larger aggregates of probable biogenic (or sometimes hydraulic) origin. The granular microfabric is composed largely of fine to medium silt particles in relatively close proximity, and is widespread in the bedded facies, often being associated with laminae, bed contacts and pockets of coarser material. The cumulate microfabric is distinguished

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3 The author is aware that the number of samples from each macrofacies type (6 bedded and 3 mottled) was not sufficient for a complete statistical classification of the microfabric types within each macrofacies. The number of subsamples for microfabric type and microfabric element frequency within each individual sample (24 and 576 respectively), however, is considered to be sufficient for a statistically valid classification within each sample. Two further samples from the mottled facies were prepared for this study but were found to be unsuitable: the samples had not been kept refrigerated before freeze-drying and a coating of bacteria had formed over the surface, making classification impossible.
Table 5.4  Scanning electron microscope fields of view at varying magnifications.

<table>
<thead>
<tr>
<th>Magnification</th>
<th>Approximate Field of View</th>
<th>Features visible</th>
</tr>
</thead>
<tbody>
<tr>
<td>~5x</td>
<td>~15 x 20mm (15000 x 20000µm)</td>
<td>Whole sample (13mm diameter) with topographical features and damaged areas.</td>
</tr>
<tr>
<td>100x</td>
<td>800 X 1200µm (0.8 x 1.2mm)</td>
<td>Lineations, usually horizontal to sub-horizontal occurring at 200µm intervals.</td>
</tr>
<tr>
<td>200x</td>
<td>400 x 600µm</td>
<td>Above-mentioned lineations in more detail, encompassing area between two or three lineations. Texture, eg. aggregated, and interaggregate voids.</td>
</tr>
<tr>
<td>500x</td>
<td>150 x 250µm</td>
<td>Clay aggregates, interaggregate voids and spatial relationships between aggregates. Bio-debris floating in clay matrix.</td>
</tr>
<tr>
<td>750x</td>
<td>100 x 150µm</td>
<td>As 500x and 1000x. Used specifically in qualitative and quantitative controlled micrograph study.</td>
</tr>
<tr>
<td>1,000x</td>
<td>75 x 125µm</td>
<td>Whole coarse silt grains. Complete clay aggregates measuring ~50µm. Juxtaposition of diatoms and medium silt particles with clay matrix. Circular burrow complex within clay matrix. Pyrite framoids infilling diatom frustules.</td>
</tr>
<tr>
<td>2,000x</td>
<td>40 x 60µm</td>
<td>Whole medium to coarse silt grains. Juxtaposition of diatoms, shell fragments and fine silt particles within clay matrix. Complete circular burrow complex. Complete disc-shaped diatoms.</td>
</tr>
<tr>
<td>5,000x</td>
<td>15 x 25µm</td>
<td>Clay honeycomb with clay platelet contacts. Whole diatoms with fine detail of their structural patterns. Mouth of individual burrow from burrow complex. Whole fine and medium silt grains. Contacts between silt grains.</td>
</tr>
<tr>
<td>10,000x</td>
<td>8 x 12µm</td>
<td>Individual clay platelets and their contacts, whole fine silt grains. Detail of mouth of burrow with micro-crystallites and mucous membrane, from circular burrow complex. Clay bridges and butresses between silt grains.</td>
</tr>
<tr>
<td>20,000x</td>
<td>4 x 6µm</td>
<td>Individual clay platelets. Few taken at this magnification as focussing proved difficult and micrographs obtained were unsatisfactory.</td>
</tr>
</tbody>
</table>
### Table 5.5 Summary of principal microfabric descriptors.

<table>
<thead>
<tr>
<th>Microfabric Descriptor</th>
<th>Fabric Description</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>AGGREGATED MICROFABRIC</td>
<td>Clay matrix dominant. Clay sized particles are grouped into assemblages of 20-50μm with curvilinear boundaries to which the particles may locally conform. Inter-aggregate voids are of various sizes and often exceed 20-50μm. Local, isolated bio-debris and silt grains may be present.</td>
<td>The general appearance is of clay-silt aggregates separated by voids of similar size. The aggregates have relatively small internal voids (2μm - 5μm) and are held together by clay platelets in edge to face, edge to edge or face to face contact, sometimes with visible mucal material. The inter-aggregate voids are usually elongate or curvilinear.</td>
</tr>
<tr>
<td>CUMULATE MICROFABRIC</td>
<td>The micrograph contains substantial biogenic debris (plus possible silt) embedded in clay matrix, thus forming a bio-debris cumulate. The clay matrix is predominant and may have either a uniform or an aggregated organisation.</td>
<td>The clay matrix is often uniform either without obvious aggregation or with only limited aggregates. Individual areas of clay may have a regular 'honeycomb' appearance.</td>
</tr>
<tr>
<td>GRANULAR MICROFABRIC</td>
<td>The micrograph is dominated by silt-sized particles with little biogenic debris. The clay matrix is subordinate and may have either a uniform or an aggregated organisation.</td>
<td>The silt particles are normally angular and are in contact at many points. Interparticle areas are loosely filled with clay matrix or are empty. Clay particles bridge many silt grains: mucal cements are largely absent.</td>
</tr>
</tbody>
</table>

### Table 5.6 Summary of microfabric element descriptors.

<table>
<thead>
<tr>
<th>Microfabric Element Descriptor</th>
<th>Description/Variations</th>
</tr>
</thead>
<tbody>
<tr>
<td>CLAY FRAMEWORK</td>
<td><strong>Clay Framework</strong> Normally composed of clay particles in face-face or edge-face contact, often with considerable overlap. Intra-framework voids typically 2μm - 5μm size. Sometimes has distinct structure of associated particles with curvilinear boundaries smaller than 25μm patch size.</td>
</tr>
<tr>
<td>SILT PARTICLE</td>
<td><strong>Silt particle isolate</strong> Single, dominant grain within a silt accumulation or set in a differing matrix.</td>
</tr>
<tr>
<td></td>
<td><strong>Silt particle accumulation</strong> Cumulate of silt-sized grains without matrix or quantitatively significant in comparison to matrix.</td>
</tr>
<tr>
<td>BIO-DEBRIS</td>
<td><strong>Bio-debris accumulation</strong> Cumulate of biogenic origin without matrix or quantitatively significant in comparison to matrix.</td>
</tr>
<tr>
<td></td>
<td><strong>Bio-debris isolate</strong> Item of biogenic origin within differing matrix.</td>
</tr>
</tbody>
</table>
by a relatively higher proportion of biogenic debris, held in a uniform matrix whose floc/aggregate structure has often been completely or partially lost.

The aggregated microfabric (Figure 5.17) is dominated by the clay matrix. Edge to face, edge to edge and face to face contacts between clay particles result in flocculation or aggregation of these particles on a scale of 20µm to 50µm. These flocules have curvilinear boundaries with inter-aggregate voids often exceeding 20µm to 50µm (occasionally 200µm), and intra-aggregate voids usually in the 2µm to 5µm range. This results in the formation of a very open structure sometimes with visible mucal material. Local isolated silt grains (20µm to 50µm) and bio-debris are to be found floating in the clay matrix. Being on about the same scale as the clay flocules which would behave hydraulically as whole particles, they were probably hydraulically similar and therefore deposited simultaneously. These flocules are usually thought to be formed within the water column, possibly by biogenic activity and due to electrical charges which occur on the particle surfaces (Bennet et al., 1991; Kranck, 1991; Syvitski, 1991). This microfabric dominates thicker beds of silty clay where there is an absence of silty laminae and areas which have been reworked by small-scale bioturbation.

The granular microfabric (Figure 5.18) is dominated by silt particles (both clast supported and matrix supported) and arises from simple hydraulic sorting of sediment material. There is little biogenic debris and the clay matrix is subordinate. Interparticle areas are either loosely filled with clay matrix or are empty. Clay particles bridge many silt grains and mucal cements are largely absent. This microfabric has been identified primarily in samples taken from bedding contacts, from silt-bed and silt laminae horizons.

The cumulate microfabric (Figure 5.19) contains substantial biogenic debris with minor silt embedded in a clay matrix. The clay matrix is predominant and may have either a uniform organisation without obvious aggregation or with only limited aggregation. Individual areas of clay may have a regular honeycomb appearance. Biogenic debris consists of: assorted mollusc and bivalve shell fragments, diatom frustules, foraminiferal tests and sponge spicules. This microfabric probably formed from the biological reworking of previously deposited particles or particle aggregates during periods of reduced input. It is the dominant microfabric of the Type III mottled horizons, although it is not restricted to these sections of the core. Except for very silty bedding contacts, a certain amount of the cumulate microfabric is encountered in all macrofabric types.

In addition to these principal types of fabric, it is possible to identify three additional types that are here termed minor microfabrics: the honeycomb, the fused sheet and the pelletised microfabrics. Example micrographs are shown in Figure 5.20. Although these minor fabrics are relatively

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Figure 5.17  Examples of the aggregate microfabric.  
(a) 6.23m bgl, borehole HW7;  
(b) 7.21m bgl, borehole HW3.
Figure 5.18 Examples of the granular microfabric from 4.00m bgl, borehole HW7. (a) shows an overall view of the granular fabric; (b) micrograph from the 750x magnification study. Note different scales.
Figure 5.19  Examples of the cumulate microfabric from borehole HW3.  
(a) from 9.75m bgl, (b) from 15.75m bgl. Note abundance of biodebris 
consisting of shell fragments, spicules and diatom frustules.
uncommon, attention is drawn to them since they may reflect genetic processes which have
geotechnical significance.

The *honeycomb microfabric* is a very regular example of a clay framework (Figure 5.20(a)). It has
only been found extensively at a few positions, notably in the mottled facies in HW3 at about 6m
depth. Its significance is unclear, although it may represent some aspect of biological, perhaps
bacterial, activity. On the other hand this fabric could be due partly to gas-generated microvoids.
Some of the voids have a rounder appearance where the clay platelets are bent around the
circumference of the void. It has been suggested by Wartel *et al.* (1991) that similar structures
could be gas-generated features although these are usually in the order of 0.5mm to 5mm in
diameter.

The *fused sheet microfabric* is dominated by extensive sheets of organic material, probably
polysaccharide, which coats the underlying sediment material (Figure 5.20(b)). It has only been
recorded at high levels (at about 4.95m depth and above) within the mottled Type II facies and
does not cover the whole sample. It is likely that it is a form of polysaccharide mat that has
formed due to the activities of epipelagic diatoms (Paterson *et al.*, 1990), or the mucous-lined
burrows of worms (Schäfer, 1972; Meadows & Tait 1989), and is responsible for some of the
organic materials discussed in Chapter Six.

The *pelletised microfabric* (Figure 5.20(c)) is another minor microfabric and is a well-developed
example of the aggregated structure. The pellets are very likely to have been biogenic in origin,
perhaps formed in the water column as aggregates or planktonic faecal pellets, or perhaps
formed or modified at the sea bed by ingestion and excretion of sediment by benthic feeders
(Syvitski, 1991).

5.5.3 Relative Frequencies of Principal Microfabrics

The results of the study on the ‘controlled’ set of micrographs are tabulated in Table 5.7 and
shown graphically in Figure 5.21. It is clear from inspection that (a) those samples from bed
contacts and silty mid-bed within the bedded facies are dominated by the granular microfabric;
(b) in the other samples from the bedded facies the granular microfabric is also relatively
common, with the exception of the sample from the fine mid-bed; (c) the samples from the
mottled facies possess a mix of aggregated and cumulate microfabrics: the Type III (densely,
finely mottled) samples have the highest relative proportion of the cumulate fabric, probably as a
result of a lower accumulation rate and hence more intense reworking, as implied by the greater
density of mottling; the Type II (densely mottled) sample had the highest proportion of the
aggregated fabric but more granular and less cumulate fabric than the other mottled samples,
probably as a result of a more steady and slightly increased accumulation rate.
Figure 5.20  Examples of the minor microfabric types.
(a) Honeycomb microfabric: borehole HW3, 9.75m bgl;
(b) Fused sheet microfabric: borehole HW7, 4.84m bgl;
(c) Pelletised microfabric: borehole HW7, 4.30m bgl.
Note different scales.
Table 5.7 Relative frequencies of the three principal microfabrics in each sample.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Aggregated</th>
<th>Cumulate</th>
<th>Granular</th>
</tr>
</thead>
<tbody>
<tr>
<td>SE1</td>
<td>21.0</td>
<td>21.0</td>
<td>58.0</td>
</tr>
<tr>
<td>SE2</td>
<td>15.8</td>
<td>31.6</td>
<td>52.6</td>
</tr>
<tr>
<td>SE3</td>
<td>4.20</td>
<td>12.5</td>
<td>83.3</td>
</tr>
<tr>
<td>SE4</td>
<td>4.50</td>
<td>0.00</td>
<td>95.5</td>
</tr>
<tr>
<td>SE5</td>
<td>0.00</td>
<td>0.00</td>
<td>100.0</td>
</tr>
<tr>
<td>SE6</td>
<td>80.0</td>
<td>13.3</td>
<td>6.70</td>
</tr>
<tr>
<td>SE7</td>
<td>52.9</td>
<td>47.1</td>
<td>0.00</td>
</tr>
<tr>
<td>SE8</td>
<td>33.4</td>
<td>45.8</td>
<td>20.8</td>
</tr>
<tr>
<td>SE9</td>
<td>57.1</td>
<td>14.3</td>
<td>28.6</td>
</tr>
</tbody>
</table>

Figure 5.21 Relative proportion of principal microfabric types in the samples at 750x magnification.
5.5.4 Classification and Frequency of Microfabric Elements

Examination of the controlled micrographs using the grid square approach described above suggests that, at the 25µm scale of the squares, three microfabric elements frequently recur. These have been termed the silt particle, clay framework and bio-debris elements. These classifications were originally sub-divided into the ‘variations’ of Table 5.6 but the results of the analyses were then recombined under the main microfabric element categories. Figures 5.22 to 5.26 show example micrographs of each of these elements respectively.

Table 5.8 and Figure 5.27 show the relative proportions of the microfabric elements in each of the principal microfabrics themselves. The general associations are as might be expected: in the granular microstructure the silt particle element is normally the most frequent; in the aggregated microstructure the clay framework is the more common; in the cumulate structure there is a higher proportion of the bio-debris element than in the others. Although there is much scatter on the triangular plot, and the three fabrics overlap in consequence, Table 5.9 shows that the distribution of elements within the microfabric types is statistically different from a uniform spread at the 95% significance level. Further, the individual contributions to the chi-square statistic (Table 5.9) shows that in each structural type the largest contribution comes from a different element and that in each case the element is the one expected from the above associations. This indicates that, despite the scatter on Figure 5.27, the threefold division of the overall images into the principal microstructures is statistically meaningful.

Table 5.10 and Figure 5.28 show the relative proportions of the microfabric elements in each of the samples. In general, those samples from bed contacts and silty beds are silt-dominated; the others are largely clay framework or bio-debris dominated. In terms of microfabric elements there is no clear distinction between the samples from the mottled facies and those from the finer areas of the bedded facies in this respect.

5.5.5 Descriptions of Miscellaneous Micrograph Features

During the systematic examination of the various fabrics discussed above, a number of observations were made of individual features that have some diagnostic significance or are merely interesting and thus worth reporting. This subsection describes these and suggests any conclusions that might be drawn from them.

Silt Particles: Quartz grains from fine silt size to medium sand are generally angular in shape with fresh unweathered faces and conchoidal fracture steps (Figures 5.29(a) – (c)). These grains are typical of those having undergone glacial fracture (Krinsley & Donahue, 1968; Eyles, 1978; Bull, 1978; Mahaney, 1995; Mahaney & Kalm, 1995) and can therefore be used as provenance indicators. It is, of course, possible to achieve similar surface textures in non-glacial sedimentary
Figure 5.22  Examples of the silt particle microfabric elements highlighted for illustration. (a) and (b) show both the 'silt particle accumulate' (SPA) and the 'silt particle isolate' (SPI) elements (V = void). Sample from 4.00m bgl, borehole HW7. Micrographs are from the 750x magnification study and show superimposed grid used in this study. Grid squares = 25μm.
Figure 5.23  Examples of the silt particle isolate (SPI) microfabric element in a clay-rich matrix. Sample from 6.30m bgl, borehole HW7. Grid as for Figure 5.22.
Figure 5.24  Examples of the clay framework (CF) microfabric element.  Sample from 7.21 m bgl, borehole HW3.  Grid as for Figure 5.22.
Figure 5.25  Examples of the bio-debris accumulate (BA) microfabric element. Sample from 7.21m bgl, borehole HW3. Grid as for Figure 5.22.
Figure 5.26 Examples of the bio-debris isolate (Bl) microfabric element.
(a) Sample from 6.32m bgl, borehole HW7;
(b) Sample from 7.21m bgl, borehole HW3. Grid as for Figure 5.22.
Table 5.8  Relative frequencies (%) of elements within principal microfabrics.

<table>
<thead>
<tr>
<th></th>
<th>Clay Framework</th>
<th>Silt Particle</th>
<th>Bio-debris</th>
<th>Totals</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aggregated</td>
<td>15.0</td>
<td>6.7</td>
<td>5.9</td>
<td>27.6</td>
</tr>
<tr>
<td>Cumulate</td>
<td>6.6</td>
<td>6.9</td>
<td>8.0</td>
<td>21.5</td>
</tr>
<tr>
<td>Granular</td>
<td>10.2</td>
<td>32.4</td>
<td>8.3</td>
<td>50.9</td>
</tr>
<tr>
<td><strong>Totals</strong></td>
<td><strong>31.8</strong></td>
<td><strong>46.0</strong></td>
<td><strong>22.2</strong></td>
<td><strong>100</strong></td>
</tr>
</tbody>
</table>

Table 5.9  Chi-square values for departures from uniform distribution.

<table>
<thead>
<tr>
<th></th>
<th>Clay Framework</th>
<th>Silt Particle</th>
<th>Bio-debris</th>
<th>Totals</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aggregated</td>
<td>4.48</td>
<td>2.86</td>
<td>0.01</td>
<td>7.35</td>
</tr>
<tr>
<td>Cumulate</td>
<td>0.01</td>
<td>1.29</td>
<td>2.26</td>
<td>3.56</td>
</tr>
<tr>
<td>Granular</td>
<td>2.22</td>
<td>3.47</td>
<td>0.80</td>
<td>6.49</td>
</tr>
<tr>
<td><strong>Totals</strong></td>
<td><strong>6.71</strong></td>
<td><strong>7.62</strong></td>
<td><strong>3.07</strong></td>
<td><strong>17.40</strong></td>
</tr>
</tbody>
</table>

* The chi-square value of 17.40 is significant at the 95% confidence level.

Figure 5.27  Relative proportion of microfabric elements within principal microfabric types; at 750x magnification.
Table 5.10  Relative frequencies of microfabric elements in each sample.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Clay Framework</th>
<th>Silt Particle</th>
<th>Bio-debris</th>
</tr>
</thead>
<tbody>
<tr>
<td>SE1</td>
<td>Bedded: Thickly laminated (1)</td>
<td>17.2</td>
<td>49.8</td>
</tr>
<tr>
<td>SE2</td>
<td>Bedded: Thickly laminated (2)</td>
<td>15.9</td>
<td>48.6</td>
</tr>
<tr>
<td>SE3</td>
<td>Bedded: Mid-bed silty</td>
<td>26.7</td>
<td>56.4</td>
</tr>
<tr>
<td>SE4</td>
<td>Bedded: Non-erosive contact</td>
<td>20.5</td>
<td>63.9</td>
</tr>
<tr>
<td>SE5</td>
<td>Bedded: Penetrative contact</td>
<td>11.3</td>
<td>83.1</td>
</tr>
<tr>
<td>SE6</td>
<td>Bedded: Mid-bed fine</td>
<td>52.4</td>
<td>24.9</td>
</tr>
<tr>
<td>SE7</td>
<td>Mottled: Type III dense (1)</td>
<td>56.0</td>
<td>16.8</td>
</tr>
<tr>
<td>SE8</td>
<td>Mottled: Type III dense (2)</td>
<td>43.9</td>
<td>24.4</td>
</tr>
<tr>
<td>SE9</td>
<td>Mottled: Type II dense</td>
<td>46.0</td>
<td>40.0</td>
</tr>
</tbody>
</table>

Figure 5.28  Relative proportion of microfabric elements in the samples at 750x magnification.
environments characterised by high energy transport (Byles, 1978; Mahaney & Kalm, 2000), but the presence of quartz rock flour in the clay fraction of these sediments offers support for the suggestion that these sediments have undergone comminution during glacial erosion. Figure 5.29(a) also shows a heavily corroded orthoclase feldspar (F). Feldspar grains are generally more susceptible to chemical corrosion than quartz (due to the selective loss of potassium) and often show evidence of weathering and chemical dissolution while quartz grains remain unaffected.

Although quartz was found to be the dominant mineral in the silt fraction, there were a number of sheet minerals in this size range. Figure 5.29(d) and (e) show medium and coarse silt-sized grains respectively: they are delaminating sheet minerals, possibly kaolinite, muscovite or biotite. A pyrite framboid can be seen at lower left in (d).

Clay particles and structures: Individual clay platelets are seen in Figure 5.30(a). It can be seen that the outlines of the platelets are often irregular although some equant clay minerals are visible on the micrograph. Figures 5.30(b) and (c) (from 9.75m in HW3 and 5.25m in HW7 respectively) show the edge-edge and edge-face arrangement of the clay minerals, probably the result of flocculation, which has resulted in the very obvious open structure that is reflected in the high water content at this level in the core.

Pellets: A number of oblong structures were found which were thought to be faecal pellets. They were of the order of 100µm in length by about 25µm in diameter and consisted of aggregates of clay-size platelets and fine silt grains. Figure 5.31(a) and (b) shows two such pellets embedded into a clay matrix from borehole HW3 at 7.21m depth. The pellet in Figure 5.31(b) shows the end of the pellet partly broken off exposing the internal structure. The pellet in Figure 5.31(c) is from the upper bedded-burrowed section of borehole HW7 at 2.43m and appears to contain more silt than those in (a) and (b). This is to be expected in the siltier horizon at 2.43m and this pellet is probably an end-on view of a faecal pellet from a different type of organism than that in (a) and (b).

Biogenic debris: A wide variety of biogenic material was found to be disseminated throughout all of the microfabric types (Table 5.7). There are a number of possible sources for this material which has been incorporated into the sediments: zoo- and phytoplankton from the photic zone of the water column (planktonic diatoms) rain down onto the sediment surface when the organisms die; those which live on and within the sediment surface (epipelagic diatoms, foraminifera, echinoderms, molluscs).

Figure 5.32 shows examples of biogenic debris including a number of diatom frustules and a foraminiferal test. An attempt has been made to identify these, with varying success (Table 5.11). A systematic micropalaeontological study is not part of this thesis, but one point worthy of note
Figure 5.29  Nature of silt-size grains. Q = quartz, F = feldspar, P = pyrites. Arrows point to fractured surface.
(a) Quartz grain with fractured surface and corroded feldspar grain. Scale bar = 20µm.
(b) Angular quartz grain with fracture surface and 
(c) closer view of fracture surface, marked with box on (b). Scale bar = 5µm.
(d) Medium silt-sized and (e) coarse silt-sized delaminating layered minerals. 
Scale bars = 10µm and 20µm respectively.
Figure 5.30  Micrographs of clay platelets and structures. See text for explanation.
Figure 5.31  (a) and (b) Oblong faecal pellets from the mottled Type III facies at 7.21 m in HW3. (c) Pellet from the bedded burrowed facies at 2.43 m in HW7.
is that diatom (j) (*Pleurosigma* sp.) has been identified as an epipelagic diatom which inhabits brackish waters. This supports the palaeo-water depth interpretation in Chapter Eight which suggests that the sediments above about 4.0m depth were laid down under intertidal conditions. The other diatoms in Table 5.11 were sampled at 7.21m and below, and have been identified as marine or coastal marine genera or species. These organisms probably experienced subtidal conditions. A number of diatom tests and foraminiferal frustules had undergone secondary mineralisation and were found to have infills of pyrite frambooids: illustrations of these are shown in Chapter Six (cf. Figure 6.8).

A greater variety of biogenic debris can be seen in the cumulative microfabric micrographs in Figure 5.19.

*Biogenic structures:* Figure 5.33(a) shows a hollow, vertical structure measuring ~10μm in diameter and made up of clay platelets in edge to edge and edge to face contact. It appeared at the top of a mottled zone where mottling was types II/III at about 7.05m bgl. The origin is unknown but it could be either a small burrow structure or due to the generation of gas.

Figure 5.33(b) shows a more complex biogenic structure (9.75m bgl in HW3) from about 150mm above the base of a mottled unit (types I/II mottles) which extended over a total depth of about 500mm. It consists of main circular structure of about 20μm to 30μm diameter, to which are attached subsidiary tubular structures about 5μm to 10μm in diameter (internal diameter 2μm to 3μm) and about 10μm to 20μm long (Figures 5.33 (c) and (d)). The main circular structure has a central void of about 10μm in diameter and the whole structure is made up of microcrystals from about 100nm to 2μm in size. These crystals are generally angular, often tabular and randomly orientated, and appear to be cemented together. A membrane-like material can be seen at the mouth of one of the subsidiary tubular structures (Figures 5.33(c) and (d)) and appears to form a lining on the inside of the tube. The whole structure has an overall spiral appearance. It may be a burrow (probably feeding) of a very small organism, or it may be part of a larger structure which is not discernible.

5.6 **INTERPRETATION AND DISCUSSION**

The aim of this discussion is to identify and explain those aspects of the sediment fabric and facies architecture which are likely to be significant for the geotechnical properties of the clay soil at Bothkennar. These will be pursued in Chapter Seven where the new geotechnical results will be presented and the associated property profiles discussed. The present section thus proceeds in two stages. First, the individual macroscopic facies are each considered and their likely environment of deposition is established and, second, the evidence of the facies succession is
Table 5.11 Identification and preferred habitats of diatom frustules shown in Figure 5.32.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Borehole</th>
<th>Depth (m)</th>
<th>Identification</th>
<th>Habitat</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>HW3</td>
<td>17.80</td>
<td><em>Paralia sulcata</em></td>
<td>Commonly found in marine inshore plankton.</td>
</tr>
<tr>
<td>b</td>
<td>HW3</td>
<td>7.21</td>
<td><em>Paralia sulcata</em></td>
<td>As above.</td>
</tr>
<tr>
<td>c</td>
<td>HW3</td>
<td>14.06</td>
<td><em>Paralia sulcata</em></td>
<td>As above.</td>
</tr>
<tr>
<td>d</td>
<td>HW3</td>
<td>9.75</td>
<td><em>Skeletonema sp.</em></td>
<td>Found in coastal marine plankton.</td>
</tr>
<tr>
<td>e</td>
<td>HW3</td>
<td>11.20</td>
<td><em>Coscinodiscus sp.</em></td>
<td>Free-living, marine, often common in the phytoplankton.</td>
</tr>
<tr>
<td>f</td>
<td>HW3</td>
<td>7.21</td>
<td><em>Thalassiosira sp.</em></td>
<td>Mainly in the marine plankton, solitary or joined by threads to form loose chains; or in mucilage masses.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>or <em>Minidiscus sp.</em></td>
<td></td>
</tr>
<tr>
<td>g</td>
<td>HW3</td>
<td>17.80</td>
<td><em>Actinocyclus senarius</em></td>
<td>Common in neritic assemblages, a component of loose material or attached to other algae on coastal sediments.</td>
</tr>
<tr>
<td>h</td>
<td>HW3</td>
<td>17.05</td>
<td><em>Actinocyclus senarius</em></td>
<td>As above.</td>
</tr>
<tr>
<td>i</td>
<td>HW3</td>
<td>11.20</td>
<td><em>Hyalodiscus laevis</em></td>
<td>A common marine genus often found attached to seaweed by mucilage pads; also found in turbulent inshore waters or lying on the sediment surface.</td>
</tr>
<tr>
<td>j</td>
<td>HW7</td>
<td>3.43</td>
<td><em>Pleurosigma sp.</em></td>
<td>Found in marine and brackish waters, usually epipellic on sand or silt but occasionally planktonic.</td>
</tr>
</tbody>
</table>

1 These refer to Figure 5.32 (a) to (j).
2 Round *et al.* (1990), page 165.
3 Round *et al.* (1990), page 140.
4 Round *et al.* (1990), page 176.
5 Round *et al.* (1990), page 132.
6 Round *et al.* (1990), page 138. Identification of (f) not certain.
7 Round *et al.* (1990), page 200.
8 Hendey (1964), plate XXIII, 1.
9 Round *et al.* (1990), page 162.
10 Round *et al.* (1990), page 580.
11 Hendey (1964), plate XXXV.
Figure 5.33 Examples of biogenic structures. (a) Hollow vertical burrow or gas generated void. (b) Complex biogenic structure, arrow points to tubular structure which is enlarged in (c) and (d). Membrane-like material (M) in (c) and (d).
used to identify possible events of geotechnical significance during the depositional period as a whole.

5.6.1 Bedded Facies

The major part of the bedded facies is believed to be the product of sedimentation under quiet, tidally dominated conditions such as those in the Forth estuary at the present day. Individual beds consist of a lower, silty layer largely composed of individual mineral grains, which progressively gives way to a hydraulically similar layer of less dense clay particle aggregates. The slightly coarser sediments at the base of the Claret Formation suggest a succession laid down in predominantly shallow, offshore marine conditions that suffered greater current activity than did the later deposits.

Near the top of the Claret Formation sediments of the bedded facies occurs only infrequently in borehole HW3 (Figure 5.1), but have a thickness of about 1.0m immediately below the shell bed in borehole HW7. The stratigraphically highest sediments in this facies (above about -1m OD) show evidence of very shallow subtidal to intertidal conditions. This includes rip-up clay flakes and layers of pellets (probably biogenic) which are a few millimetres in size and locally abundant, occurring along distinct horizons above sharp, cross-cutting, erosive boundaries. These distinctive features are not seen in the bedded units lower in the succession and suggest possible subaerial exposure and current conditions sufficient to cause local re-erosion of the sediments. Other evidence comes from the presence of burrows (sections 5.2.2 and 5.4.1) probably associated with low intertidal to subtidal species and the completion of the sequence by an erosion surface colonised by intertidal species such as *Cerastoderma edule*.

These findings are in agreement with the water depth model to be (described in Chapter Eight) which suggests the onset of intertidal conditions from around this depth in the succession. Thus the overall interpretation is that the bedded facies formed under initially subtidal conditions, during which wave and current activity were of limited significance. Deposition continued under intertidal conditions during which wave and current activity increased in importance causing some re-erosion of a possibly desiccated surface. The sediment was eventually colonised by (low) intertidal species when deposition had finally ceased.

5.6.2 Mottled Facies

The mottled facies is believed to be the equivalent of the bedded facies now disturbed by bioturbation. The mottling is considered to be the result of burrowing, probably by polychaetes and oligochaetes, both of which groups are known to be common in the Forth mudflats at the present day (section 5.4.2). Although polychaete burrows can vary greatly in shape, size and configuration, they often have smooth walls with a mucal lining (Howard & Frey, 1973). The
central tubes which have occasionally been observed in some of the mottles also have similar smooth walls and mucal coatings (Figure 5.8). The presence of faecal pellets (Figure 5.31) within the mottled facies is a further indication of biogenic activity. Although it has been suggested (Hawkins et al., 1991) that the mottles were caused by plant root systems, this explanation appears less likely in view of the water depths involved (below) and the general absence of plant material associated with the mottles (section 5.4.2).

The succession of the subfacies through Type I to Type III mottles is believed to reflect the depth structure of the infaunal community responsible for the mottling. Thus the larger sporadic Type I mottles represent organisms able to exist in larger, well irrigated, burrows below the oxygenated surface layer, whereas the smallest, dense, mainly horizontal (Type III) mottles represent organisms in very small burrows and thus confined to the near surface layers. The bulk of the organisms (responsible for Type II mottles) exist at intermediate depth. If the community is disturbed by an influx of sediment, it becomes re-established at a higher level, so creating an apparent cycle of mottling.

5.6.3 Laminated Facies

The finely laminated silt to clay alternation, together with the symmetrical ripples and lenticular form of the beds, suggests that the sediments of the laminated facies were deposited in a tidal setting involving transitions from bedload to sediment load transport during the tidal cycle (Reineck & Wunderlich, 1968; Klein, 1971). The erosive contacts and apparently linear subsurface form of the unit suggest the fill of a tidal channel analogous to modern clay counterparts in close proximity.

The transition between the top of the laminated beds and the mottled facies is gradational over about a 10cm section. As tidal channels are not permanent features and channel switching would be expected over time, this would account for the resumption of mottled deposits which overlie the laminated beds.

5.6.4 Interpretation of the SEM Microfabrics

The origins of the three principal microfabrics can be considered in terms of the processes involved in their formation.

The aggregated microfabric is most likely to be the result of normal water column processes (Bennet et al., 1991). These include flocculation and pelletisation. During flocculation, clay platelets undergo mutual attraction in saline and brackish waters and are held together by van der Waals forces. When they settle out of the water column they form an open structure with interaggregate voids on a similar scale to the size of the aggregates themselves. Aggregates also
form within the water column when minerogenic particles are trapped by mucous excreted by phytoplankton. Pelletisation is caused within the water column by the ingestion of particles by filter feeding organisms which often results in the binding together of these particles by sticky mucous. This process can also occur within the sediment since burrowing polychaetes, oligochaetes and some molluscs ingest the sediments as part of their feeding processes.

Although the clay framework microfabric element dominated in samples SE6 to SE9, the aggregated microfabric which was calculated as 80% in SE6 (bedded facies: mid-bed fine) (Table 5.5) has been reduced in the samples from the mottled facies (SE7 – SE9), probably due to reworking of the sediment fabric during bioturbation. Although the open clay framework often still remains, the frequency of interaggregate voids are reduced. The fused sheet microfabric is an extreme example of this process where bioturbation appears to collapse the clay framework and the clay particles become smeared with mucous.

The granular microfabric is more frequent in silty horizons, particularly at silty contacts which is to be expected. The reduced clay particle content appears to be the result of winnowing by tidal currents.

In the cumulate microfabric a relatively high concentration of biogenic debris is incorporated into the clay matrix. This can be attributed to a reduced rate of sedimentation which leads to accumulation and reworking of biogenic material from phytoplankton from within the water column and organisms which live within and on the sediment surface. This is attributed to a condensed sequence which is indicative of a hiatus in the sediment profile: the highest frequencies of the cumulate microfabric occurred within samples SE7 and SE8 from the mottled Type III facies (Table 5.7), which usually occurred at the top of a sediment sequence and was seen to be truncated by a silty parting or lamina.

The microfabric obviously plays a role in the geotechnical properties of the sediments. This is explored in detail in Chapter Eight (section 8.4.1). Although the data are not sufficient to draw definite conclusions, amongst the suggestions made in that chapter will be that the relative dominance of the three microfabrics is a consequence of the sedimentation history of the deposit and that this is reflected in very small scale in the profiles of a number of properties, in particular that of void index, shear strength and water content.

5.7 SUMMARY
The sediments of the Claret Formation can be divided into three principal macrofacies: the bedded facies, the mottled facies and the laminated facies on the basis of primary sedimentary structures and the nature and extent of bioturbation. The bedded facies is believed to be the product of primary sedimentation under subtidal to intertidal conditions, although there exists a
distinctive subfacies, restricted to the top of the sequence, which is believed to be the product of immediately subtidal to intertidal conditions. The mottled facies is a bioturbated equivalent of the bedded facies and intermediate forms are common. The sediments of the laminated facies occur as the infill to eroded channels within the other sediments and are interpreted as deposits formed within tidal channels cut into the main body of the Claret Formation.

Biogenic features occur throughout all the facies. In addition to the mottling, a variety of burrow structures are also found, ranging from larger shafts, funnel- and U-shaped burrows down to fine anastomosing structures. Comparison of the various fossil features with the present invertebrate fauna of the Forth mudflats suggests many of the organisms possibly responsible for them.

Examination of the sediments under the SEM has allowed the microstructure to be categorised into three types, termed the aggregated, the granular and the cumulate microfabrics. They are considered respectively to have originated by the accumulation of pelletised/flocculated particles; by the selective winnowing of fines to leave a siltier unit; and by the biological reworking of the sediment under conditions of reduced sedimentation. It is suggested that the aggregated microfabric is relatively more frequent in the bedded facies away from silty partings and in the mottled facies; the granular microfabric is relatively more frequent in the bedded facies; the cumulate microfabric is less frequent overall but is most common in the mottled facies.
CHAPTER SIX

PROPERTIES OF THE CLARET FORMATION AT bothkennar
PART 2: SEDIMENT COMPOSITION

6.1 INTRODUCTION

6.2 MINERALOGICAL COMPOSITION

6.3 PARTICLE SIZE DISTRIBUTION

6.4 ORGANIC MATERIAL

6.5 INORGANIC GEOCHEMISTRY

6.6 CONCLUSIONS
CHAPTER SIX

PROPERTIES OF THE CLARET FORMATION AT BOTHKENNAR

PART 2: SEDIMENT COMPOSITION

6.1 INTRODUCTION

This chapter describes the composition of the material that forms the sediments of the Claret Formation at Bothkennar. Mineralogically they are a rather complex suite, derived ultimately by glacial action on the metamorphic rocks of the southern Highlands and the sediments and volcanic rocks of the Midland Valley (cf. Chapter Two). Their particle size distribution, although much modified by Holocene depositional processes, also reflects this glaciogenic derivation. They contain organic material input from the estuary in which they accumulated and have a pore water composition that in part reflects modern estuarine conditions and in part also reflects presumed interstitial processes.

6.2 MINERALOGICAL COMPOSITION

The bulk mineral composition was determined at 32 positions in boreholes HW3 and HW7 (approximately one metre depth spacing or less) using the X-ray powder diffraction (XRD) technique described in Chapter Four. Analyses were also carried out on the silt- and clay-sized fractions separately, in order to determine the distribution of the minerals between the size fractions. Additionally, in order to investigate the clay mineralogy in more detail, samples from the clay-sized fraction were analysed after air drying at 20°C, after furnace drying at 550°C and after glycerolation.

The main series of diffraction traces was obtained on powdered oven-dried material, giving a presumed random orientation. This typically produced a trace showing around sixty peaks (up to 2θ angle = 60°), of which only one quarter are strong or very strong1. About one-half are weak or very weak. The minerals were identified principally from their d-spacings and relative peak intensities, using tables of natural or synthetic minerals (Borg & Smith, 1969; Brindley & Brown, 1980). From these tables candidate minerals were selected and the principal reflection planes identified. An identification was considered acceptable if it was possible to distinguish on the

1 The strongest peak is always that at 26.7° (2θ) [d=3.33Å] due to the 101 reflection from low quartz. This is typically around five times stronger than the next strongest reflection and has been used to characterise the relative strength of the other reflections. Normally in this work a peak has been considered significant only if its strength is more than 1% of this quartz peak. Due to the variability of the relative intensities, in the following Tables peaks have been classified merely as very strong (v), strong (s), moderate (m) or weak (w) if they are approximately >10%, 5%-10%, 2%-4.99% or 1%-1.99% of the strongest quartz peak respectively. A peak is described as very weak (vw) if it is <1% of the quartz peak (1% ±2 times the intensity of the background peaks); some such peaks are consistent features of the diffraction traces and can be identified from tables, whereas others occur only sporadically.
trace an integral series of peaks arising from a known family of crystallographic planes within that mineral. The minerals so identified form a consistent suite that is seen on all the diffraction traces and it will be shown that this agrees with the general mineralogy of the likely bedrock source. Other peaks, which are only seen sporadically, are suggestive of particular minerals but do not confirm their presence.

A series of representative diffraction traces is reproduced in Figure 6.1. Disregarding for the moment the identification of the individual peaks, they show that the mineral assemblage is of uniform composition throughout the full depth of the Claret Formation at Bothkennar. This conclusion is maintained when all 32 traces are compared. However, the relative strengths of the peaks vary, which is thought to be due in part (with some qualifications discussed below) to variations in the relative proportions of the minerals present in one sample against another. This may be due to small but significant changes in the relative proportion of sizes within the silt-sized fraction of the sediment and to variations in the silt:clay ratio (cf. section 6.3.2 below).

Examination of Figure 6.1 shows that the peaks vary in both width and shape: some are asymmetric and some have distinct 'shoulders'. These features arise mainly from the large number of overlapping reflections that are present, and from the layered structure of some of the minerals. In general, the strength of the reflections is governed by the proportion of the minerals present, the atomic structure on the reflection plane and the cumulative effect of overlapping peaks. Thus, some care must be taken when identifying mineral families and varieties: some of the particular problems that have arisen are discussed below.

Figure 6.2 shows the results obtained on a sample from 2.43m depth in borehole HW7. This is considered to be a typical diffraction trace and will be used to illustrate the detailed identification of the minerals in the suite. Inter-comparison of the whole set of diffraction traces shows minor variations in the calculated d-spacings (possibly due to instrumental error: cf. section 6.2.1) and some variation in the peak strengths relative to the main quartz peak. The first problem can be addressed by using the quartz peaks as an internal reference (section 6.2.1) and the second by the broad categorisation of peaks as strong, moderate etc. Comparison also shows that there are often some peaks which occur sporadically on the different diffraction traces: it is thought that these arise from relatively less common minerals not present in all samples or are chance observations of those reflections from common minerals that are normally obscured by other peaks.

Sixty-one individual peaks are noted on Figure 6.2 and most are assigned to a mineral family in Table 6.1. The peaks are numbered for convenience using a set of identification numbers by which individual peaks will be referenced throughout this section of the thesis. Comparison of
Figure 6.1 Selection of X-ray diffraction traces at the BSCRS illustrating the uniformity of mineralogical composition of the Claret Formation.
Figure 6.2  Typical X-ray diffraction trace used to identify peaks. Numbered peaks are identified in Table 6.1.
Table 6.1 XRD Analysis of whole sample: assignment of peaks to a mineral family (cf. Figure 6.2).

<table>
<thead>
<tr>
<th>Peak no.</th>
<th>Angle (2θ)</th>
<th>d-Spacing (Å)</th>
<th>Corrected d-Spacing (Å)</th>
<th>Relative Intensity</th>
<th>Probable Mineral Family</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>6.36</td>
<td>13.886</td>
<td>13.897</td>
<td>1.86 w</td>
<td>Chlorite</td>
</tr>
<tr>
<td>3</td>
<td>12.46</td>
<td>7.097</td>
<td>7.103</td>
<td>3.59 m</td>
<td>Kaolinite</td>
</tr>
<tr>
<td>4</td>
<td>12.64</td>
<td>6.999</td>
<td>7.005</td>
<td>5.26 s</td>
<td>Chlorite</td>
</tr>
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<td>5</td>
<td>13.98</td>
<td>6.332</td>
<td>6.337</td>
<td>1.02 w</td>
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<td>6</td>
<td>17.92</td>
<td>4.947</td>
<td>4.951</td>
<td>4.31 m</td>
<td>Mica</td>
</tr>
<tr>
<td>7</td>
<td>18.94</td>
<td>4.682</td>
<td>4.686</td>
<td>1.63 w</td>
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</tr>
<tr>
<td>8</td>
<td>19.91</td>
<td>4.457</td>
<td>4.461</td>
<td>2.05 m</td>
<td>Mica/Kaolinite</td>
</tr>
<tr>
<td>9</td>
<td>20.97</td>
<td>4.234</td>
<td>4.237</td>
<td>15.64 vs</td>
<td>Quartz</td>
</tr>
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<td>10</td>
<td>22.15</td>
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<td>4.014</td>
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<td>3.942</td>
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</tr>
<tr>
<td>12</td>
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<td>3.645</td>
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<td>16</td>
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<td>3.183</td>
<td>21.71 vs</td>
<td>Feldspar/Pyroxene</td>
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<td>2.630</td>
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<th>d-Spacing (Å)</th>
<th>Corrected d-Spacing (Å)</th>
<th>Relative Intensity</th>
<th>Probable Mineral Family</th>
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<td>2.592</td>
<td>2.05 m</td>
<td>Chlorite</td>
</tr>
<tr>
<td>33</td>
<td>35.09</td>
<td>2.555</td>
<td>2.557</td>
<td>3.80 m</td>
<td>Mica/Kaolinite</td>
</tr>
<tr>
<td>34</td>
<td>35.39</td>
<td>2.535</td>
<td>2.537</td>
<td>2.78 m</td>
<td>Olivine/Pyroxene/Chlorite</td>
</tr>
<tr>
<td>35</td>
<td>36.13</td>
<td>2.484</td>
<td>2.486</td>
<td>2.14 m</td>
<td>Kaolinite/Olivine</td>
</tr>
<tr>
<td>36</td>
<td>36.65</td>
<td>2.450</td>
<td>2.452</td>
<td>10.11 vs</td>
<td>Quartz/Mica</td>
</tr>
<tr>
<td>37</td>
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<td>2.418</td>
<td>2.425</td>
<td>2.94 m</td>
<td>Pyrite</td>
</tr>
<tr>
<td>38</td>
<td>37.84</td>
<td>2.376</td>
<td>2.378</td>
<td>1.84 w</td>
<td>Chlorite/Kaolinite/Mica</td>
</tr>
<tr>
<td>39</td>
<td>38.62</td>
<td>2.330</td>
<td>2.332</td>
<td>0.90 vw</td>
<td>Kaolinite</td>
</tr>
<tr>
<td>40</td>
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<td>2.276</td>
<td>2.278</td>
<td>7.43 s</td>
<td>Quartz</td>
</tr>
<tr>
<td>41</td>
<td>40.38</td>
<td>2.232</td>
<td>2.234</td>
<td>4.31 m</td>
<td>Quartz</td>
</tr>
<tr>
<td>42</td>
<td>40.82</td>
<td>2.209</td>
<td>2.211</td>
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<td>Pyrite</td>
</tr>
<tr>
<td>43</td>
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<td>2.160</td>
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</tr>
<tr>
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<td>2.125</td>
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<td>45</td>
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<td>2.012</td>
<td>2.014</td>
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</tr>
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<td>1.991</td>
<td>5.52 s</td>
<td>Mica</td>
</tr>
<tr>
<td>47</td>
<td>45.88</td>
<td>1.976</td>
<td>1.978</td>
<td>4.95 m</td>
<td>Quartz</td>
</tr>
<tr>
<td>48</td>
<td>47.50</td>
<td>1.913</td>
<td>1.915</td>
<td>1.92 w</td>
<td>Pyrite</td>
</tr>
<tr>
<td>49</td>
<td>48.21</td>
<td>1.886</td>
<td>1.888</td>
<td>0.57 vw</td>
<td>Chlorite</td>
</tr>
<tr>
<td>50</td>
<td>49.79</td>
<td>1.830</td>
<td>1.831</td>
<td>7.01 s</td>
<td>Not assigned</td>
</tr>
<tr>
<td>51</td>
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<td>1.829</td>
<td>1.830</td>
<td>4.60 m</td>
<td>Not assigned</td>
</tr>
<tr>
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<td>1.815</td>
<td>1.816</td>
<td>16.45 vs</td>
<td>Quartz</td>
</tr>
<tr>
<td>53</td>
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<td>1.801</td>
<td>1.802</td>
<td>2.40 m</td>
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</tr>
<tr>
<td>54</td>
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<td>1.798</td>
<td>1.799</td>
<td>2.01 m</td>
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</tr>
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<td>55</td>
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<td>1.782</td>
<td>0.64 vw</td>
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</tr>
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<td>1.739</td>
<td>2.45 m</td>
<td>Not assigned</td>
</tr>
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<td>57</td>
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<td>1.671</td>
<td>7.05 s</td>
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<td>1.671</td>
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<td>Mica</td>
</tr>
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<td>1.657</td>
<td>1.658</td>
<td>2.60 m</td>
<td>Quartz</td>
</tr>
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<td>60</td>
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<td>1.633</td>
<td>1.634</td>
<td>4.24 m</td>
<td>Pyrite</td>
</tr>
<tr>
<td>61</td>
<td>59.07</td>
<td>1.563</td>
<td>1.564</td>
<td>0.83 vw</td>
<td>Pyrite</td>
</tr>
</tbody>
</table>
the observed diffraction angles of these peaks (converted to d-spacings) with published tables (Bailey, 1980; Brown, 1980; Brown & Brindley 1980) shows that the peaks fall into five groups:

- The standard series of (low) quartz reflections;
- Several integral reflection series from the basal planes of the layered minerals biotite, muscovite (illite), chlorite and kaolinite, together with some of their stronger non-basal reflections;
- Characteristic groups of reflections arising from the feldspar family, distinguished only as plagioclase or alkali feldspars;
- Sporadic reflections from various ferromagnesian minerals, notably pyroxenes;
- Sporadic reflections from non-silicate minerals, principally oxides and pyrite, mainly visible at diffraction angles >40°.

The following sections examine the XRD data for each of the above mineral groups in more detail.

The identifications have been corroborated by other means whenever possible. Visual examination of the sand and coarse silt fraction has shown quartz, biotite, muscovite, feldspar and occasional dark (ferromagnesian) minerals. Particle morphology has been observed under the SEM and, in combination with EDX² analysis, has suggested all the above minerals and such oxides as haematite, ilmenite and rutile. The presence of sulphides has also been confirmed: pyrite framboïds have been viewed under the SEM and the coloration of the sediments (black, oxidising to light grey) suggests the presence of disseminated iron monosulphide.

6.2.1 Quartz

The principal mineral in the Claret Formation sediments at Bothkennar is quartz. This is suggested both by the relative strengths of its reflections, which dominate all the diffraction traces, and by simple visual observation of the coarser particles. This finding is not unexpected in view both of the abundance of quartz in the presumed parent rocks and its mechanical and chemical resistance to degradation.

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² It must be remembered that there can be interference from surrounding particles, particularly those lying beneath thin, platey minerals (cf. Chapter Four: section 4.4.1.2) and amorphous iron oxide coatings. Therefore, the observed EDX spectrum often contains extraneous peaks. This is also evident in the standard EDX spectra of Welton (1984). EDX analyses must therefore be used in combination with other methods of mineral identification.
Figure 6.3 illustrates an example of a quartz grain seen under the SEM and also shows the corresponding element composition determined by energy-dispersive X-ray analysis (EDX). The silicon peak is obviously the dominant peak on the EDX graph: indicative of quartz. Oxygen (atomic number 8) does not appear as only elements with atomic number >16 (Na, sodium) yield a peak on the EDX spectrum. The gold (Au) peak is generated by the gold coating used during sample preparation.

Table 6.2 details those peaks from Table 6.1 which have been assigned to quartz on the basis of their relative strength and comparison with the published d-spacings (Brown, 1980: table 6.6) of quartz species. The observed strengths and spacings most closely match those of low quartz. However, there is a small inconsistency throughout Table 6.2 between the measured and the 'standard' spacings which, in view of the almost invariant values for the quartz lattice, allows an estimate to be made of the error in the diffraction trace as a whole. It can be seen that at a diffraction 2θ angle of 20.97° the measured d-spacing is 0.026Å below the standard value and that this difference reduces to 0.002Å at an angle of 50.22° or greater. This suggests that, for this sample, at 20.97° the measured diffractometer angle exceeds the true value by +0.14° and that the error reduces to +0.01° over the interval in question.

The range of errors over this interval for a set of six untreated XRD samples (analysed on the same day) is shown in Figure 6.4. It can be seen that for d-spacings in the range 2Å to 3.5Å the correction does not exceed 0.012Å, which for routine identification is unimportant. At larger d-spacings the graph indicates a rapidly growing correction: although this may be unrealistic it is undetermined and so the possibility exists that higher d-spacings (e.g. the basal spacing of chlorite or the micas) may be underestimated (perhaps by 0.1 Å to 0.2 Å) if the raw diffractometer results are used without correction. However, this does not prevent the identification of these minerals at the family level. The sample from 2.43m is represented by blue triangles on Figure 6.4 and shows consistently higher errors than the other samples. Despite this it is possible to identify the minerals at the family level and the fact that this was possible using the sample with the greatest d-spacing errors suggests that the results can be used with confidence.

6.2.2 Layered Minerals

This category includes the micas (principally muscovite and illite, biotite and phlogopite), kaolinite and chlorite. Table 6.3 lists those peaks that are believed to arise from the mica minerals. These show an integral series of basal reflections, arising from a basal spacing of around 10Å, and a number of non-basal reflections. This indicates that the dioctahedral (muscovite/illite) micas are probably present in both the 1M and 2M, structural states and that the trioctahedral (biotite/phlogopite) micas are in the 1M state. It is probable that in the
Figure 6.3  (a) Electron micrograph and (b) EDX spectrum for quartz [Si] acquired from the point marked by white O on the micrograph. The gold (Au) peaks are the result of the conductive coating applied to the SEM sample.
Table 6.2  Detailed XRD Analysis (2.43m HW7) - Quartz

<table>
<thead>
<tr>
<th>Peak no.</th>
<th>Angle (2θ)</th>
<th>Observed d-Spacing (A)</th>
<th>Standard Spacing(^1) (A)</th>
<th>Relative Intensity</th>
<th>Reflection plane</th>
</tr>
</thead>
<tbody>
<tr>
<td>9</td>
<td>20.97</td>
<td>4.234</td>
<td>4.260</td>
<td>vs</td>
<td>100</td>
</tr>
<tr>
<td>18</td>
<td>26.74</td>
<td>3.331</td>
<td>3.343</td>
<td>100%</td>
<td>101</td>
</tr>
<tr>
<td>36</td>
<td>36.65</td>
<td>2.450</td>
<td>2.458</td>
<td>vs</td>
<td>110</td>
</tr>
<tr>
<td>40</td>
<td>39.56</td>
<td>2.276</td>
<td>2.282</td>
<td>s</td>
<td>102</td>
</tr>
<tr>
<td>41</td>
<td>40.38</td>
<td>2.232</td>
<td>2.237</td>
<td>m</td>
<td>111</td>
</tr>
<tr>
<td>44</td>
<td>42.55</td>
<td>2.123</td>
<td>2.128</td>
<td>s</td>
<td>200</td>
</tr>
<tr>
<td>47</td>
<td>45.88</td>
<td>1.976</td>
<td>1.980</td>
<td>m</td>
<td>201</td>
</tr>
<tr>
<td>52</td>
<td>50.22</td>
<td>1.815</td>
<td>1.817</td>
<td>vs</td>
<td>112</td>
</tr>
<tr>
<td>57</td>
<td>54.94</td>
<td>1.670</td>
<td>1.672</td>
<td>s</td>
<td>202</td>
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<tr>
<td>59</td>
<td>55.40</td>
<td>1.657</td>
<td>1.659</td>
<td>m</td>
<td>103</td>
</tr>
</tbody>
</table>

\(^1\) Standard d-spacing for low quartz from Brindley & Brown (1980).

Figure 6.4  X-ray diffraction d-spacing correction calculated for quartz peaks. Sample types are for information only. See text for explanation.
Table 6.3  Detailed XRD results - Dioctohedral micas: Muscovite and Illite and Trioctohedral Micas: Biotite and Phlogopite

<table>
<thead>
<tr>
<th>Peak no.</th>
<th>Angle (2θ)</th>
<th>Observed d-Spacing (Å)</th>
<th>Intensity</th>
<th>Probable Reflection Plane</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Dioctahedral(^1)</td>
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<tr>
<td>2</td>
<td>8.99</td>
<td>9.826</td>
<td>s</td>
<td>001</td>
</tr>
<tr>
<td>6</td>
<td>17.92</td>
<td>4.947</td>
<td>m</td>
<td>002</td>
</tr>
<tr>
<td>8</td>
<td>19.91</td>
<td>4.457</td>
<td>m</td>
<td>020</td>
</tr>
<tr>
<td>11</td>
<td>22.55</td>
<td>3.939</td>
<td>m</td>
<td></td>
</tr>
<tr>
<td>14</td>
<td>24.42</td>
<td>3.642</td>
<td>m</td>
<td></td>
</tr>
<tr>
<td>16</td>
<td>25.30</td>
<td>3.517</td>
<td>m</td>
<td></td>
</tr>
<tr>
<td>18</td>
<td>26.74</td>
<td>3.331</td>
<td>100%(^3)</td>
<td>003</td>
</tr>
<tr>
<td>20</td>
<td>27.57</td>
<td>3.233</td>
<td>s</td>
<td></td>
</tr>
<tr>
<td>27</td>
<td>31.35</td>
<td>2.851</td>
<td>w</td>
<td></td>
</tr>
<tr>
<td>33</td>
<td>35.09</td>
<td>2.555</td>
<td>m</td>
<td>200, 131</td>
</tr>
<tr>
<td>36</td>
<td>36.65</td>
<td>2.450</td>
<td>vs(^3)</td>
<td></td>
</tr>
<tr>
<td>38</td>
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</tr>
<tr>
<td>46</td>
<td>45.58</td>
<td>1.989</td>
<td>s</td>
<td>005</td>
</tr>
</tbody>
</table>

\(^1\) From comparison with table 1.15 (Bailey, 1980).

\(^2\) From comparison with table 1.12 (Bailey, 1980).

\(^3\) Mica peaks probably masked by quartz.
dioctahedral micas the 1M reflections arise from illite and the 2M₁ reflections from muscovite, simply because the 1M structure is the most common in the illites and the 2M₁ the most common in crystalline muscovite (Bailey, 1980).

It should be remembered that illite is not a well defined mineral (Bailey, 1980). Examination of Figure 6.2 shows that the basal (low angle) reflections are both broad and asymmetric. These features are common in diffraction traces from illites and are caused respectively by stacking disorders (the 1Md structure) and by interlayering of smectite layers in the illite structure. It is believed that many natural illites contain up to 10% interlayered smectite (Reynolds, 1980). The presence of the smectite layers causes corresponding changes in the chemical composition: EDX analyses (Figure 6.5) indicate the presence of iron and magnesium in many of the clay mineral particles, which again is common in illites. Thus the illite in the Claret Formation appears typical in both its structure and chemical composition.

The identification of biotite by means of XRD is problematic. Several of the basal reflections of biotite overlap with those of illite or muscovite (Table 6.3) and the non-basal reflection (peak 36) which does not correspond to either of the common dioctahedral (muscovite/illite) structures is masked by a very strong quartz peak at the same angle. In principal the biotite peaks might also arise from phlogopite (>70% Mg in the trioctahedral layer); however, EDX analyses of mica crystals seen under the SEM have shown a moderate iron content and thus correspond to the composition of biotite. Although XRD analyses for biotite are inconclusive, the EDX analyses together with the optical microscopic examination of the silt- and sand-size fraction (section 6.2.5) confirms its presence. Figure 6.6(a) and (c) illustrates two typical delaminating layered minerals found under the SEM and shows their corresponding elemental compositions determined by EDX (Figure 6.6(b) and (d). Both of these particles have been identified as biotite.

Kaolinite is a breakdown product of feldspars and as such should be expected in the sediments at Bothkennar (cf. section 6.2.3). The evidence, however, is inconclusive. The basic problem arises since the basal spacing of chlorite (14Å) is twice that of kaolinite (7Å) and so the basal reflections of the kaolinite are often obscured by the stronger chlorite peaks and, in addition, several of the non-basal peaks are obscured by the various non-basal mica reflections. However, peaks 3 and 4 probably represent kaolinite and chlorite respectively: the 002 chlorite peak standard d-spacing is 7.05Å whereas the 001 kaolinite peak standard d-spacing is 7.14Å. At these values of d-spacing the underestimate due to machine error would account for the difference between the observed and standard values. Peaks 15 and 39 are also attributed to kaolinite alone. More detailed investigations contracted out to the University of Cambridge (section 6.2.5) have confirmed the presence of kaolinite.
Figure 6.5  (a) Electron micrograph of typical illite grains.  (b) EDX spectrum from the point marked by white O on the micrograph. The gold (Au) peaks are the result of the conductive coating applied to the SEM sample. Note irregular shape of the grains.
Figure 6.8  (a) and (c) Electron micrographs of delaminating biotite grains.  (b) and (d) EDX spectrum for each grain acquired from the point marked by white O on the micrograph. The gold (Au) peaks are the result of the conductive coating applied to the SEM sample (these are suppressed in (d)).
Table 6.4 lists the peaks that are believed to be due to the presence of chlorite. These form a complete integral series up to fifth order from a basal spacing of 13.89 Å, together with some non-basal reflections. The presence of chlorite is further suggested by the collapse of some peaks upon heating to 550°C, a feature commonly seen in clay grade chlorites (cf. section 6.2.5). Detailed comparison of the d-spacings with published examples (Bailey, 1980: table 1.22) suggests that the Ilb structural form is present (the most common natural form) and the sequence of peak strengths (001 peak weaker than 002) suggests an iron-rich variety. The morphology of Chlorite minerals found under the SEM is not that of the typical equant crystals used as examples by Welton (1984) and peaks from surrounding particles generally also appear on the spectrum. The identification of chlorite under the SEM (visually) and by EDX analysis is, therefore, not without its problems but the presence of chlorite was confirmed by XRD analyses (above).

6.2.3 Feldspars

The feldspars are a complex group of minerals that show a range of compositions and crystallographic structures. They are capable of producing a large number of diffraction peaks, which can be difficult to interpret in detail in the presence of other minerals, and so XRD analysis alone is unable to distinguish precise species. To do so it is necessary to use both chemical and optical data in combination. In this research a primary division has been made simply into plagioclase (calcium - sodium) feldspars and alkali (potassium - sodium) feldspars, which can be distinguished by XRD analyses, and to supplement these with EDX data when possible.

Table 6.5 details those peaks assigned to the feldspars. The reflections at around 3.85 Å (peak 12) and 2.92 Å (peak 25) simply shows the presence of feldspar and can be produced by either plagioclase or alkali feldspar. The strong reflection from a d-spacing of around 3.18 Å (peak 21) is indicative of plagioclase whereas the moderate to strong reflections from the spacings of 3.30 Å (peak 19) and from around 2.88 Å (peak 26) are indicative of alkali feldspar. The reflections listed in Table 6.5 have been assigned tentatively from those given in table 6.8 of Brown (1980).

It is likely that potassium feldspars are the more common alkali feldspar based on their more frequent appearance in EDX analyses and the sporadic presence in diffraction traces of peaks attributable to sanidine, the high temperature potassium feldspar. The presence of the latter is also suggested by micrographs such as Figure 6.7(a) which occasionally show heavily corroded silt-sized particles. The EDX analysis (Figure 6.7(b)) shows the typical spectrum of a potassium feldspar. Analyses on other feldspar grains suggest that the sanidine particles have preferentially lost potassium.
Table 6.4  Detailed XRD Analysis - Chlorite

<table>
<thead>
<tr>
<th>Peak no.</th>
<th>Angle (2θ)</th>
<th>Observed d-Spacing (Å)</th>
<th>Relative Intensity</th>
<th>Reflection Plane</th>
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<tbody>
<tr>
<td>1</td>
<td>6.36</td>
<td>13.886</td>
<td>w</td>
<td>001</td>
</tr>
<tr>
<td>4</td>
<td>12.64</td>
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<td>s</td>
<td>002</td>
</tr>
<tr>
<td>7</td>
<td>18.94</td>
<td>4.682</td>
<td>w</td>
<td>003</td>
</tr>
<tr>
<td>16</td>
<td>25.30</td>
<td>3.517</td>
<td>s</td>
<td>004</td>
</tr>
<tr>
<td>28</td>
<td>31.74</td>
<td>2.817</td>
<td>w</td>
<td>005</td>
</tr>
<tr>
<td>32</td>
<td>34.60</td>
<td>2.590</td>
<td>m</td>
<td>202</td>
</tr>
<tr>
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<td>35.39</td>
<td>2.535</td>
<td>m</td>
<td>201</td>
</tr>
<tr>
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</tr>
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<td>49</td>
<td>48.21</td>
<td>1.886</td>
<td>vw</td>
<td>206</td>
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Table 6.5  Detailed XRD Analysis - Feldspars

<table>
<thead>
<tr>
<th>Peak no.</th>
<th>Angle (2θ)</th>
<th>Observed d-Spacing (Å)</th>
<th>Relative Intensity</th>
<th>Probable Mineral Variety</th>
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<td>5</td>
<td>13.98</td>
<td>6.332</td>
<td>w</td>
<td>Plagioclase</td>
</tr>
<tr>
<td>10</td>
<td>22.15</td>
<td>4.011</td>
<td>m</td>
<td>Plagioclase</td>
</tr>
<tr>
<td>12</td>
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<td>3.846</td>
<td>w</td>
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</tr>
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<td>23.66</td>
<td>3.757</td>
<td>m</td>
<td>Plagioclase</td>
</tr>
<tr>
<td>19</td>
<td>26.98</td>
<td>3.303</td>
<td>vs</td>
<td>Alkali Feldspar²</td>
</tr>
<tr>
<td>21</td>
<td>28.04</td>
<td>3.180</td>
<td>vs</td>
<td>Plagioclase</td>
</tr>
<tr>
<td>22</td>
<td>28.60</td>
<td>3.119</td>
<td>w</td>
<td>Plagioclase</td>
</tr>
<tr>
<td>23</td>
<td>29.51</td>
<td>3.025</td>
<td>m</td>
<td>Plagioclase</td>
</tr>
<tr>
<td>25</td>
<td>30.57</td>
<td>2.922</td>
<td>m</td>
<td>Undifferentiated¹</td>
</tr>
<tr>
<td>26</td>
<td>31.05</td>
<td>2.878</td>
<td>m</td>
<td>Alkali Feldspar²</td>
</tr>
</tbody>
</table>

¹ Both plagioclase and the alkali feldspars show reflections in this region.
² The d-spacing suggests the high temperature K-felspar (sanidine).
Figure 6.7  Example of a typical feldspar grain. (a) Electron micrograph and (b) EDX spectrum for Feldspar acquired from the point marked by red O on the micrograph. The gold (Au) peaks are the result of the conductive coating applied to the SEM sample. Note the corroded nature of the grain surface.
6.2.4 Other Minerals

The sediments contain a variety of other minerals, none of which can be recognised by X-ray diffraction alone. Of those which produce peaks on the diffraction traces, the ferromagnesian minerals such as the pyroxenes occur most commonly: nearly all diffraction traces have a number of such peaks, although the details differ from sample to sample. Usually the reflections are seen at higher diffraction angles (above about 28°) and correspond to d-spacings of about 3Å or less. In the case of Figure 6.2, for example, peaks 21 and 23 possibly arise from ferromagnesian minerals, although there are other possible candidates such as feldspars. Peak 27 could equally arise from pyroxene or mica. However, peaks 24 and 31 have been attributed to pyroxene and the presence of ferromagnesian minerals in some samples has been confirmed unequivocally either visually (in the sand or coarse silt fraction (cf. section 6.2.5)) or by EDX analysis of tabular or equant silt grains seen under the SEM.

Minor carbonates, oxides and sulphides have also been identified by visual, chemical or SEM/EDX methods and possible peaks found on some diffraction traces. Amongst those so recognised have been rutile (titanium dioxide), haematite (iron oxide) and ilmenite (mixed iron-titanium oxide). Pyrite occurs throughout the sediment as framboids, often associated with foraminiferal tests and shell detritus (illustrated for example by Figure 6.8). Metastable amorphous iron monosulphides (FeS) are probably responsible for the general black colour of the sediments and it has been suggested that black fine-grained hydrotroilite (FeS·H₂S) imparts a grey or black colour to many Holocene sediments (Berner, 1967; 1980; 1984). Hydrotroilite is poorly diffracting and is therefore easily masked by small amounts of quartz and clay minerals, and is also extremely susceptible to air oxidation. It is therefore difficult to detect by X-ray diffraction or chemical methods. In the lighter grey weathered zone of the Claret Formation these monosulphides have probably oxidised to pyrite (FeS₂). The speed with which the black sediments at Bothkennar fade to a paler grey colour could indicate the presence of hydrotroilite. There is good visual and chemical evidence (from the Eh-pH results below) that hydrated iron oxides/hydroxides (and their manganese analogues) are present in the upper weathered zone but their amorphous nature precludes identification by X-ray diffraction.

6.2.5 Size Fractionation of Minerals

Minerals which are found in the silt- and clay-size fraction of a deposit are often found preferentially in some size ranges as a result of the differential effects of glacial fracture, crushing and grinding due to a combination of breakage along cleavage planes (planes of weakness within a crystal), fracture (breakage along planes other than cleavage planes) and the hardness of the mineral.
Figure 6.8 (a) Electron micrograph of pyrite frambooids within a foraminiferal test; (b) enlargement of boxed area in (a); (c) enlargement of boxed area in (b). Sample from 17.80m, borehole HW3.
Table 6.6 shows the hardness classification, the nature of cleavage and fracture, and the relative resistance to weathering for the minerals found in the Claret Formation. Quartz, for example, undergoes ‘conchoidal fracture’ (due to its crystal structure it does not fracture along cleavage planes), it has a hardness of seven and is resistant to weathering and alteration. This shows that the minerals which are most abundant in the Claret Formation are most resistant to weathering and, apart from quartz, there is no direct relationship between hardness and abundance. It appears, therefore, that the minerals present in the Claret Formation are present, in the first instance, due to their resistance to weathering, but their distribution between the silt- and clay-size fraction is a function of hardness, cleavage and fracture, and is discussed below.

6.2.5.1 Composition of the silt-sized fraction

Figure 6.9 shows a diffraction trace obtained on the nominal 32μm-63μm size fraction (it is likely that there is some contamination from finer particles). Compared with the whole sample traces, it is noticeable that the peaks are sharp and narrow, indicating well developed crystallinity. The dominant minerals are quartz with subsidiary feldspar, muscovite, biotite, chlorite and minor ferromagnesian minerals (probably hornblende): the clay mineral peaks (particularly kaolinite) are suppressed relative to the whole sample traces.

Examination of this fraction by means of optical microscopy (Figure 6.10(a) and (b)) showed that quartz (the colourless, glassy particles) formed the major constituent of this fraction, probably due to a combination of its resistance to weathering, its hardness and its lack of cleavage. The conchoidal fracture of quartz gives the glacially fractured quartz grains their characteristic angular shape and stepped concave fracture faces (cf. Figure 5.29). Figure 6.10(a) also shows a few opaque pink and opaque white minerals, possibly feldspars. Figure 6.10(c) from the fine sand fraction (retained on the 125μm sieve) confirms the presence of biotite.

6.2.5.2 Composition of the clay-sized fraction

The composition of the clay fraction is especially important in engineering geology as it controls plasticity and many related geotechnical properties. This is discussed in Chapters Seven and Eight where the results presented below are related to the geotechnical behaviour of the deposits of the Claret Formation.

A set of samples were analysed in the Department of Earth Sciences at the University of Cambridge (see section 4.4.1.1 for details) in order to identify and quantify the minerals present in the clay fraction. Three XRD analyses were carried out on the clay fraction of each sample: air dried untreated, glycerolated at 105°C, and heated to 550°C. The resultant XRD traces are shown in Figure 6.11.
Table 6.6 Classification of resistance to abrasion and weathering for minerals found in the Claret Formation at Bothkennar.

<table>
<thead>
<tr>
<th>Mineral/Mineral family</th>
<th>Mohs' Hardness Classification (^1,2)</th>
<th>Cleavage (^1,2)</th>
<th>Fracture (^2)</th>
<th>Relative resistance to weathering (^3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Olivine</td>
<td>6.5 - 7</td>
<td>Indistinct</td>
<td>Conchoidal</td>
<td>1</td>
</tr>
<tr>
<td>Feldspar (Ca-plagioclase)</td>
<td>6 - 6.5</td>
<td>2 good</td>
<td>Uneven</td>
<td>2</td>
</tr>
<tr>
<td>Pyroxene</td>
<td>5 - 6.5</td>
<td>Prismatic, good</td>
<td>Uneven</td>
<td>3 - 4</td>
</tr>
<tr>
<td>Amphibole</td>
<td>5 - 6</td>
<td>Good to perfect</td>
<td>None to uneven</td>
<td>5</td>
</tr>
<tr>
<td>Feldspar (Na-plagioclase)</td>
<td>6 - 6.5</td>
<td>2 good</td>
<td>Uneven</td>
<td>6</td>
</tr>
<tr>
<td>Biotite</td>
<td>2 - 3</td>
<td>Perfect basal</td>
<td>None</td>
<td>7</td>
</tr>
<tr>
<td>Feldspar (alkali)</td>
<td>6 - 6.5</td>
<td>2 perfect</td>
<td>Conchoidal to uneven</td>
<td>8</td>
</tr>
<tr>
<td>Muscovite</td>
<td>2.5 - 3</td>
<td>Perfect basal</td>
<td>None</td>
<td>9</td>
</tr>
<tr>
<td>Kaolinite</td>
<td>2 - 2.5</td>
<td>Perfect basal</td>
<td>None</td>
<td>10 (^4)</td>
</tr>
<tr>
<td>Chlorite</td>
<td>2 - 3</td>
<td>Perfect basal</td>
<td>None</td>
<td>10 (^5)</td>
</tr>
<tr>
<td>Illite</td>
<td>1 - 2</td>
<td>Perfect basal</td>
<td>None</td>
<td>10 (^6)</td>
</tr>
<tr>
<td>Quartz</td>
<td>7</td>
<td>None</td>
<td>Conchoidal</td>
<td></td>
</tr>
</tbody>
</table>

\(^1\) Deer et al. (1964).
\(^2\) Hamilton et al. (1992).
\(^3\) Henderson (1982).
\(^4\) Kaolinite is a weathering product of feldspar and under normal conditions is an end product of the weathering process.
\(^5\) The chlorite group of minerals are formed during metamorphism by a reaction involving aluminium-rich hornblende (an amphibole) or epidote and by the degradation of ferromagnesian minerals and is an end product of weathering/alteration.
\(^6\) The illite group of minerals develop by the alteration of micas and potassium feldspars.
Figure 6.9  X-ray diffraction trace from the coarse silt fraction. Numbered peaks are identified in Table 6.1.
Figure 6.10 Micrographs from the silt- and sand-size fraction.
(a) and (b) Sample of the silt-size fraction: borehole HW3, 6.87m and 18.96m respectively. The majority of the grains are quartz, but note also: dark ferromagnesium minerals [F], milky feldspar [Fs], and muscovite [M].
(c) Sample from the fine sand fraction (6.87m) showing biotite flake [B] and adjacent angular quartz grain [Q].
Figure 6.11(a) shows a comparison between the untreated sample and that heated at 550°C. The heating is mainly concerned with chlorite and a number of peaks are seen to have changed after heating to 550°C. The 001 peak at \( \sim 14.1 \text{Å} \) shifted to \( \sim 13.7 \text{Å} \), and the other basal peaks (marked 002, 003 and 004) collapsed after heating. Both occurrences are diagnostic of chlorite (Brown & Brindley, 1980). Figure 6.11(b) shows a comparison between the untreated sample and the glycerolated sample. It can be seen that a new peak appeared at d-spacing 17.7Å (2θ = 4.9\(^\circ\)): this is indicative of smectite in a sample treated with glycerol (MacEwan & Wilson, 1980; C.V. Jeans, pers. comm.). The results of the semi-quantitative analyses carried out by Dr. Jeans are summarised in Table 6.7.

### 6.2.6 Implications for Sediment Provenance

The principal minerals in all of the assemblages have consistently been found to be quartz, feldspar, mica (illite, biotite and muscovite) and chlorite, with other ferromagnesian minerals, sulphides and oxides as minor constituents. These minerals are all typical of the mixed rock suite from which the Claret Formation has been ultimately derived.

Quartz is ubiquitous in the sedimentary and igneous rocks of the area except, of course, in the olivine basalt lavas. It would therefore be expected to be found in any derived sediments. The angular, fractured nature of the majority of the quartz grains found in the Bothkennar sediments would appear to confirm that these sediments are glacially derived. Krinsley and Donahue (1968), Krinsley and Doornkamp (1973), Whalley and Krinsley (1974), and Eyles (1978) and Bull (1978) have investigated the nature of sand grains as a method of classifying their morphology with respect to sedimentary processes and have suggested that angular, fractured particles were probably diagnostic of a glacially derived sediment. However, Whalley (1978; 1979) and more recently Wright (1995) and Wright et al. (1998), have found that subglacial processes which cause the rounding of some of the edges of subglacially derived particles produced similar surface textures to particles derived from non-glacial sedimentary environments characterised by high energy transport. An abundance of angular, fractured mineral particles within both the sand and silt fraction, together with quartz rock flour in the clay fraction of a sediment (as is the case with the Claret Formation deposits at Bothkennar) would, however, support the theory that those sediments had undergone comminution during glacial erosion (Eyles 1978).

The Dalradian metasediments (grits and mica schists) north of the Highland Boundary Fault and the sedimentary Old Red Sandstone strata of the Midland Valley are possible sources of muscovite, biotite, and Fe-chlorite (cf. Chapter Two: sections 2.2.1 and 2.2.2). Layered minerals in the sedimentary Carboniferous strata included muscovite and biotite (cf. section 2.2.3). The occurrence of these minerals in the Bothkennar sediments would be no surprise in view of the ice flow paths presented in Chapter Two (Figure 2.1).
Figure 6.11  Comparisons of XRD traces from the clay-size fraction (borehole HW3, 16.76m). See text for explanation.
(a) untreated sample and heated at 550°C,
(b) untreated sample and glycerolated sample.
Table 6.7  XRD analyses of the clay fraction of the Claret Formation. Interpretation by C.V.Jeans, University of Cambridge.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Percentage of minerals present in the clay fraction¹</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Quartz²</td>
</tr>
<tr>
<td>7.23m</td>
<td>20.9</td>
</tr>
<tr>
<td>16.78m</td>
<td>18.4</td>
</tr>
</tbody>
</table>

¹ These analyses were semi-quantitative using the external binary standard mixture method explained in Brindley & Brown, (1984: pp. 414 - 417), (Jeans  *pers comm.*, 1999).

² XRD analyses performed on unoriented samples. Analyses on the remaining minerals were performed on oriented samples.
Both potassium and plagioclase feldspars are common constituents of the igneous and sedimentary rocks found widely in central Scotland and both are therefore expected in the sediments of the Claret Formation at Bothkennar. Feldspar grains are more susceptible to chemical corrosion than quartz and glacially derived feldspars generally show more evidence of weathering and chemical dissolution compared to the sharp angular faces and edges of quartz grains. Kaolinite, the breakdown product of feldspar minerals would also be expected and their presence was confirmed by Jeans (pers comm.). The widespread Devonian and Carboniferous lavas are a possible source of ferromagnesian minerals including pyroxene and olivine, whilst amphibole is found in the metamorphic rocks of the highland region of Scotland.

The presence of iron monosulphide at Bothkennar indicates an availability of sulphate and organic material whose breakdown by anaerobic sulphate reducing bacteria (such as Desulfovibrio desulfuricans) releases hydrogen sulphide (H$_2$S) (Berner, 1970; 1980; 1984; Ruddy, 1997). In the presence of iron compounds black, amorphous iron monosulphides (FeS) are formed. These in turn can be converted to pyrite (FeS$_2$) in the presence of free sulphate, supplied by for example the saline (or brackish) water at the sediment surface. Iron sulphides, therefore, are more likely to be authigenic, rather than having been transported along with the main body of sediment and, therefore, their presence does not aid the understanding of the sediment provenance.

6.3 PARTICLE SIZE DISTRIBUTION

6.3.1 General Character of the Particle Size Distribution

The full grading over the range 1μm - 250μm has been obtained on 130 samples from borholes HW3 and HW7 at an interval of ~150mm, using a Micromeritics SediGraph. Figures 6.12(a) and (b) show summary profiles for the two borholes. The results show that sediments are almost entirely clayey-silts (clay content <50%) and that there is no clear difference in grading between the bedded and mottled facies, despite their obviously different visual appearance. The sand-sized content is generally less than ~5% by weight, other than immediately above the Bothkennar Gravel Formation (~24%) and, very occasionally, in the uppermost 5.3m, for example at 2.3m (11%) and 5.3m (8%).

The median size-modal size-clay % profiles (Figure 6.13) can be divided into four general zones (A-D). Below about 14.7m depth (zone A) the median size (Figure 6.13(a)) is in the range 3μm to 8μm and the modal size (Figure 6.13(b)) is quite variable (mainly 15μm to 30μm). There is a noticeable transition at this depth (into zone B) within which the median size is lower (~2μm to 4μm) and the mode is less variable (~15μm to 23μm). The transition between zones A and B occurs at a bedded to mottled facies transition, although other such transitions do not lead to such prominent breaks in grading. At 10.6m depth there is a further transition (into zone C),
Figure 6.12(a) Profile of particle size distribution for borehole HW3.
Figure 6.12(b) Profile of particle size distribution for borehole HW7/8.
above which the median size increases to 1μm to 7μm with a much increased scatter. The modal size does not increase correspondingly, although there is also an increase in scatter. At about 4.1m there is a further sharp transition (into zone C) where the median envelope narrows to between 5μm and 8μm. Above that point the modal size varies between about 15μm and 25μm. In the case of the percentage of clay (Figure 6.13(c)), zone A corresponds to a relatively narrow envelope in the range ~35% to 45% and zone B to a slight increase to 35% to 50%. In Zone C there is a very considerable increase in scatter and the envelope widens from ~30% to over 50%, while in zone D the envelope narrows again to between 35% to 40%.

Figure 6.14 shows the corresponding profiles for borehole HW7. The transition that occurred at about 4.10m in borehole HW3 (zone C to D) is repeated here and there is a general increasing trend in the median size (Figure 6.14(b)) and a decrease in the clay content (Figure 6.14(c)) above this point.

Figure 6.15 shows a series of ternary diagrams in which these data have been subdivided by facies. Figure 6.15(a) is a conventional sand-silt-clay ternary diagram. It is immediately clear that, since the sand content is very limited (<5% for most samples), this type of plot is not the best for separating the data points. For comparison, the same data is plotted in Figure 6.15(b) as sand + coarse silt, medium silt and fine silt + clay. This shows that although the bedded and mottled facies overlap almost completely, there is a better spread of data points and differences between the individual samples can be better appreciated. A further aspect of the grading in general is shown in Figure 6.15(c). When the ternary distribution of the silt fraction alone is considered, it is found that most of the variation occurs in the proportion of coarse silt; the variation in the medium:fine silt ratio is much less. There is a possible hydrodynamic reason for this, since the coarsest particles will be the first to be lost as the velocity of a flow reduces. The figure also suggests that variations in the coarse silt fraction may be the most useful in discriminating between the facies, an idea that will be examined in section 6.3.3.

6.3.2 Discussion of the grading curves and profile zonation

Three parameters - median size, modal size and percentage clay - have been chosen to characterise the grading curve. The nature of their inter-relationship is demonstrated in Figure 6.16(a) for samples selected on the basis of their differing median sizes. The curves are generally very similar at their coarse end: the variation occurs at their fine end and in the curve shape. These features are captured by the median diameter (d50), which (for these samples) varies between ~5.5μm and 2.3μm and describes the overall position of the curve, and by the strength of the mode\(^3\). The percentage of clay is in some sense redundant if d50 is used (or vice

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\(^3\) It should be noted that the modal size shown in Figures 6.13 and 6.14 is that of the dominant mode and therefore does not illustrate the polymodal nature of the samples.
Figure 6.13 Profiles of median grain size, modal grain size and percentage clay for borehole HW3.
Figure 6.14 Profiles of median grain size, modal grain size and percentage clay de HW7.
Figure 6.15  Ternary compositions of samples from boreholes HW3 and HW7. (Bedded burrowed from HW7 only).
Figure 6.16 Comparison of cumulative and frequency particle size distributions for selected samples from the bedded and the mottled facies, borehole HW3.
versa) but it is a useful parameter which is significant for the engineering description of the sediment (it is widely associated with its plasticity) and so it is retained here.

The overall shape of the curve arises from the polymodal distribution of particle size, a probable consequence of their original glacigenic origin, allied to differences in the dominant modes between the samples. This is illustrated by Figure 6.16(b), which shows the particle size data presented in the form of frequency curves. In these particular samples there are a limited number of recurrent size modes at around 14-15\(\mu\)m, 8-9\(\mu\)m, 4-5\(\mu\)m and 1-2\(\mu\)m. The relative size of these modes, however, varies considerably between the samples and it is this, together with the differences in the median size (Figure 6.16(a)), that is the distinguishing feature.

The division of the profile into four zones raises two questions: what mechanisms have caused the changes and what does this imply for the interpretation of the depositional conditions. Although the latter will be addressed in Chapter Eight, it is convenient to anticipate some key results at this stage.

If it is assumed that the grading represents the finer fraction of the glacial till found widely in central Scotland (cf. Chapter Two), then it may be supposed that the source material was fairly consistent in terms of available particle sizes and thus that variation is largely the result of subsequent hydrodynamic sorting by estuarine processes. Typically in such a case, sorting would occur by loss of traction (due probably to the slowing of the basal current) and so the coarser particles would be lost first (cf. Figure 6.15(c)). This would probably bring one of the lower size modes (Figure 6.16(b)) into greater prominence and it may be argued from Figure 6.13(b) that changes in modal size are a sensitive indicator of such hydrodynamic changes. The concomitant lowering of the median size and increase in the percentage of clay size are also evident.

Taking Figures 6.13 and 6.14 as a guide, it is possible to develop an interpretation of the threefold division as a higher current velocity unit (zone A), a lower current velocity unit (zone B) and a return to a higher current velocity unit (zones C and D). These currents will have been tidal or wave-driven. The reasons for the changes are entirely speculative but may include decreasing water depth and alteration of the geometry of the estuary as the sediment infill accumulated. Such changes could account for the reduction in current velocity from unit A to unit B, although the reason for a rapid transition is unclear. In the case of zones C and D there is an association with other sedimentological and biological evidence of shallow water (cf. Chapter Five) and so the increased current velocities may have been due to stronger wave action at seabed as water become more shallow and eventually intertidal (possibly at D).
6.3.3 Differences in Grading Between the Facies

The bedded and mottled facies are visually distinguishable on the basis of the extent of post-depositional bioturbation. This is apparently selective and appears subjectively to develop preferentially in the finer beds. For example, the material from the mottled units often had a smooth, greasy feel when moulded between the fingers, whereas that from the other units had a slightly frictional feel. Furthermore, when the split cores were being cleaned by osmotic knife, the mottled units would normally produce a smooth, shiny surface whereas the non-mottled units (the bedded facies) would offer more resistance to the knife and often had a somewhat matt surface once cleaned. These subjective observations suggest that there is some difference in grading between the two facies and the data was therefore examined further to test this hypothesis.

Figure 6.17(a)-(f) shows cumulative grading curves for samples from both boreholes HW34 and HW7, differentiated by facies and subfacies. It is apparent that nearly all the curves are contained within a similar, narrow, envelope, other than in the case of the burrowed subfacies, where the spread is much greater, and the basal subfacies where the two lowest samples in the core have a much higher sand content. Examination shows that the greater spread in the burrowed subfacies is due to the inclusion of better sorted samples which are lacking in the clay-size fraction: these are the infillings of the burrows, which introduce a degree of heterogeneity at the centimetre scale. This is, of course, emphasised by the small size of the samples used by the SediGraph, which were positioned precisely to include or exclude these features. The narrower envelope of the mottled Type III samples could be due to the small sample number (n = 6), but it is possible that it results from these mottles generally being confined to narrow horizons in the upper part of a mottling sequence.

Figure 6.18(a), (b) shows these data in a more compact form on a ternary diagram (the choice of apices was discussed earlier). Once again, the data points fall in a tight group and the bedded and the mottled facies have therefore been separated (cf. Figure 6.15). In Figure 6.18(a) the outlying data points, as expected, belong very largely to the burrowed subfacies: if these are excluded the spread of points is greatly reduced. In Figure 6.18(b) it is evident that there is much overlap between the types I and II mottles, whilst the spread of Type III mottles is much less.

These results raise the question of whether the facies really are distinguishable by grading, which is counter to the subjective impression mentioned earlier. To test this further, the values of median size, modal size and clay percentage were plotted as frequency histograms (Figures 6.19(a)-(c) to 6.20(a-c)). These give the visual impression that the bedded and mottled

---

4 Particle size data from the bedded (burrowed) subfacies in the uppermost section of the Claret Formation in borehole HW3 have been excluded from Figures 6.17 to 6.21. It is believed that selective sampling in this section resulted in unrepresentative samples.
Figure 6.17(a-c) Cumulative grading curves for the bedded facies from borehole HW3, (burrowed subspecies samples from HW7).
Figure 6.17(d - f) Cumulative grading curves for the mottled facies from borehole HW3 and HW7.
Figure 6.18 Ternary compositions of samples from the mottled and bedded facies, boreholes HW3 and HW7 (burrowed subfacies borehole HW7 only).
Figure 6.19  Variation of median size within the mottled facies (a-c) and bedded facies (d-f), boreholes HW3 and HW7 (bedded burrowed from HW7 only).
Figure 6.20 Variation of modal size within the mottled facies (a-c) and bedded facies (d-f), boreholes HW3 and HW7. (bedded burrowed from HW7 only).
Figure 6.21 Variation of clay percentage within the mottled facies (a-c) and bedded facies (d-f), boreholes HW3 and HW7 (bedded burrowed from HW7 only).
Table 6.8  Particle size comparisons of bedded and mottled facies, boreholes HW3 and HW7 (excluding the burrowed subfacies).

(a) Descriptive statistics

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<thead>
<tr>
<th></th>
<th>Bedded facies</th>
<th>Mottled facies</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Median size (μm)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean</td>
<td>4.50</td>
<td>4.21</td>
</tr>
<tr>
<td>Standard deviation</td>
<td>1.53</td>
<td>1.69</td>
</tr>
<tr>
<td><strong>Modal size (μm)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean</td>
<td>21.9</td>
<td>18.6</td>
</tr>
<tr>
<td>Standard deviation</td>
<td>5.37</td>
<td>5.01</td>
</tr>
<tr>
<td><strong>Clay size (%)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean</td>
<td>41.9</td>
<td>42.2</td>
</tr>
<tr>
<td>Standard deviation</td>
<td>4.37</td>
<td>4.93</td>
</tr>
</tbody>
</table>

(b) t-Test: Two-Sample Assuming Unequal Variances

<table>
<thead>
<tr>
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<th>Bedded</th>
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</tr>
</thead>
<tbody>
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<td><strong>Median particle size</strong></td>
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</tr>
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<td>97</td>
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<td>t Stat</td>
<td>0.902</td>
<td>3.199</td>
</tr>
<tr>
<td>P(T&lt;=t)</td>
<td>0.369</td>
<td>0.002</td>
</tr>
<tr>
<td>t Critical</td>
<td>1.984</td>
<td>1.985</td>
</tr>
</tbody>
</table>

No significant difference  Significant at 99% level  No significant difference

(c) F-Test: Two-Sample for Variances

<table>
<thead>
<tr>
<th></th>
<th>Bedded</th>
<th>Mottled</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Median particle size</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean</td>
<td>4.50</td>
<td>4.21</td>
</tr>
<tr>
<td>Variance</td>
<td>2.34</td>
<td>2.87</td>
</tr>
<tr>
<td>Observations</td>
<td>49</td>
<td>51</td>
</tr>
<tr>
<td>df</td>
<td>48</td>
<td>50</td>
</tr>
<tr>
<td>F</td>
<td>0.816</td>
<td>1.119</td>
</tr>
<tr>
<td>P(F&lt;=f)</td>
<td>0.240</td>
<td>0.347</td>
</tr>
<tr>
<td>F Critical</td>
<td>0.621</td>
<td>1.605</td>
</tr>
</tbody>
</table>

No significant difference  No significant difference  No significant difference
facies may differ, although the extent to which this is significant is unclear. To establish this, the means and variances of all three parameters were tested to find whether any differences between the bedded and mottled facies were statistically significant. The burrowed subfacies was excluded from these analyses on the grounds that it is clearly heterogeneous.

The results are presented in Table 6.8. The mean values of the modal particle sizes are statistically different (at the 95% level or greater) between the bedded and mottled facies, while the means of the median and percentage clay size are not. The variances, however, of all three parameters are not statistically different. The modal size in the bedded facies is about 18% larger than in the mottled facies, due to a relative increase in the coarser modes of the polymodal distribution. This probably explains the subjective difference in texture\(^5\). It is presumed that it is in fact an increase or decrease in this coarser material that generates the two distinctive facies, in that the burrowing organisms responsible for the bioturbation are limited by the amount of the coarser silt particles in their substrate. Evidently they do not colonise the sediment if it contains too many coarser particles.

### 6.4 ORGANIC MATERIAL

As described in Chapter Four, the organic content of the sediment was investigated using fresh samples that were collected from a new borehole (HW9). The organic material was characterised in four separate ways: potassium dichromate oxidation was used to determine the total organic content; sulphuric acid-phenol colorimetry to determine the content of monosaccharide residues; Kjeldahl reduction to determine organic nitrogen; and soxhlet extraction using a methanol-toluene azeotrope to estimate the lipid content. The last three of these are general proxy measures that can be used respectively to characterise soil polysaccharides and other carbohydrates; proteins, peptides and other nitrogenous residues; and oils, fats, and waxes.

#### 6.4.1 Nature and Occurrence of the Organic Component

Total organic material (Table 6.9) comprises between 2% and 4% by weight of the samples tested here, which is typical for the upper part of the Bothkennar sequence (Hawkins et al., 1989; Hight et al., 1992a; Paul et al., 1992a). From this figure it is possible to estimate the annual organic productivity of the palaeo-Forth estuary. Given an average sediment accumulation rate at Bothkennar of about 10mm a\(^{-1}\) (Barra and Paul, 1999), this organic content equates to an organic carbon accumulation rate in the order of 200g m\(^{-2}\) a\(^{-1}\). This value is typical of modern

\(^5\) It should be noted that due to the small-scale selective sampling mentioned above, it is probable that the modal and median size of the bedded facies were underestimated: silt laminae within the bedded facies were often excluded from the SediGraph samples. The differences between the bedded and mottled facies, therefore, are probably an underestimate, although the interpretation of the results would stand. A higher percentage of coarser material in the bedded facies could result in the percentage clay size also being significantly different between the two facies.
Table 6.9 Geochemical composition: borehole HW9.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Total organic content</th>
<th>Total organic carbon(^1)</th>
<th>Monosaccharide residues</th>
<th>Kjeldahl nitrogen</th>
<th>Methanol-toluene extract</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.43</td>
<td>3.9</td>
<td>2.3</td>
<td>0.32</td>
<td>0.12</td>
<td>1.86</td>
</tr>
<tr>
<td>2.61</td>
<td>3.3</td>
<td>1.9</td>
<td>0.55</td>
<td>0.11</td>
<td>1.53</td>
</tr>
<tr>
<td>3.01</td>
<td>2.6</td>
<td>1.5</td>
<td>0.56</td>
<td>0.12</td>
<td>2.01</td>
</tr>
<tr>
<td>3.19</td>
<td>2.4</td>
<td>1.4</td>
<td>0.39</td>
<td>0.10</td>
<td>1.83</td>
</tr>
<tr>
<td>3.41</td>
<td>3.2</td>
<td>1.9</td>
<td>0.49</td>
<td>0.11</td>
<td>2.01</td>
</tr>
<tr>
<td>3.69</td>
<td>3.9</td>
<td>2.3</td>
<td>0.59</td>
<td>0.13</td>
<td>2.21</td>
</tr>
<tr>
<td>4.09</td>
<td>3.1</td>
<td>1.8</td>
<td>0.45</td>
<td>0.11</td>
<td>1.65</td>
</tr>
<tr>
<td>4.26</td>
<td>3.1</td>
<td>1.8</td>
<td>0.60</td>
<td>0.14</td>
<td>2.39</td>
</tr>
<tr>
<td>4.63</td>
<td>3.0</td>
<td>1.7</td>
<td>0.60</td>
<td>0.13</td>
<td>1.37</td>
</tr>
<tr>
<td>4.81</td>
<td>3.7</td>
<td>2.1</td>
<td>0.64</td>
<td>0.13</td>
<td>2.08</td>
</tr>
</tbody>
</table>

\(^1\) Conversion factor Total organic content to Total organic carbon = 0.58.
This is based on the potassium dichromate method of organic content determination which assumes that soil organic matter contains an average of 58% of carbon by mass (BS 1377: 1975).
temperate estuaries (Wollast, 1991) and shows that in this respect conditions in the Forth were similar to those at the present day. The variability in the percentage of total organic material can also be explained most simply by the temporal variability of the sedimentation rate, as deduced from $^{14}$C dating (Paul & Barras, 1998), and does not necessarily imply major changes in organic productivity. Such variability is consistent with those SEM images (cf. section 5.5.2) that suggest occasional condensed sequences associated with an increase in reworked biogenic debris (Paul et al., 1992b).

A variety of organic materials and structures to be found in intact samples from the Bothkennar clay were presented in Chapter Five. These features include in particular biogenic structures with mucal structural linings (Figure 5.33) and probable faecal pellets (Figure 5.31). In one of these examples, individual soil particles appear to be cemented into larger structures by organic material. This is a key observation that provides some insight into the role played by organic materials in the mechanical behaviour of the Bothkennar sediments.

6.4.2 Geochemical Composition

Table 6.9 details the composition of the organic component in terms of the three gross parameters described above: monosaccharide residues, Kjeldahl nitrogen and the methanol-toluene extractable fraction. As shown on Table 6.9 and in Figure 6.22, the weight percentage of monosaccharide residue varies with depth by a factor of around two, whereas organic nitrogen is relatively constant with depth. The average monosaccharide content, if assumed to be composed only of glucose units (1 mole = 180g) equates to ~27 µmoles g$^{-1}$, which again is typical of many inshore sediments at the present day (Degens & Mopper, 1975). The weight percentage of the methanol-toluene extract is more variable with depth but generally correlates with the total organic material. These three components together account for up to 100% of the total organic material by weight and it is presumed that any remainder comprises materials, possibly of high molecular weight, which are resistant to the procedures employed in this work. The margins of error for the methanol-toluene method are believed to be quite wide, which is thought to be reason why the components apparently add up to >100% in a few cases.

The results indicate an overall C:N ratio in the range 11 to 16, a relatively low value that indicates a marine rather than a terrestrial origin for the material since terrigenous organic material, which is often dominated by cellulose and its products, typically has a C:N ratio up to 100 or higher (Tyson, 1996: table 22.1). When the weight percentages of monosaccharide residues, Kjeldahl nitrogen and methanol-toluene extractable material are plotted on a ternary diagram (Figure 6.23), the compositions fall along a line on which the nitrogen and monosaccharides maintain an almost constant weight ratio of 1:4.37±0.63. If the sugar is a hexose (six carbon), this corresponds to a consistent nitrogen:carbon atomic ratio close to 1:2 (1:2.04±0.30). The reason for
Figure 6.22 Profile of total organic material and organic components: borehole HW9.

Figure 6.23 Ternary composition of organic material in borehole HW9.
this value is unclear, although it is possible that it arises from some particular but unidentified amino-carbon material, such as a mucopolysaccharide (e.g. from diatoms) or chitin (e.g. from crustaceans, polychaetes and zooplankton). Both of these materials would be expected to be present in the sediments in view of the organisms identified in Chapter Five.

6.4.3 Origin of the Organic Component

Although the term 'organic' can suggest a fibrous or peaty material of terrestrial plant origin, such plant remains are seen only very occasionally in the samples. It is thought instead that the major proportion of the organic material at Bothkennar, as is the case in most Holocene estuarine sediments, originated from planktonic or benthic organisms present in the water and sediments of the contemporaneous estuary.

Planktonic organisms are responsible for 'marine snow': a general term for detritus composed of inorganic aggregates, organic mucal agglomerates and zooplankton faecal pellets (Syvitski, 1991), evidence of which can be seen frequently in electron micrographs such as those in Figures 5.8, 5.31 and 5.32, and also illustrated in Paul et al. (1992b). In the case of the benthic organisms, studies of the modern benthic biota of the intertidal mudflats near Bothkennar (McLusky, 1987; Moore, 1987) have revealed that they are dominated by worms (polychaetes and nematodes), crustaceans and molluscs. Many species in these various groups produce mucopolysaccharides, which they use as structural materials in burrow systems (Howard & Frey, 1973). The intertidal mudflats also support an epipelagic diatom flora, which are known elsewhere to bind sediment surfaces by mucopolysaccharides exuded as a means of locomotion during the tidal cycle (Holland et al., 1974; Frostick & McCave, 1979; Paterson et al., 1990). The body tissues of the organisms themselves are high in lipids and also supply a range of nitrogenous materials to the sediment.

Following deposition and burial, these various organic materials degrade in a variety of ways, the exact details of which control the mix of compounds found in the sediment. There will have been a relative loss of nitrogen, proteins and sugars will have been partially consumed by bacterial action and ill-specified 'complexes' will have formed, perhaps using clay surfaces as catalytic templates (Hedges, 1977; Syvitski & Murray, 1981; Harvey et al., 1983). In its final state in the sediment the organic material will thus consist of a mixture of identifiable structural components, amorphous materials and secondary degradation products. The biochemical details of these products were not examined in the present study but, as will be shown in sections 7.3.5 and 7.8, some geotechnical properties of the sediment are related to the relatively simple characterisation of sections 6.4.1 and 6.4.2.
6.5 INORGANIC GEOCHEMISTRY

Although the full study of the geochemistry of the Claret Formation was beyond the scope of the projects on which this thesis is based, certain aspects were studied as an adjunct to the geotechnical investigations. An interest in possible cementation led to the extraction of iron and manganese by two comparative methods in order to investigate the presence of amorphous oxides and hydroxides; in turn this produced information about the weathering profile in the upper part of the sequence. The obvious fact that the Bothkennar research site is located immediately adjacent to the Forth estuary led to an interest in the chemistry of the interstitial pore water, particularly into the presence of cations that might modify the engineering behaviour of the soil. This in turn led to additional knowledge about the possible evolution of the interstitial water and aided the comparison of Bothkennar with other soft clay sites as discussed in Chapter Eight.

The investigations described in this section are based on boreholes HW3, HW7 and HW9. Borehole HW3 provided a profile of sodium, potassium, magnesium and calcium throughout the local thickness of the Claret Formation down to the Bothkennar Gravel Formation, albeit at a rather coarse spacing (~1m). Neither Eh-pH nor acid-extractable cations were measured in this borehole. In order to remedy this, further measurements of both Eh-pH and of the acid-extractable cations iron, manganese and titanium were made to a depth of ~4.8m in borehole HW9 adjacent to HW3. In addition, further profiles of these parameters were obtained to ~6m depth in borehole HW7 during the later phase of the study, as described in Chapter Four.

The analytical methods employed for this part of the work have been described in detail in Chapter Four. Measurements of Eh and pH were made using a bench meter with suitable selective electrodes. Dithionate-citrate-bicarbonate (DCB) digestion was used to extract oxides and hydroxides (Mehra & Jackson, 1960) and the more aggressive nitric acid-hydrogen peroxide digestion (Krishnamurty et al., 1976) was used for comparison as a measure of total inorganic iron, including that contained within free oxides and clay colloids. The digested extracts were assayed for iron by atomic absorption spectrophotometry. Sodium, potassium, magnesium and calcium ions were extracted using washing and acetate leaching techniques and analysed using atomic absorption spectrophotometry. From these results the cations have been determined as total extracted weight (in milliequivalents)/soil solid weight.

6.5.1 Eh and pH

Figure 6.24 shows the profiles of acidity (pH) and redox potential (Eh) obtained in the upper part of the sequence. The pH shows a slight but perceptible increase with depth from slightly acid at surface, to neutral at about 0.4m, to slightly alkaline below this depth. Below about 2.5m depth the groundwater alkalinity is constant at around pH 8 and the redox potential is around zero.
Figure 6.24  Geochemical profiles: (a) pH (acidity) and (b) Eh (redox potential) from the trial pit and borehole HW9.
Figure 6.25 Eh-pH data plotted in iron hydroxide stability fields for the trial pit (TP and borehole HW9. (Stability field source: Rowell, 1981: figure 7.18)
volts, which is probably a result of the organic content whose decomposition has created an oxygen deficiency. Above this depth, as the aerated percolation zone is approached, the acidity falls to around pH 6 and the redox potential rises to around 0.5V. [A] to [C3] on Figure 6.24 refer to the clusters of data points on Figure 6.25.

The Eh-pH conditions are an important control on inorganic cementing agents, since these are usually precipitated from interstitial water. The example of iron hydroxide (cf. section 6.5.2) is presented in Figure 6.25, which shows a plot of the measured Eh-pH values superimposed on the stability fields of various iron hydroxides: it suggests that ferric hydroxide may act as a cement throughout much of the profile, since the majority of points lie well outside the dissolution field of aqueous ferrous iron. The clusters of data points represent: a zone of greater oxidation above the water table [A], a zone of decreasing oxidation below the water table [B], a zone [C1,2] where the sediments become anaerobic (Eh < 0.2V). The cluster of higher values of pH [C3] is thought to be due to shell contamination in samples from a zone below the shell bed.

The interpretation of the lower Eh values is open to some debate, since the measurement of low Eh in sulphur-bearing sediments is difficult and it is normal to regard such Eh measurements as only approximate. In the present case it is reasonable to say that there is an Eh gradient with depth from moderately oxidising to weakly reducing. Despite the limitations of this approach, it has some value for the identification of possible cementing agents when direct methods (e.g. XRD) fail due the amorphous nature of many cements.

6.5.2 Iron, Manganese and Titanium

At depth, the sediments are mainly black (5Y2.5/1) due to the presence of finely disseminated iron monosulphide, with occasional very dark gray (5Y3/1) and dark gray (5Y4/1) horizons. Above about 4m below surface, the sediments are lighter in colour (very dark gray to dark gray: 5Y3/1 to 5Y4/1) due to the oxidation of the monosulphide to the disulphide (pyrite). This colour difference allows the recognition of the two common oxidation zones based on the species of iron sulphide: the monosulphide zone and the pyrite zone. The pyrite zone extends upwards to about the water table, normally about 1m below ground surface, at which depth it is replaced by a further, limited zone of irregular iron-staining that takes the form of yellowish-red (5YR4/6) streaks and rings, the latter with light gray (10YR7/1) cores, developed against a gray to grayish brown (10YR5/1 to 10YR5/2) background. This zone is probably developed as a result of the formation of iron oxides and hydroxides and in places appears to be associated with a local increase in the proportion of DCB extractable iron. The boundary between this zone and the underlying pyrite zone is not always well defined, particularly if the sediments contain shell debris.
Figure 6.26 shows the profile (to 6m depth) of iron, manganese and titanium. Iron is quantitatively dominant amongst these as would be expected. Below the crust the acid-extractable iron fraction (principally colloidal hydroxides plus clay coatings and free oxides) comprises 3.96 ± 0.21% of the dry soil weight. This proportion shows some variation with depth that is indicative of depletion, translocation and relative enrichment within the upper part of the profile, as evidenced also by the colour changes that were described above. This in turn is probably the result of leaching in locally more acid areas and subsequent reprecipitation at lower levels in response to changes in both redox and pH. The DCB extractable fraction (principally amorphous oxides and hydroxides) represents about 25% of the acid-extractable fraction as a whole and is relatively more variable at around 0.94 ± 0.22% of the dry soil weight, perhaps reflecting more closely local changes in redox conditions that lead to the increased precipitation of iron compounds (cf. Figure 6.25).

Manganese and titanium show less variation with depth, although both follow the major variations in the iron profile, the extent of the correlation being shown by Figures 6.27 and 6.28 respectively. Manganese shows the better correlation, possibly since some manganese dioxide may have formed as a secondary mineral in association with ferrihydrite (Taylor, 1987), whereas titanium dioxide occurs widely as a detrital mineral and so need not be well-correlated with iron, despite the probable occurrence of the mixed oxide ilmenite (Paul & Tovey, pers. comm.).

6.5.3 Exchangeable Cations

Figure 6.29 shows the profiles of four major exchangeable cations (sodium, potassium, magnesium and calcium) down the full depth of borehole HW3. Several features are apparent. All the profiles are generally similar in form (indicating that, other than calcium, the cation concentrations are well correlated) and overall conform to an approximately parabolic shape. In detail there are a number of inflections in the curves (for example in the sodium curve at 14m depth) which appear to correlate with facies changes, although the rather wide sampling interval makes it difficult to see the correspondence clearly.

A more detailed sodium profile from borehole HW7 (Figure 6.30) shows a slightly different distribution, with two distinct zones above a depth of about 6m. In this borehole there is a steady decrease in sodium content from 6m depth to around 2m depth, which may be due to the infusion of fresh water from precipitation onto the surface and perhaps also from the adjacent, slightly higher, carse lands to the west. This zone of infusion appears to be broadly coincident with the iron pyrite oxidation zone and may thus reflect the input of oxygenated waters from the surface. The higher zone, from 2m depth to surface, contains almost no sodium chloride and is interpreted as a fully desalinated zone, developed largely by vertical leaching in response to precipitation and improved soil drainage.

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Figure 6.26 Composite depth profile of iron, manganese and titanium in borehole HW7 and trial pit TP1. (acid = acid extractable, DCB = DCB extractable).
Figure 6.27 Correlation of acid-extractable cations: Fe with Mn, (borehole HW7 only).

Figure 6.28 Correlation of acid-extractable cations: Fe with Ti, (borehole HW7 only).
Figure 6.29  (a) Profiles of exchangeable cations for borehole HW3,
(b) Expanded profiles from (a) to separate Ca"", Mg"" and K".
Figure 6.30 Profile of exchangeable sodium for borehole HW7.
Figure 6.31 explores some aspects of the possible origin and evolution of the exchangeable cation profiles. It is assumed that in the first instance the pore water originated as estuarine water that was trapped during deposition. Figure 6.31(a) shows the pore-water concentration of sodium chloride that would be obtained if all the exchangeable sodium were combined as sodium chloride and dissolved in the current pore water content. This figure indicates, however, that although the concentration does not exceed that of normal seawater [line B], the peak value exceeds the range probably expected in the estuary itself [interval A] (Riley & Chester, 1971; Webb & Metcalfe, 1987) and thus implies that there may have been some relative increase in the sodium content since deposition. A possible mechanism for such an increase is the expulsion of pore water during selfweight consolidation, whereas dissolved cations may have been retarded by adsorption onto clay surfaces and so become relatively more concentrated in the remaining pore fluid.

Figure 6.31(b) indicates that, subsequent to deposition, there has been a change in the proportions of the cations relative to seawater. The vertical lines (colour coded) represent the ratio in seawater of each of the cations (potassium, magnesium and calcium) with sodium. Clearly the ratio seen in the sediment differs from this and is dependent on position in the profile. The suggestion is that at the top and base of the section there has been, in the cases of magnesium and calcium, a relative enhancement in relation to sodium: since, however, the ratio in the mid-profile is not dissimilar to that in seawater it may be that the 'enhancement' is in fact due to a loss of sodium (as can be seen in Figure 6.29) rather than a gain in the other cations. Conversely, in the case of potassium there appears to have been a definite relative increase, since the ratio even at mid-profile is above that found in seawater by a factor of four to five. The most likely source of additional potassium may well be the partial decomposition of minerals such as muscovite (following degradation to illite) or selective loss of potassium from alkali feldspar (leading to the corrosion seen under the SEM cf. Figures 5.29 and 6.7).

A possible mechanism for the loss of cations from the profile may be diffusion towards lower concentrations in the less saline water near the ground surface and in the coarse sediments of the Bothkennar Gravel Formation. The approximately parabolic shape of, for example, the sodium profile in Figure 6.29 is reminiscent of a pore pressure isochrone during consolidation under two-way drainage. However, the loss of sodium is unlikely to have been due simply to pore water migration during consolidation, since the time required for full consolidation of a 20m layer of clay with the properties of the Bothkennar sediment ($C_v \sim 10 \text{ m}^3 \text{ yr}^{-1}$: two way drainage) can be shown to be less than 100 years (M.A.Paul, pers. comm.) and thus the observed parabolic profile will not have persisted for the lifetime of the deposit ($\sim 3000$ yrs) due to consolidation alone.
Figure 6.31  (a) Profile of equivalent NaCl concentration in present-day pore water, borehole HW3.  
(b) Ratios of major exchangeable cations against exchangeable sodium.  
The lines represent the fixed ratio of the relevant ion against sodium in seawater.
This invites the speculation that diffusion is involved, since the controlling equation (the
diffusion equation) is the same for both consolidation and for diffusion and so the shape of the
isochrones is the same. Such timescales are similar to those associated with diffusion in other
saline clays; for example, Quigley and Crooks (1983) have calculated that in the case of a 35m
thick Leda clay deposit at Hawkesbury, Ontario, a period of around 10,000 years has been
required to produce a parabolic salinity profile similar to that seen at Bothkennar. If this
hypothesis is correct, the loss of sodium relative to magnesium and calcium seen at the top and
base of the profiles in Figure 6.31(b) may be due to the slower diffusion of these divalent species
through the clay structure. A similar relative loss is also seen in the case of potassium which may
diffuse more slowly due to its ionic size.

6.6 CONCLUSIONS

The sediments of the Claret Formation are composed of a mixed mineral suite that comprises
quartz, feldspar, mica (illite, biotite and muscovite) and chlorite, with other ferromagnesian
minerals, sulphides and oxides as minor constituents. These minerals are all typical of the mixed
rock suite from which the Claret Formation has been ultimately derived.

The sediments are almost entirely silty clays or clayey silts with a sand content usually less than
5%. The bedded and the mottled facies are statistically distinct in terms of their modal particle
size, although they do not differ statistically in their median particle size and percentage clay
content. The grading profiles can be divided into four general depth zones based on values of
the median and modal size, the percentage of clay and the width of the envelope which contains
the profiles. It is believed that these zones have a genetic origin that relates to the changing
water depth and tidal conditions during the depositional period.

The organic material is derived largely from estuarine organisms, and plant tissues such as
leaves, stems and root fibres are found only rarely. The material occurs mainly as mucal sheets
or amorphous coatings which cement individual soil particles into larger aggregates or pellets.
Its composition can be characterised in terms of three gross parameters: monosaccharide
residues, Kjeldahl nitrogen and a methanol-toluene extractable fraction. The C:N ratio is
relatively low and indicates a marine rather than a terrestrial origin for the material.

The pH shows a slight but perceptible increase from slightly acid at surface to slightly alkaline
below ~2m depth. Below this depth, the groundwater alkalinity is constant at around pH 8. The
redox potential at this depth is around zero volts, which is probably a result of the organic
content, whose decomposition has created an oxygen deficiency.

Iron is quantitatively the dominant acid extractable cation and below the crust comprises around
4% of the dry soil weight. The DCB extractable iron is relatively more variable. There is some
suggestion that the Eh-pH conditions in the upper sediment profile are appropriate for the
formation of iron oxides or hydroxides, which are likely to be DCB extractable.

The full depth profiles of four major exchangeable cations (sodium, potassium, magnesium and
calcium) are generally similar in form and, other than calcium, they conform to an approximately
parabolic shape. A more detailed sodium profile above a depth of about 6m shows two distinct
zones, the upper being almost desalinated. The ratios of the major cations against sodium show
that although the pore water is likely to have originated as trapped estuarine water during
deposition, it has undergone subsequent modification leading to the relative loss of sodium and
enhancement in potassium concentration. It is proposed that these effects are respectively the
results of long term diffusion and the breakdown of potassium-bearing minerals such as
feldspars and micas.
CHAPTER SEVEN

PROPERTIES OF THE CLARET FORMATION AT BOTHKENNAR
PART 3: GEOTECHNICAL PROPERTIES

7.1 INTRODUCTION

7.2 INDEX PROPERTIES

7.3 DISCUSSION OF THE CONTROLS ON INDEX PROPERTIES AT BOTHKENNAR

7.4 WATER CONTENT AND BULK DENSITY

7.5 LIQUIDITY INDEX

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7.7 COMRESSIBILITY AND YIELD STRESS

7.8 UNDRAINED SHEAR STRENGTH

7.9 SUMMARY
CHAPTER SEVEN

PROPERTIES OF THE CLARET FORMATION AT BOTHKENNAR
PART 3: GEOTECHNICAL PROPERTIES

7.1 INTRODUCTION

This chapter describes the new geotechnical results that have been obtained from the sediments of the Claret Formation at Bothkennar during the course of this research. The programme on which this Thesis is based had only limited geotechnical objectives in view of the other geotechnical investigations that were ongoing at the same time (Hight et al., 1992a) and so this research focussed on just two particular geotechnical questions, both of which had a geological dimension:

- In what ways are the geotechnical properties of the Bothkennar sediments in the remoulded state governed by their composition?

- In what ways are profiles of geotechnical properties governed by the facies architecture and thus by their inferred depositional history?

The first question centres on the relative roles played by grading, exchangeable cation content and organic content in the determination of index properties and the extent to which the bedded and mottled facies may differ in these properties. The second question largely centres on the geotechnical profiles obtained in the Claret Formation and their relationship to the inferred depositional history of the sediments. A central concept in this aspect of the work has been that the vertical profile of a geotechnical property reflects, in part, the sequence of depositional events that occurred as the sediment accumulated. This approach requires the palaeoenvironmental interpretation that is to be developed in the next chapter and so will be considered further in that context.

The locations of the boreholes mentioned in this chapter are shown in Chapter Four (cf. Figure 4.1). Use has been made principally of the paired boreholes HW3/9 and HW7/8. As described earlier, HW3 is the full profile of the Holocene deposits at Bothkennar, down to the buried gravel layer, and so acts as the principal geological reference borehole for the site. HW9 was a shallow (4.8m) borehole adjacent (~1m) to HW3, from which samples were collected specifically for organic geochemical testing. Boreholes HW7/8 sampled the deposits down to a depth of 6.4m as part of a detailed study of the upper portion of the Bothkennar sediments:
HW7 was used for sedimentological and fabric samples, HW8 (positioned about 1m from HW7) for geotechnical testing.

7.2 INDEX PROPERTIES

7.2.1 General concepts

The effect of geotechnical structure is often demonstrated by comparing the properties of an intact sample of a clay to properties measured in the remoulded or destructured state. Two of the simplest, but most useful, properties of destructured clays are their index properties (the liquid and plastic limits), which were introduced by Atterberg (1911) for the classification of soil. The liquid limit (abbreviated as LL or wL) is the water content at which the soil exhibits a defined behaviour (e.g. a given penetration) in a test such as the cone penetration test (e.g. BS 1377 1990: British Standards Institute, 1990). Conversely, the plastic limit (PL or wP) is the water content at which the soil exhibits another defined behaviour (fracturing of a thin thread) in a standard test. The difference between the two is known as the plasticity index (PI or Ip). A plot of plasticity index against liquid limit (Ip/wL), commonly known as a ‘plasticity chart’, offers a useful guide in which the soil is classified according to its position relative to a sloping straight line (the A-line). First introduced by Casagrande (1948), this line represents the boundary between inorganic clays (above) and organic silts and clays (below).

An interesting observation is that natural soils do not plot all over the plasticity chart at random. Instead, they are often confined to well-defined bands that lie approximately parallel to the A-line. Skempton (1970) showed that a range of sedimentary clays, which were largely of marine origin, all plotted along a narrow line displaced to the left of the A-line by a few percent of water content. Skempton’s line, which has since become known as the marine clay line, appears characteristic of a large number of such materials. Similarly, modern glacial tills from Spitsbergen were shown by Boulton and Paul (1976) to follow an analogous line (which they called the T-line), displaced to the left again of the marine clay line. The reason that different genetic types of soil fall along separate but clearly defined lines is probably the result of their grading: in particular the relative proportion of sand:silt:clay. This proportion appears to be reasonably consistent between soils of a common genetic origin (i.e. their grading curves are similar) and so causes them to plot in a consistent manner on the plasticity chart in a manner first described by Dumbleton and West (1966) and later elaborated by Trentor (1999).

The plasticity index itself is controlled by the proportion of the clay-size fraction, its mineralogy and other aspects of the soil geochemistry (including the dominant amorphous component and the porewater chemistry). Where the clay-size fraction is dominated by rock flour (consisting for example of quartz and feldspar), the plasticity will normally be low unless there is a significant non-peat organic content. Conversely, if the dominant clay mineral were illite, the plasticity
would be greater, with liquid limits perhaps in the range of 60% to 90% (Culshaw et al., 1991), the higher values occurring should organic material also be present. Swelling clays such as smectites have intrinsically high values of plasticity and are not greatly influenced by organic material. Conversely, in the absence of organic materials and where inorganic amorphous material dominates, the plasticity is characteristic only of the minerals themselves.

For a particular soil, the plasticity increases with the percentage of clay-sized particles. This is demonstrated by plotting, for samples of given type of soil, the plasticity index against the clay-sized content for each sample: the results fall around a straight line through the origin. The slope of this line, the ratio of \( I_p \) to clay %, is referred to as the activity of the clay. Clays with high activities, such as the smectites, are termed active clays and exhibit plastic properties over a large range of water content values (Skempton, 1953) whereas those with low activities and short plasticity indices are termed inactive. This so-called activity chart also gives a simple indication of the dominant clay minerals present in the soil.

### 7.2.2 Index property profiles at Bothkennar

In its natural state, Bothkennar clay is a soil of medium to high plasticity. The liquid limit varies considerably with depth, having a maximum value of >80% at around 7m to 10m depth and reducing to a minimum of ~50% both above and below this level. The plastic limit shows much less variation. Figures 7.1(a) and (b) respectively show the profiles of index limits in boreholes HW3/HW9 and HW7/HW8. It is known from earlier work (Hawkins et al., 1989) that the plasticity reduces if the sample is air-dried prior to testing (as recommended in the earlier edition of BS1377: British Standards Institute, 1975), a result that might indicate an effect from organic material. In the present study, this effect was further investigated by treating the samples with hydrogen peroxide (\( \text{H}_2\text{O}_2 \)) in order to destroy organic material prior to testing (cf. Chapter Four, section 4.5.3). Thus, in the present work, the liquid and plastic limits have been determined by tests from natural water content, on air dried and on peroxide treated material. The effects of these treatments are also shown on Figure 7.1(a) and (b) and are summarised in Table 7.1. It can be seen that, as expected, the liquid limits for the air dried and \( \text{H}_2\text{O}_2 \) treated samples show a marked decrease from their corresponding value from natural water content, whereas the plastic limits show only a relatively small change. There is generally a greater decrease in both liquid and plastic limits for the \( \text{H}_2\text{O}_2 \) treated than for the air dried samples, which is interpreted as the result of destroying partially (air drying) or almost completely (peroxide) the organic component of the sediment.

The plasticity chart for boreholes HW3/HW9 and HW7/HW8 is shown in Figure 7.2. The untreated samples plot around the A-line, which is to be expected of silty clay material with variable organic content. The treated samples also lie close to the A-line but it was expected that they would all plot above the A-line indicating an inorganic material. Those that plot below the
Figure 7.1(a) Atterberg limits for boreholes HW3 and HW9
Figure 7.1(b) Atterberg limits for borehole HW7/8
(key as in Figure 7.1(a))
Table 7.1  Summary of Index Limits from the Claret Formation at Bothkennar.

(a) Boreholes HW3/HW9.

<table>
<thead>
<tr>
<th></th>
<th>At Natural w%</th>
<th>Air Dried ¹</th>
<th>H₂O₂ Treated</th>
</tr>
</thead>
<tbody>
<tr>
<td>Liquid Limit</td>
<td>47% – 87%</td>
<td>45% - 74%</td>
<td>36% - 56%</td>
</tr>
<tr>
<td>Plastic Limit</td>
<td>23% – 45%</td>
<td>25% - 42%</td>
<td>20% - 34%</td>
</tr>
<tr>
<td>Plasticity Index</td>
<td>20% – 53%</td>
<td>20% - 32%</td>
<td>14% – 26%</td>
</tr>
<tr>
<td>Activity</td>
<td>0.42 – 1.41</td>
<td>0.45 – 0.85</td>
<td>0.28 – 0.69</td>
</tr>
</tbody>
</table>

¹ These values are for borehole HW3 only, air dried test not done on HW9 material.

(b) Boreholes HW7/HW8.

<table>
<thead>
<tr>
<th></th>
<th>At Natural w%</th>
<th>Air Dried</th>
<th>H₂O₂ Treated</th>
</tr>
</thead>
<tbody>
<tr>
<td>Liquid Limit</td>
<td>50% – 73%</td>
<td>43% - 58%</td>
<td>43% - 62%</td>
</tr>
<tr>
<td>Plastic Limit</td>
<td>27% – 42%</td>
<td>25% - 34%</td>
<td>21% - 35%</td>
</tr>
<tr>
<td>Plasticity Index</td>
<td>23% – 36%</td>
<td>18% - 24%</td>
<td>21% – 28%</td>
</tr>
<tr>
<td>Activity</td>
<td>0.56 – 0.91</td>
<td>0.46 – 0.61</td>
<td>0.52 – 0.62</td>
</tr>
</tbody>
</table>

Table 7.2  Index limit summary statistics subdivided by facies, Boreholes HW3/HW9.

<table>
<thead>
<tr>
<th></th>
<th>Bedded Facies</th>
<th>Mottled Facies</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>LL</td>
<td>PL</td>
</tr>
<tr>
<td>Mean</td>
<td>65.8</td>
<td>34.6</td>
</tr>
<tr>
<td>Median</td>
<td>65.0</td>
<td>35.0</td>
</tr>
<tr>
<td>Standard Deviation</td>
<td>12.70</td>
<td>7.05</td>
</tr>
</tbody>
</table>
Figure 7.2 Plasticity chart for boreholes HW3 (red), HW7/8 (blue) and HW9 (black). Filled circles - tested from natural water content; Open circles - tested after peroxide treatment.
line (which would normally indicate an organic material) could suggest either that the treatment method did not remove all of the organic material, or that the plastic limit determination was too high. A lower plastic limit would have increased the plasticity index and raised the plotted points for the treated samples above the A-line. Since the liquid limit reduced markedly after treatment and it can be difficult to determine the plastic limit reliably, it is thought that the latter explanation is the more probable.

Figures 7.3(a) and (b) respectively show profile of activity for samples from HW3/HW9 and HW7/HW8. Figure 7.4 shows the composite activity chart for these samples. It can be seen that in its natural state the material has an activity of ~ 0.5 to 1.4. This is an unusually large range, which at its upper end is characteristic of an expanding lattice clay. However, the XRD results (Chapter Six, section 6.1) indicate that, although some smectite is present, the clay-sized fraction is dominated by other, inactive minerals and thus the activity seems anomalously high. The reduced activity of the material after peroxide treatment (~0.35 to 0.6) is more consistent with such a composition and it thus seems that it may be the organic component which is responsible for the raised activity in the natural material.

7.3 DISCUSSION OF THE CONTROLS ON INDEX PROPERTIES AT BOTHKENNAR

7.3.1 General

A comparison of the profiles of index properties with those of various compositional parameters indicates that there can be a mutually consistent pattern. Some examples are illustrated in Figure 7.5, which shows the comparative profiles, for boreholes HW3/ HW9, of the index limits, the activity, the percent sand:silt:clay, median particle size, the principal exchangeable cations and the total organic content. It is apparent from inspection that there is a strong positive correlation between the liquid limit and sodium concentration and a weaker negative correlation with median particle size. It is, conversely, much less clear whether there is any mutual variation of the index properties with the total organic content; nor is there any clear correlation of the plastic limit with any of the compositional parameters shown in Figure 7.5.

These correlations can be explored statistically. It is, however, necessary to make some initial caveats. The first is well known: that correlation does not necessarily imply causation, although it will be suggested that the observed correlations do indeed have a physico-chemical basis. Secondly, it is recognised that, when any two parameters are correlated depthwise in a profile, their mutual correlation may be only apparent, the link being their separate correlations with some third, unknown, parameter which also varies with depth. Thirdly, for practical reasons, borehole HW9 only penetrated the more accessible, upper part of the sequence, which may thus
Fig. 7.3(a) Profile of clay mineral activity boreholes HW3 and HW9.
Figure 7.3(b) Profile of clay-fraction activity borehole HW7/8.
Figure 7.4 Activity chart for boreholes HW3, HW7/8 and HW9. Samples tested from natural water content (filled symbols) and after treatment with hydrogen peroxide (open symbols).
Figure 7.5 Comparative profiles of index properties and possible explanatory variables (boreholes HW3 and HW9)
be over-represented in the data set\(^1\). Finally, the sampling positions within the cores were constrained by the need to avoid any visible disturbance and by the need to articulate with those for other geotechnical tests. Thus, they do not represent a sampling pattern designed for statistical purposes. For these reasons the interpretation of the results presented in this section proceeds cautiously and only simple, pairwise, correlations are explored. The inter-dependent, multivariate nature of the dataset is acknowledged but is not explored further.

7.3.2 Variation of index properties with sedimentary facies

Figure 7.6 and Table 7.2 compare the index properties between the bedded and mottled facies. The immediate impression is that the two facies differ little. This is confirmed by the statistical results presented in Table 7.3, which shows that there is no significant inter-facies difference in the variance of any parameter. These results are in general agreement with the inter-facies variation in grading reported in Chapter Six (section 6.3.3), where it was shown that the two facies differ only in their modal size (bedded: 21.9\(\mu\)m, mottled: 17.6\(\mu\)m). On this basis, they would be expected to be similar in their index properties and the slightly coarser mode in the bedded facies would be expected to reduce its plasticity (e.g. Dumbleton & West 1966) but, in this case, has not done so to any significant degree. These results are obviously inconclusive and it is thus necessary to examine more closely the various possible controls on plasticity. Amongst these may be hypothesised the grading, the adsorbed (exchangeable) cations and the nature and amount of the organic content, and these will be examined in sections 7.3.3 to 7.3.5.

7.3.3 Variation of index properties with grading

The relationship between the four principal index properties (liquid and plastic limit, plasticity index and clay fraction activity) and each of three grading parameters (percentage of clay-size, modal grain size and median grain size) is shown in Figures 7.7 to 7.9. The best-fit linear regression line to these data is also shown to indicate the trend. The value of the (Pearson) correlation coefficient for each relationship and its level of statistical significance is stated in Table 7.4.

The results are at first sight surprising: there appears to be no correlation between any index property and any grading parameter. Although there is an apparent visual (inverse) correlation of the liquid limit and median size profiles, this is not statistically significant for this sample. Similarly, the lack of significant correlation between the percentage clay size and any index property is not what is intuitively expected and is apparently counter to most classical studies (e.g. Skempton, 1970).

\(^1\) In order not to bias the dataset further the analysis is restricted to samples from HW3/HW9, since HW7/HW8 also sampled only the topmost four metres of the Claret Formation.
Figure 7.6 Comparison of index properties by facies, borehole HW3/9.
Table 7.3  Comparison of index limits between the bedded and mottled facies.

**t-Test: Two-Sample Assuming Unequal Variances**

<table>
<thead>
<tr>
<th></th>
<th>LL-bedded</th>
<th>LL-mottled</th>
<th>PL-bedded</th>
<th>PL-mottled</th>
<th>PI-bedded</th>
<th>PI-mottled</th>
<th>Act-bedded</th>
<th>Act-mottled</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>65.75</td>
<td>68.47</td>
<td>34.58</td>
<td>32.21</td>
<td>31.17</td>
<td>36.26</td>
<td>0.86</td>
<td>0.86</td>
</tr>
<tr>
<td>Variance</td>
<td>161.30</td>
<td>101.60</td>
<td>49.72</td>
<td>24.51</td>
<td>91.79</td>
<td>44.76</td>
<td>0.09</td>
<td>0.05</td>
</tr>
<tr>
<td>Observations</td>
<td>12</td>
<td>19</td>
<td>12</td>
<td>19</td>
<td>12</td>
<td>19</td>
<td>12</td>
<td>19</td>
</tr>
<tr>
<td>df</td>
<td>20</td>
<td></td>
<td>18</td>
<td></td>
<td>18</td>
<td></td>
<td>19</td>
<td></td>
</tr>
<tr>
<td>t Stat</td>
<td>-0.628</td>
<td></td>
<td>1.018</td>
<td></td>
<td>-1.611</td>
<td></td>
<td>-0.009</td>
<td></td>
</tr>
<tr>
<td>P(T&lt;=t)</td>
<td>0.537</td>
<td></td>
<td>0.322</td>
<td></td>
<td>0.125</td>
<td></td>
<td>0.993</td>
<td></td>
</tr>
<tr>
<td>t Critical</td>
<td>2.086</td>
<td></td>
<td>2.101</td>
<td></td>
<td>2.101</td>
<td></td>
<td>2.093</td>
<td></td>
</tr>
</tbody>
</table>

No significant difference  No significant difference  No significant difference  No significant difference

**F-Test: Two-Sample for Variances**

<table>
<thead>
<tr>
<th></th>
<th>LL-bedded</th>
<th>LL-mottled</th>
<th>PL-bedded</th>
<th>PL-mottled</th>
<th>PI-bedded</th>
<th>PI-mottled</th>
<th>Act-bedded</th>
<th>Act-mottled</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>65.75</td>
<td>68.47</td>
<td>34.58</td>
<td>32.21</td>
<td>31.17</td>
<td>36.26</td>
<td>0.86</td>
<td>0.86</td>
</tr>
<tr>
<td>Variance</td>
<td>161.30</td>
<td>101.60</td>
<td>49.72</td>
<td>24.51</td>
<td>91.79</td>
<td>44.76</td>
<td>0.09</td>
<td>0.05</td>
</tr>
<tr>
<td>Observations</td>
<td>12</td>
<td>19</td>
<td>12</td>
<td>19</td>
<td>12</td>
<td>19</td>
<td>12</td>
<td>19</td>
</tr>
<tr>
<td>df</td>
<td>11</td>
<td></td>
<td>18</td>
<td></td>
<td>11</td>
<td></td>
<td>18</td>
<td></td>
</tr>
<tr>
<td>F</td>
<td>1.588</td>
<td></td>
<td>2.029</td>
<td></td>
<td>2.051</td>
<td></td>
<td>1.769</td>
<td></td>
</tr>
<tr>
<td>P(F&lt;=f)</td>
<td>0.185</td>
<td></td>
<td>0.088</td>
<td></td>
<td>0.085</td>
<td></td>
<td>0.136</td>
<td></td>
</tr>
<tr>
<td>F Critical</td>
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<td></td>
<td>2.374</td>
<td></td>
<td>2.374</td>
<td></td>
<td>2.374</td>
<td></td>
</tr>
</tbody>
</table>

No significant difference  No significant difference  No significant difference  No significant difference
Figure 7.7 Relationship of index limits and related parameters to percentage of clay-sized particles. Blue = untreated, Red = treated samples.
Figure 7.8 Relationship of index limits and related parameters to modal particle size. Blue = untreated, Red = treated samples.
Figure 7.9  Relationship of index limits and related parameters to median particle size. Blue = untreated, Red = treated samples.
<table>
<thead>
<tr>
<th>Parameter</th>
<th>Median size</th>
<th>Modal size</th>
</tr>
</thead>
<tbody>
<tr>
<td>Liquid limit (natural)</td>
<td>Not significant at 95% level (n=31)</td>
<td>Not significant at 95% level (n=31)</td>
</tr>
<tr>
<td>Plastic limit (natural)</td>
<td>Not significant at 95% level (n=31)</td>
<td>Not significant at 95% level (n=31)</td>
</tr>
<tr>
<td>Plasticity index (natural)</td>
<td>Not significant at 95% level (n=31)</td>
<td>Not significant at 95% level (n=31)</td>
</tr>
<tr>
<td>Activity (natural)</td>
<td>Not significant at 95% level (n=31)</td>
<td>Not significant at 95% level (n=31)</td>
</tr>
<tr>
<td>Liquid limit (treated)</td>
<td>Not significant at 95% level (n=21)</td>
<td>Not significant at 95% level (n=21)</td>
</tr>
<tr>
<td>Plastic limit (treated)</td>
<td>Not significant at 95% level (n=21)</td>
<td>Not significant at 95% level (n=21)</td>
</tr>
<tr>
<td>Plasticity index (treated)</td>
<td>Not significant at 95% level (n=21)</td>
<td>Not significant at 95% level (n=21)</td>
</tr>
<tr>
<td>Activity (treated)</td>
<td>Not significant at 95% level (n=21)</td>
<td>Not significant at 95% level (n=21)</td>
</tr>
</tbody>
</table>

Table 7.4: Correlation of index limits with grading parameters: boreholes HW3 and HW9.
Further consideration, however, suggests that this failure to show a clear association may arise for one of three reasons:

- The majority of samples are of similar grading and so any differences due to this cause are likely to be small. These may be masked by errors of measurement, particularly of the plastic limit and of the percentage clay-size, both of which are for different reasons difficult to determine reliably;

- The mineralogy of the clay-sized fraction, which is likely to be a determining parameter, is dominated by low-activity clays and inactive rock flour. Thus relatively small variations in the percentage of clay-sized particles will lead to only small variations in the plasticity index, which may be masked in a single variate analysis by the stronger effect of geochemical variations.

The overall conclusion is that the relatively small variations in grading seen in the Claret Formation at Bothkennar do not cause statistically significant variations in the index properties. This does not, however, imply that these sediments lack the mechanism whereby variations in grading cause variations in plasticity: these would indeed be expected if the sediment were more variable.

7.3.4 Variation of index properties with pore water chemistry

The relationships between the index limits and the concentrations of the major exchangeable cations (sodium, magnesium, potassium and calcium) are shown in Figures 7.10 to 7.13: the values and statistical significance of the correlation coefficients are given in Table 7.5. The results show that in the cases of sodium, potassium and magnesium there is a strong, positive correlation between the weight of cation (per 100g soil) and all the index properties. Some care should be taken in the interpretation of this result, since these cations are also strongly correlated with one another and so there is not necessarily any causative link with every one of them. However, a similar relationship with sodium concentration has been found in the Holocene clays of Belfast Lough (see section 8.5.1) and its is possible that increased sodium concentration raises the liquid limit by increasing the dispersion and thus increasing the specific surface (particle surface area/unit wt).

The data also show that the correlations are changed by hydrogen peroxide treatment. In all cases the correlation with activity and with plastic limit is raised and that with plasticity index is lowered. The results for liquid limit are mixed. Since the purpose of the peroxide treatment is to remove organic materials, it may be supposed that the improved correlations result from the

---

2 The expected mineralogical activity is around 0.5 and so (from Table 6.8) the standard deviation in percentage clay-size of around 4% - 5% would lead to an expected standard deviation in plasticity index of around 2% - 2.5%.
Figure 7.10  Relationship of index limits and related parameters to exchangeable sodium. Blue = untreated, Red = treated samples.
Figure 7.11 Relationship of index limits and related parameters to exchangeable potassium. Blue = untreated, Red = treated samples.
Figure 7.12  Relationship of index limits and related parameters to exchangeable magnesium. Blue = untreated, Red = treated samples.
Figure 7.13  Relationship of index limits and related parameters to exchangeable calcium. Blue = untreated, Red = treated samples.
Table 7.5  Correlation of index limits with exchangeable cations: boreholes HW3 and HW9.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Sodium</th>
<th>Potassium</th>
<th>Magnesium</th>
<th>Calcium</th>
</tr>
</thead>
<tbody>
<tr>
<td>Liquid limit (natural)</td>
<td>0.783</td>
<td>0.755</td>
<td>0.633</td>
<td>Not significant at 95%</td>
</tr>
<tr>
<td></td>
<td>significance &gt;99%</td>
<td>significance &gt;99%</td>
<td>significance &gt;99%</td>
<td>level</td>
</tr>
<tr>
<td>Plastic limit (natural)</td>
<td>0.380</td>
<td>0.402</td>
<td>0.360</td>
<td>Not significant at 95%</td>
</tr>
<tr>
<td></td>
<td>significance &gt;95%</td>
<td>significance &gt;95%</td>
<td>significance &gt;95%</td>
<td>level</td>
</tr>
<tr>
<td>Plasticity index (natural)</td>
<td>0.785</td>
<td>0.731</td>
<td>0.636</td>
<td>Not significant at 95%</td>
</tr>
<tr>
<td></td>
<td>significance &gt;99%</td>
<td>significance &gt;99%</td>
<td>significance &gt;99%</td>
<td>level</td>
</tr>
<tr>
<td>Activity (natural)</td>
<td>0.678</td>
<td>0.402</td>
<td>0.471</td>
<td>Not significant at 95%</td>
</tr>
<tr>
<td></td>
<td>significance &gt;99%</td>
<td>significance &gt;95%</td>
<td>significance &gt;99%</td>
<td>level</td>
</tr>
</tbody>
</table>

| Liquid limit (treated)     | 0.812        | 0.563        | 0.633        | Not significant at 95%   |
|                           | significance >99% | significance >95% | significance >99% | level                   |
| Plastic limit (treated)    | 0.723        | 0.455        | 0.533        | Not significant at 95%   |
|                           | significance >99% | significance >95% | significance >95% | level                   |
| Plasticity index (treated)| 0.667        | 0.508        | 0.550        | - 0.474                  |
|                           | significance >99% | significance >95% | significance >99% | significance >95%       |
| Activity (treated)         | 0.757        | 0.535        | 0.654        | - 0.438                  |
|                           | significance >99% | significance >95% | significance >99% | significance >95%       |

n = 21 in all case
removal of some of the variation associated with these. The reasons why the correlation with plasticity index should be weakened is unclear, although it may arise since this parameter is a function of the liquid limits which shows a variable response. Of note is the fact that treatment reveals a negative correlation of both plasticity index and activity with calcium concentration, a result that may perhaps be explained by the role of calcium in preventing the expansion of the clay lattice.

Overall, the cation data allow the following conclusions to be drawn:

- There is a strong, positive correlation between the index properties and the concentration of each of the cations sodium, potassium and magnesium. This does NOT necessarily imply an equally strong causative link since the cation concentrations are strongly correlated with one another;

- In the cases of the plastic limit and the activity, this correlation is increased after treatment with hydrogen peroxide. It is decreased in the case of the plasticity index and shows no clear pattern in the case of the liquid limit;

- The plasticity index and activity are negatively correlated with calcium after peroxide treatment.

7.3.5 Variation of index properties with organic geochemistry

One implication of the reduction in liquid and plastic limit by hydrogen peroxide treatment is that the organic component of the sediment is in some way instrumental in changing the Atterberg limits. Skempton and Petley (1970) showed that in peaty soils there was a relationship between the total organic content and the Atterberg limits. However, in this present work no significant statistical relationship has been found between the total weight of organic material and either the index properties. This is indicated by Figure 7.14 and Table 7.6. Thus in the Bothkennar sediments the total organic content appears to be a poor predictor of the Atterberg limits. After the treatment with hydrogen peroxide, the reduction in the liquid limit is greater than in the plastic limit, although the magnitude of this reduction is variable and the correlation with the original percentage of total organic material is not statistically significant. Both of these findings suggest that not all the weight of organic material is equally effective in changing the Atterberg limits, possibly because in the sediment some of the material exists as relatively large 'particles' which do not directly interact with the clay minerals.

However, if the variation of individual monosaccharide, nitrogenous and toluene-methanol extractable components with that of the various index properties are considered separately (Figures 7.15 to 7.17), it is found that statistically significant relationships do exist. Table 7.6 shows that in the case of monosaccharide residues there is a positive correlation between the
Figure 7.14 Relationship of index limits and related parameters to total organic content. Blue = untreated, Red = treated.
Table 7.6  Correlation of index limits with organic components: borehole HW9

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Total organic content $^1$</th>
<th>Monosaccharide residues</th>
<th>Kjeldahl nitrogen</th>
<th>Methanol-toluene extract</th>
</tr>
</thead>
<tbody>
<tr>
<td>Liquid limit (natural)</td>
<td>Not significant at 95% level</td>
<td>0.679</td>
<td>0.740</td>
<td>Not significant at 95% level</td>
</tr>
<tr>
<td>Plastic limit (natural)</td>
<td>Not significant at 95% level</td>
<td>Not significant at 95% level</td>
<td>0.765</td>
<td>Not significant at 95% level</td>
</tr>
<tr>
<td>Plasticity index (natural)</td>
<td>Not significant at 95% level</td>
<td>0.712</td>
<td>Not significant at 95% level</td>
<td>Not significant at 95% level</td>
</tr>
<tr>
<td>Activity (natural)</td>
<td>Not significant at 95% level</td>
<td>0.734</td>
<td>Not significant at 95% level</td>
<td>Not significant at 95% level</td>
</tr>
<tr>
<td>Liquid limit (treated)</td>
<td>Not significant at 95% level</td>
<td>0.728</td>
<td>0.792</td>
<td>Not significant at 95% level</td>
</tr>
<tr>
<td>Plastic limit (treated)</td>
<td>Not significant at 95% level</td>
<td>Not significant at 95% level</td>
<td>0.703</td>
<td>Not significant at 95% level</td>
</tr>
<tr>
<td>Plasticity index (treated)</td>
<td>Not significant at 95% level</td>
<td>0.848</td>
<td>0.718</td>
<td>Not significant at 95% level</td>
</tr>
<tr>
<td>Activity (treated)</td>
<td>Not significant at 95% level</td>
<td>0.868</td>
<td>Not significant at 95% level</td>
<td>Not significant at 95% level</td>
</tr>
</tbody>
</table>

$^1$ Also includes borehole HW3 (n = 21)  
  n = 10 in all other cases
Figure 7.15 Relationship of index limits and related parameters to monosaccharide residues. Blue = untreated, Red = treated samples.
Figure 7.16  Relationship of index limits and related parameters to Kjeldahl nitrogen. Blue = untreated, Red = treated samples.
Figure 7.17 Relationship of index limits and related parameters to methanol-toluene extract. Blue = untreated, Red = treated samples.
weight percentage of that component and the liquid limit, plasticity index and activity. The correlation between these parameters and Kjeldahl nitrogen is not significant in the latter two cases, although with this component there is a significant positive correlation with the plastic limit. There is no significant correlation between the toluene-methanol extractable fraction and any component, although this may result from the small size of the dataset and is sensitive to single, individual values (Paul & Barra, 1999).

Table 7.6 also shows that, perhaps unexpectedly, the above correlations persist after peroxide treatment and indeed are stronger in all cases but one. Since the purpose of the treatment is to remove organic material, it would be expected that such correlations would no longer occur: that they do suggests that the peroxide treatment is not completely effective. Furthermore, the fact that the correlation coefficient increases after treatment (as it does in the case of some cations) suggests that, although the absolute value of the index limits has been lowered, the reduction in the scatter of their values has improved the correlation. The underlying mechanism is not clear. It may be that some scatter in the index limits arises from the presence of organic materials that are ‘inefficiently’ attached to the clay particles and that treatment selectively removes these, leaving that material which is more closely associated with the clay surfaces and so continues to modify the index limits. This implies that the inefficient material also modifies the index properties but to a variable extent that is controlled more by the local spatial configuration of the material than by its mass. Some evidence for this is provided by electron micrographs of organic materials which in some cases show sheets or coatings that variably cover the clays on which they are developed (cf. Figure 5.20(b)).

Thus, the following conclusions can be drawn from the organic data:

- There is no significant correlation between the index properties and the total organic content, nor with the weight percentage of the methanol-toluene extractable organic component;

- There is a significant positive correlation between various index properties and the weight percentage of the monosaccharide and Kjeldahl nitrogen components, although the same properties and components are not involved in both cases;

- The strength of the correlations is nearly always increased by treatment with hydrogen peroxide, perhaps due to selective removal of some part of these components.

7.3.6 Summary and overview of Index Properties at Bothkennar

In the more general context, it can be proposed that soil plasticity can be modelled by appropriate variables that describe the grading, the cations and the organic components. This model can be termed the grading-salinity-organics or GSO model. The use of the term salinity
rather than cation acknowledges the fact that some species of cation must always be present on
the clay mineral surfaces and that the enhancement of plasticity is seen in the context of
relatively saline porewater. Figures 7.18 and 7.19 illustrate the GSO model schematically in
terms of the conventional activity and plasticity charts respectively. It is proposed that the
baseline plasticity of a soil is determined by the mineralogy of its clay-sized fraction and that this
is expressed by the clay-fraction activity. The percentage of clay-sized particles then determines
the plasticity index in the manner proposed by Skempton (1953). The location of this baseline is
shown on Figure 7.18 as an envelope and on 7.19 as a small region near the A-line. From this
baseline, the plasticity will develop in terms of each of the GSO components.

On the activity chart (Figure 7.18) the grading component causes the envelope to spread around
the baseline (due to grading within the clay-sized fraction) and to shift downwards due to the
presence of non-clay minerals in the coarser fraction (up to 425μm in the conventional plasticity
tests). On the plasticity chart (Figure 7.19) the grading component causes the region to expand
in the direction above and perpendicular to the A-line, such that the soil plots onto one of a
family of grading-based lines such as the marine clay line (Skempton, 1970) or the till line
(Boulton & Paul, 1976; Trentor, 1999). The salinity and organics components both cause
enhancement of the plasticity relative to the baseline. On the activity chart this is seen as a
movement of the plotted positions upward by some variable amount, the potential movement
due to the organics being the greater. On the plasticity chart this enhancement is seen as a
limited expansion of the baseline area along the direction of the A-line due to salinity alone and
as a larger expansion plus a shift along the A-line when both organics and salinity are involved.

7.4. WATER CONTENT AND BULK DENSITY

7.4.1 General concepts

In a soil mass of uniform density the vertical effective stress will increase linearly with depth.
This in turn will cause the soil to undergo selfweight compression (often termed consolidation\(^3\)
or autocompaction in the geological literature) as the voids reduce in size and water is squeezed
out. The bulk density of a soil is a function of its water content, grain density and degree of
saturation. Knowing\(^4\) the specific grain density \(G_w\) also known as specific gravity), it is possible
to calculate bulk density from the water content using the relationship:

\[^3\]The terms consolidation, compaction and compression tend to be used differently by geotechnical engineers and
geologists. In geotechnical parlance, compression is a permanent reduction of volume under load and consolidation
is the time-dependent process by which it occurs as water is expelled from the soil pores. The term compaction
is reserved for the expulsion of air from a partly saturated soil, often by mechanical means. In geological parlance the
terms consolidation and compaction are both used for the general reduction in pore space that occurs as a sediment is
buried, ages and starts to become a rock. The term compression is seldom encountered.

\[^4\]If \(G_w\) has not been measured, an approximate bulk density can be obtained by assuming a specific gravity of 2.7 for a
clay-rich material.
Figure 7.18  Illustration of the GSO model in terms of the activity chart.

Figure 7.19  Illustration of the GSO model in terms of the plasticity chart.
\[ \rho_b = \rho_w \left[ \frac{G_s + eS_r}{1 + e} \right] \] (7.1)

and substituting for \( e \), where

\[ e = \frac{wG_s}{S_r} \] (7.2)

(where \( \rho_b \) = bulk density, \( \rho_w \) = the density of the porewater, \( G_s \) = specific gravity, \( e \) = void ratio, \( S_r \) = the saturation ratio = 1 for a fully saturated soil).

Thus, conversely to the water content, the bulk density will increase with depth and its profile will be the mirror image of that of the water content.

In structured clay soils, the water content and density profiles follow the theoretical examples in a general manner but usually with several important exceptions in detail. Although the overall envelope of water content values will decrease with depth, any particular profile will exhibit a greater or lesser degree of scatter due to small variations in structure. On the larger scale, mineralogical or facies differences will lead to systematic differences in the rate of reduction with depth. Where there is a proportionately greater silt fraction, the void ratio will be smaller and consequently the water content will be lower, owing to the fact that non-clay minerals (such as quartz) which make up the silt fraction do not exhibit the same open microstructure that is found in flocculated clay minerals.

In many natural deposits, the water table does not extend up to the ground surface. Thus, they have developed a desiccated crust in which the water content is reduced owing to evapotranspiration to the atmosphere. In some natural soft clays the water content profile also deviates considerably from the theoretical curve for some distance below the desiccated crust, either continuing to increase with depth or remaining almost constant. This pattern is combined with corresponding patterns in the undrained shear strength and yield stress (below). Unlike the other parts of the profile, there is no simple mechanical explanation such as selfweight compression or present-day desiccation: instead it seems to represent a zone in which effective stresses were higher in the past and have now reduced, so introducing a degree of overconsolidation. Several explanations can be advanced for this history: the most likely in the case of a geologically young clay sediment is that the water table has fluctuated since deposition. If the sediment is an estuarine clay, then this may be linked to changes in relative sea level as discussed in the previous chapter. This part of the profile may simply be termed the transition zone (Paul et al., 1992a): a term that carries no genetic connotation.
7.4.2 Water content profiles at Bothkennar

The water content profiles for boreholes HW3 and HW7 are shown in Figures 7.20 and 7.21 respectively. It is useful to discuss these profiles section by section in terms of both the local water content gradient, which indicates the local compressibility, and the local scatter of the water content values, which indicates the local heterogeneity of the material.

Figure 7.20 shows that in gross form there is a general increase in water content from the water table down to about 6.0m depth, below which the profile envelope is approximately unchanging down to about 7.2m depth. From this depth down to 17.2m the general trend is one of falling water content, there is then a slight increase between 17.2m and 19.2m depth, after which values fall off markedly to the base of the profile.

The obvious comment to be made is that this water content profile does not conform to that expected for a uniform, normally consolidated clay, since both the slope and the width of the water content envelope vary with depth down the profile. There is also evidence (for example the increasing water content with depth in the upper profile) that the stress history is not one of simple selfweight consolidation and so it is reasonable to suppose that longer-term changes in the depositional environment may be involved. This is a complex problem that is discussed in the next chapter: it is sufficient here just to elucidate the stress history itself.

Inspection of Figure 7.20 shows that a number of generalised zones (A-G) can be identified within the profile. In zones A and B the water content decreases with depth only slowly or not at all, which probably indicates a local change in index properties and correlates with the basal subfacies identified in Chapter Five. In zone C there is a general but punctuated decrease with depth, which is the behaviour expected in a normally consolidated sediment, although there are obviously a number of discontinuities in the profile. Careful comparison with the facies log indicates that these often mark the boundaries between the various units, although care must be taken to allow for the effect of disturbance in some parts of the core. The maximum water content is reached in zone D, over which distance it is noticeably scattered, and in zones E, F and G the water content then decreases upwards until the water table is reached.

Taken as a whole, the shape of the water content profile and width of the encompassing envelopes have the following general features which may be tentatively explained in terms of their sedimentology and depositional history:

- The uppermost part of the profile (down to about 1.2m) consists of a desiccated crust in which the water content increases steadily downwards to the groundwater level (zone G);

- Values between the base of the shell bed (1.8m) and about 4m increase downward (zones E and F). It will be argued later that this pattern is the result of the intertidal and immediately
Figure 7.20  Water content with facies log for borehole HW3. Note demarkation of envelope widths and values outside the main envelopes.
subtidal origin of these deposits, which have suffered wave action and episodic exposure, leading to complex stress histories of greater magnitude than simple selfweight loading. Also, the general decrease in water content upward in this horizon is possibly due to a progressive rise of the upper sediment column within the tidal frame (due to isostatic uplift) and therefore progressively longer exposure under increasingly severe conditions;

- In the main sequence below about 4m there is generally a positive correlation between the width of the profile envelope and the facies type: the mottled facies has a narrower envelope whilst the bedded facies has a much wider envelope. This is due to bioturbation in the mottled facies resulting in a more or less even distribution of silt, whereas the silt in the bedded facies is confined to more discrete horizons;

- Below about 13m the envelope generally becomes narrower despite the fact that the bedded facies is dominant below this depth. This is due to a reduction in void ratio as a result of overburden pressure. The water content of the clay-rich part of the bedded sediment is therefore reduced and therefore similar to lower values of silt-rich horizons;

- The base of the profile (below 19.3m) shows a sharp decrease in water content due to the generally sandier nature of the deposit.

There are a number of extreme values in the profile which are placed some way outside the main envelopes of values. These arise from a variety of causes. In many cases these values are found when there is a rapid, localised change in sediment type, for example where the subsample consisted of silt or silty sand amid generally more clayey material. Where changes in sediment type were rapid but not isolated, for example within the laminated facies, there is a found to be greater scatter of values with a correspondingly wider envelope. These observations show that the variations in scatter of values and width of envelopes down the profile are usually dependent on differences in silt and clay content. The variations can often also be related to facies differences, as the silt content in the mottled facies is usually fairly evenly disseminated through the mottled horizon whereas the silt content in the bedded and laminated facies are normally confined to more discrete horizons. It is clearly not practicable to discuss every extreme point individually and so examples of the above features (numbers 1 to 5 on Figure 7.20) have been selected and are commented upon in Table 7.7.

Figure 7.21 shows the water content profile from borehole HW7. This profile extends to a depth of 6.4m and so encompasses only zones D to G of Figure 7.20. The profile shows broadly the same pattern as that in Figure 7.20, despite the sediments having a higher proportion of the bedded facies and lower overall values of plasticity index (Table 7.1). Thus, the two profiles may represent similar histories in which it is again possible to associate discontinuities in water content with the detailed boundaries between the bedded and the mottled facies.
Table 7.7 Evaluation of a selection of extreme values on the moisture content profile for borehole HW3.

<table>
<thead>
<tr>
<th>Point No.</th>
<th>Depth</th>
<th>Point w%</th>
<th>w% above &amp; below</th>
<th>Explanation</th>
<th>w% validity</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1.32m</td>
<td>61</td>
<td>49, 48</td>
<td>Burrows and holes present within shell bed. Below water table so possible introduction of water during sampling. No other tests done to check determination.</td>
<td>Possibly anomalous</td>
</tr>
<tr>
<td>2</td>
<td>1.83m</td>
<td>64</td>
<td>39, 46</td>
<td>From immediately below shell bed. No other tests done to check determination.</td>
<td>Not known</td>
</tr>
<tr>
<td>3</td>
<td>9.38m</td>
<td>48</td>
<td>61. 69</td>
<td>Sampled on 5-10mm thick silt/fine sand layer. Oxford density trace shows high peak at this depth. No other tests done to check determination.</td>
<td>Valid</td>
</tr>
<tr>
<td>4</td>
<td>16.63m</td>
<td>55</td>
<td>48, 47</td>
<td>Within silty-clay sequence with irregular staining — no obvious sudden change in sediment type. w% determination from a bulk density cell at the same depth was more in line with values above and below.</td>
<td>Probably anomalous</td>
</tr>
<tr>
<td>5</td>
<td>19.71m</td>
<td>46</td>
<td>37, 37</td>
<td>Sample shelly and cracked. May have allowed ingress of water.</td>
<td>Possibly anomalous</td>
</tr>
</tbody>
</table>

Table 7.8 Effect of pore fluid salinity on the calculation of grain density and specific gravity.

<table>
<thead>
<tr>
<th></th>
<th>Salinity (%)</th>
<th>Density (Mg m⁻³)</th>
<th>Mean Gₚ</th>
<th>Mean Sᵣ</th>
</tr>
</thead>
<tbody>
<tr>
<td>Minimum</td>
<td>6.6</td>
<td>1.00107</td>
<td>2.61</td>
<td>0.976</td>
</tr>
<tr>
<td>Weighted mean</td>
<td>17.4</td>
<td>1.01153</td>
<td>2.65</td>
<td>0.989</td>
</tr>
<tr>
<td>Maximum</td>
<td>35.4</td>
<td>1.02508</td>
<td>2.68</td>
<td>0.999</td>
</tr>
</tbody>
</table>

Max error 4 1.3% 1.3%

1 Conversion: salinity = chloride x 1.80655 (conversion factor from Riley & Chester, 1971). This value of salinity is an underestimate by ~0.2%, since chloride alone is not a full measure of chlorinity due to the omission of bromine.

2 Taken from standard tables (Riley & Chester, 1971: table 2.1) for 20°C.

3 From slope and reciprocal of Figure 7.23.

4 This is the maximum error introduced by the use of the single weighted mean value of density for the whole profile. It does not include any error introduced under note 2.
Figure 7.21 Comparative water content log and facies profile for borehole HW 7/8.
7.4.3  Bulk density profiles at Bothkennar

The bulk density was measured using a thin-walled cutting ring of volume 12.25 cm\(^3\) as described in Chapter Four, section 4.5.2.1. This procedure produced not only an independent measurement of bulk density (instead of a value calculated from the water content) but also gave a second measurement of water content from a sample of known volume. This is exploited in the following subsection to obtain a novel method for the evaluation of mean mineral density and mean degree of saturation, parameters which can be rather difficult to measure directly.

Below the water table, a comparison between bulk density and the water content (ring measurement) profiles (Figure 7.22 (a) and (b)) shows that they are close mirror images of each other and generally reflect the same localised variations down the core. This suggests that the soil is close to full saturation. As in the case of the water content profile, the bulk density profile can be split into a number of zones (A to G) of generally consistent trend. A and B have values which generally decrease upwards; the trend of C (the main part of the sequence) and D is generally one of decreasing density upward, although there is a wider envelope of values within the laminated horizons. Sections E and F show increased values of density, probably due to desiccation during intertidal exposure; in F the values are more scattered than in E.

However, there is some evidence that the soil is not 100% saturated. Figure 7.22(c) shows the bulk density profile that can be calculated from the (ring) water contents using equation (7.1) (with \(G_s = 2.65\), see below). If a comparison is made between these values and those actually measured using the ring (shown in Figure 7.22(b)), it can be seen that there are a number of small discrepancies. Although these might be partly attributable to measurement error, there is also some reason to believe that the sediment is not fully saturated. For example, immediately upon coring, the fresh sediment smells strongly of hydrogen sulphide; methane would also be expected in view of the anaerobic character of the subsurface. Furthermore, close examination of the fabric reveals pinhole voids, sometimes at the centre of mottles (as noted by Hawkins et al., 1991) which are indicative either of the presence of gas or are relics of the original burrow voids.

7.4.4  A note on the determination of mean soil saturation and mean mineral density

The degree of saturation and mineral density are both difficult to measure directly with high accuracy. However, their mean values (over the whole core) can be obtained from equation (7.3) below, since re-arrangement of equation (7.1) yields:

\[
\rho_s \frac{1 + w}{\rho_b} = \frac{1}{G_s} + \frac{w}{S_r}
\]  

(7.3)
Figure 7.22 (a) Water content and (b) bulk density profiles; (c) derived bulk density (see text), for borehole HW3.
Figure 7.23 shows the graph of this expression with the right-hand side term plotted against water content. Equation (7.3) shows the slope of a linear regression line to be \( \frac{1}{\rho_w S_r} \) and its y-axis intercept to be \( \frac{1}{\rho_s G_s} \). This gives a very convenient method of calculating the mean values of these parameters.

From Figure 7.23 the mean particle specific gravity \( (G_s) \) is calculated as 2.65 and the mean saturation ratio \( (S_r) \) as 0.989. Both these values appear to be reasonable (although \( G_s \) is a little low) in view respectively of the mineral composition described in section 6.2 and the evidence of gas described earlier. In this calculation, a value is required for the pore fluid density, which is a function of the salinity, and thus actually varies down the profile. However, the effect of this variation is small and, if a constant, weighted mean value of 1.01153 Mg m\(^{-3}\) is used the consequent error in both density and degree of saturation is around ±1.3% at most (Table 7.8).

### 7.5 LIQUIDITY INDEX

#### 7.5.1 General concepts

The profiles discussed thus far have been ‘raw’ profiles, in the sense that the parameter values have been plotted directly against depth without any attempt to normalise them against other parameters. This is not always a satisfactory approach, since the variation in a raw parameter may be the consequence of several factors, whose interplay is not apparent just from inspection. For example, some of the changes in water content down a profile are the result of compositional variations that are reflected by the liquid and plastic limits. It is possible to remove their effect by comparing the water content against the liquid and plastic limit via a parameter termed the liquidity index \( (I_L) \).

The liquidity index relates the natural water content of a soil to the interval between the liquid and plastic limits as a proportion of the plasticity index:

\[
I_L = \frac{w - w_p}{w_L - w_p} \quad (7.4)
\]

At the liquid limit \( I_L \) equals one and at the plastic limit it equals zero. In the intact state the value of the liquidity index clearly reflects the level of structure in the soil: those in which \( I_L > 1 \) have an open structure that is stabilised in some way, since in the remoulded state they would be a slurry. In many estuarine clays there occurs a point, deeper than the desiccated crust, below which the \( I_L \) decreases with depth. It has been found in many such profiles (see for example Chapter Eight) that, at this point, the liquidity index is around unity or a little above and that it
Figure 7.23  Relationship of bulk density and water content in borehole HW3.
then decreases logarithmically with depth. The rate of reduction is such that a liquidity index of 0.5 is reached a depth of 10m to 30m in most cases (Skempton, 1970: fig. 22). The use of the liquidity index is also valuable because of its connection with the undrained shear strength, as will be discussed below.

7.5.2. Liquidity index profiles at Bothkennar

Figures 7.24(a) and 7.25(a) show respectively the liquidity index profiles for boreholes HW3 and HW7. Compared with the corresponding water content profiles, the profiles for both boreholes show much less variability and the water content zonation A-D cannot be recognise as a result. In part, of course, this arises from the use of fewer data points (only those positions are plotted at which the index limits were tested); however, there is also considered to be a genuine reduction in scatter. In HW3 the trend of the liquidity index below about 7m - 8m depth indicates simple selfweight compression, with a reduction from a value close to unity at around 7.5m to a value around 0.5 at 20m depth, which is consistent with the picture presented by Skempton (1970). This suggests that the local variability in water content simply follows changes in index properties and that the underlying stress-controlled change is more simple.

In HW7 the liquidity index remains almost constant between the level of the Cerastoderma bed at around 2m depth down to a depth of ~7m, and in HW3 there is a slight decrease over this interval. This contrasts in both cases with the water content, which increases with depth over the same interval in both boreholes. This contrast implies both that the water content profile is dominated by changes in index limits and also, since the liquidity index does not increase with depth, any imposed inter-tidal stress history has not strongly overprinted the selfweight trend below a depth of around two metres.

7.6 Voids INDEX

7.6.1 General concepts

The use of the liquidity index brings a disadvantage, in that, although the water content is directly controlled by the compressibility, this parameter is not introduced explicitly, but only indirectly through its relationship with both \( w_i \) and \( I_r \). For this reason Burland (1990) introduced the concept of the void index \( (I_v) \), a parameter directly analogous to the liquidity index, in which the current void ratio \( (e) \) of a sample is expressed as a proportion of the interval between two defined values \( e_{1000} \) and \( e_{100} \), these being the void ratios of the soil in the remoulded state at vertical effective stresses of 100kPa and 1000kPa respectively:

\[
I_v = \frac{e - e_{100}}{e_{1000} - e_{100}} \quad (7.5)
\]
Figure 7.24  Profiles of (a) liquidity index and (b) void index: borehole HW3
Figure 7.25 Profiles of (a) liquidity index and (b) void index for borehole HW7/8.
The interval from $e_{100}$ to $e_{1000}$ is equal to the compression that occurs in the remoulded state due to a 10-fold increase in stress, which Burland (1990) has termed the *intrinsic compressibility*.

Re-examination of Skempton's (1970) data for structured marine clays has shown (Burland, 1990) that when $I_v$ is plotted against vertical effective stress all these clays lie in a narrow band that has been termed the *sedimentation compression line* (SCL). In the remoulded state, the same clays follow a different line termed the *intrinsic compression line* (ICL). The distance from the ICL of the void index of any in situ sample can be taken as a measure of geotechnical structure. If the void indices down a given profile from a soft, structured clay are shown as a connected series of points it is usually found that they fluctuate around the SCL, indicating that the level of soil structure varies in detail, presumably as the result of small scale variations in sedimentary fabric. Work by Burland (1990) and by Hight *et al.* (1992a) has shown that the Bothkennar clay follows the sedimentation compression line, although there are strong departures at around 7m depth and at around 12m depth, which have been attributed to changes in depositional conditions. These observations will be discussed further in Chapter Eight.

### 7.6.2 Void index profiles at Bothkennar

Figures 7.24(b) and 7.25(b) show respectively the void index profiles for boreholes HW3 and HW7. The profile from HW3 broadly follows the SCL from a depth of about three metres downwards. This suggests that below this depth the sediment is both geotechnically structured (since it follows the SCL) and is normally consolidated, presumably as a result of selfweight compression. It is probable that the points to the left of the SCL arise from sample disturbance. Above a depth of three metres one point lies to the right of the SCL, indicating a more open structure, and at still shallower depths the profile then falls increasingly to the left of the SCL, indicating a more compact structure. This pattern may arise from the (believed) intertidal origin of these sediments: the more compact structure is evidence of overconsolidation due to desiccation in the intertidal zone.

The profile from HW7 (Figure 7.25(b)) shows a more complex pattern. Below about two metres depth there appear to be three similar subunits, each with a consistent pattern in which the void index and, hence, the level of geotechnical structure decreases upwards over a distance of 1m to 1.5m. This pattern is not coincident with the disturbance caused by sampling nor does it correspond to the visible facies transition. The steady upwards reduction within each unit indicates an increasing degree of mild densification, perhaps associated with some repetitive processes within the intertidal zone, although the points remain close to the SCL and so the sediment cannot be said to be strongly overconsolidated.
7.7 COMPRESSIBILITY AND YIELD STRESS

7.7.1 General concepts

Compression is the reduction in water content that occurs under increased effective stress. If water content or void ratio is plotted against the logarithm of vertical effective stress, a compression or e-log p (generally referred to as the 'e-p') curve is obtained. This consists of an initially shallow curve which bends at the approximate overburden pressure (\(p_u\)) to form the virgin consolidation curve, which is approximately a straight line until some limiting stress is reached and the soil structure breaks down. As the name suggests, the curve shows the response of the soil to increases of load beyond those previously sustained during its history and so, in the case of a natural deposit, represents the compression of a clay due to the addition of more material during the sedimentation process. The effective vertical yield stress\(^5\) is the stress at which there is a change from elastic to plastic behaviour. This is seen as a change of slope on the void ratio/pressure curve and is interpreted as the maximum vertical stress imposed on a point in the soil column in the past. The yield stress ratio is the ratio of the effective vertical yield stress to effective overburden pressure (which is the weight of soil minus the weight of water above the point). If this ratio equals one then the soil is said to be 'normally consolidated' and at no time in its history was the clay subjected to pressures greater than the existing overburden pressure. Thus, the yield stress occurs at an effective stress approximately equal to the in situ value.

If the yield stress exceeds the current in situ stress the yield stress ratio is greater than one and the soil is said to be overconsolidated. Thus it was, at some time in its history, subjected to pressures greater than the existing overburden pressure. A geologically young clay can become overconsolidated by a number of processes, for example the removal of material by erosion or by drying in the intertidal zone or in the upper part of the soil profile to form a desiccated crust. The vertical yield stress does not revert back to its original value when the pressure is relieved and in the case of an increase in yield stress due to drying, the soil does not recover if the water content is subsequently raised. In reality, most natural soft clay deposits will show some degree of overconsolidation in the uppermost layers and, therefore, the profile will show a combination of a normally consolidated and an overconsolidated profile\(^6\) as indicated by Skempton (1948).

\(^5\) In the past yield stress was often referred to as 'preconsolidation pressure'. Burland (1990) pointed out that the yield stress could include a component of 'increased resistance to compression due to ageing' and the term 'preconsolidation pressure' should be 'reserved for situations in which the magnitude of such a pressure can be established by geological means'. Yield stress should more correctly be referred to as 'vertical yield stress'. The ratio between vertical yield stress and the effective overburden pressure should thus be termed the 'yield stress ratio' and only referred to as the 'overconsolidation ratio' when the stress history is known.

\(^6\) It is necessary to study the geotechnical properties of a deposit in conjunction with a visual study in order to be able to say whether overconsolidation is due to erosion or to intermittent exposure in the intertidal zone, since sedimentological and biological evidence need also to be taken into account. The presence of a desiccated crust is usually obvious from visual inspection since weathering usually produces easily recognisable changes in the visible profile and its geotechnical character.
It should be noted that the yield stress can incorporate a component which is over and above the present in situ effective stress without there having been any increase in past load. This was formerly described as an apparent pre-consolidation pressure and led researchers to seek explanations in terms of stress history. It was however, shown by Bjerrum (1967), that a rise in yield stress ratio can occur simply as a result of the re-arrangement of the clay structure under constant load, a process termed secondary consolidation. Although this behaviour often gets scant mention in geotechnical texts, by comparison with primary consolidation which is usually given an extended mathematical treatment, it is in fact of great significance in natural clays and appears to be a normal consequence of their ageing.

7.7.2 Yield stress profiles at Bothkennar

Profiles of yield stress and yield stress ratio (YSR) from borehole HW7 are shown on Figures 7.26(a) and (b). Figure 7.26(a) also shows for comparison the in situ effective stress profile reported by Hight et al. 1992a. From the effective stress profile, the YSR has been calculated for each data point (Figure 7.26(b)). Below a depth of about 3m, the yield stress ratio is constant at around 1.6: this conventionally indicates that the deposits here are lightly overconsolidated. The fact that the ratio is constant with depth also indicates that it cannot be the result of post-depositional erosion, since this would lead to a YSR decreasing with depth towards an asymptotic value of unity. Between 3m depth and the base of the shell bed, the YSR increases to about 2.5. These values are similar to those obtained by Nash et al. (1992a), which are also shown for comparison on Figure 7.26.

This pattern of YSR may have arisen from a number of inter-related processes. Subaerial exposure during the tidal cycle will cause desiccation and the development of suction stresses. These may be expected to be greater at the higher levels in the tidal frame (i.e. higher in the profile) due to the longer period of exposure. On this basis inter-tidal effects appear to cease below around 3m depth. The relatively constant YSR below this depth does not relate easily to normal concepts of overconsolidation: however, similar values have been reported from 'normally consolidated' estuarine sediments in other areas (Bjerrum, 1967; Crooks & Graham, 1976; Hawkins et al., 1989), as discussed in Chapter Eight (cf. Table 8.4). These authors suggest various reasons for the elevated YSR, including secondary consolidation and cementation by both organic and inorganic materials, all of which processes may have operated at Bothkennar.

7.8 UNDRAINED SHEAR STRENGTH

7.8.1 General concepts

Undrained shear strength ($S_u$) is a measure of the resistance of a soil to failure in shear under a condition of constant water content (Head, 1992; Craig, 1997). Undrained shear strength can be
Figure 7.26  Profiles of (a) yield stress and (b) yield stress ratio (YSR) from the upper part of the sequence at Bothkennar.
measured by several methods: the triaxial cell, fall cone and rotating vane (either *in situ* or laboratory vane) are those normally employed. These methods are described in Head (1992). There are some differences between the results obtained by the different methods due to the typical size of the samples involved, the extent of sample disturbance inherent in the test and the distribution of the stress field it imposes. For the present purpose these differences are not great and will not be considered further unless they are pertinent to the discussion.

In a normally consolidated soil the undrained shear strength increases linearly with depth i.e. the undrained shear strength ($S_u$) increases proportionally with the effective overburden pressure (vertical effective stress: $p_e$): Bjerrum (1967) described this as “one of the most characteristic properties of normally consolidated clays”. A plot of $S_u$ against $p_e$ is thus equivalent to a $S_u$/depth profile. The slope of the plot ($ds/dp_e$) ranges from 0.2 to 0.5 for many clays (Skempton, 1957). This value is dependent on the plasticity index of the clay (Skempton, 1957): if, for different clays, $S_u/p_e$ is plotted against plasticity index ($I_p$), the points are found to fall on a single line, indicating that the greater the plasticity, the higher the shear strength. In the remoulded state, there is also a close relationship between undrained shear strength and liquidity index (Skempton & Northey, 1953). This is to be expected since the undrained shear strength is controlled by the *in situ* water content and this is related to both the plasticity index and liquid limit.

In the intact state the relationship follows the same general form but at any given value of liquidity index the strength is increased by a factor termed the sensitivity of the soil, which is the ratio of the undrained shear strength in the intact state to that in the remoulded state. Sensitivity is attributed to the presence of bonds between clay particles, whose breakage leads to the collapse the clay structure. Extremely high sensitivity levels are found in soils known as quick clays, which occur in areas such as Norway and eastern Canada, where they present serious geotechnical hazards. The cause of this condition is generally accepted to be a metastable structure in which the clay is maintained at a very high liquidity index, a condition which arises due to post-depositional reduction in the liquid limit following leaching of originally saline pore fluid. The depositional water content remains unchanged and so the soil is brought to a very high liquidity index at which its remoulded strength is negligible, thus leading to the extremely high sensitivity.

### 7.8.2 Undrained shear strength profiles at Bothkennar

The profile of undrained shear strength (intact and remoulded) from borehole HW3 is shown in Figure 7.27(a) and the sensitivity profile in Figure 7.27(b). Below about 4m depth the intact strength increases linearly with depth. The envelope of values is generally narrow, although
some scatter results from facies variations which account for the minor breaks. The profile can
be divided into three main sections that correspond to the main sequence of silty clays (A), the
transition zone (B) and the desiccated crust (C) (including the shell bed). The main sequence
has a general increasing trend with depth which is consistent with a normally consolidated
deposit. Values increase from a minimum of about 18kPa at 4m depth to 50kPa at 20m; in
detail, however, many small variations and hiatuses are present, which can be matched to local
sedimentological features and hence are often correlated with facies changes. Within zone (B)
the values do not vary greatly with depth, which behaviour is consistent with a lightly-
overconsolidated material. In the crust, the strength increases rapidly upwards to nearly 100kPa
at surface, probably as a result of an increased effective stress due to partial saturation.

In Figure 7.27(a) it can be seen that in the main sequence, the remoulded shear strength shows a
slight increasing trend with depth down to about 14.3m, where the value is 9kPa. Between this
depth and about 17.0m the profile is fairly uniform (between about 4kPa and 5kPa) and below
17.7m values range between 3.5kPa and 13kPa. The sensitivity is typically between 5 and 10,
and the form of the profile largely follows the detail of the intact strength profile.

In a fully remoulded clay soil the undrained shear strength is largely determined by the liquidity
index, as shown originally by Skempton and Northey (1953). Although some variation exists
between soils of different geological origin (as shown in Figure 7.28), any one soil type follows a
smooth curve when its remoulded undrained shear strength $S_u$ is plotted against liquidity
index $I_L$. Figure 7.28 shows this to be the case at Bothkennar and that the removal of the organic
material does not alter this basic relationship above $I_L = 0.99$. Below this value, the untreated
soil retains some additional strength which suggests that the organic component acts as a
cement that may either partially survive remoulding or has the ability to partially reform (i.e. a
form of thixotropic hardening) afterwards. In the intact state there is also a relationship
between liquidity index and undrained strength although there is considerably more scatter in
this case.

7.9 SUMMARY

In its natural state Bothkennar clay is a soil of medium to high plasticity and is more plastic than
expected from its clay mineralogy. The index properties differ little between the bedded and
mottled facies. There is a strong, positive correlation between the index properties and the
concentration of the sodium cation. Although they correlate only poorly with the total organic
content, when the organic material is characterised in terms of separate monosaccharide,
nitrogen and methanol-toluene extractable components, the liquid limit shows a statistically
Figure 7.27 Profiles of (a) undrained shear strength and (b) sensitivity for borehole HW3.
Figure 7.28  Relationship of undrained shear strength to liquidity index in the intact and remoulded states.
significant positive correlation with the first two of these components and the plastic limit shows a statistically significant positive correlation with the nitrogen component.

In gross form there is a general increase in water content from the water table down to about 4m depth, below which the profile envelope is approximately unchanging down to about 7.2m depth. From this depth down to 17.2m the general trend is one of falling water content, there is then a slight increase between 17.2m and 19.2m depth, after which values fall off markedly to the base of the profile. This profile does not conform to that expected for a uniform, normally consolidated clay but the shape of the water content profile and the width of the encompassing envelopes can be explained in terms of their sedimentology and depositional history. The evidence is that the stress history is not one of simple selfweight consolidation and so it is reasonable to suppose that longer-term changes in the depositional environment may be involved. Profiles of liquidity index show much less variability than corresponding water content profiles, which suggests that the local variability in water content is dominated by changes in index properties and that the underlying stress-controlled change in normalised water content is more simple.

The void index profile follows the SCL fairly closely downwards from a depth of about three metres. This suggests that below this depth the sediment is both geotechnically structured (since it follows the SCL) and is normally consolidated, presumably as a result of selfweight compression. Above this depth there is some indication of a more open structure, and at still shallower depths the profile then falls increasingly to the left of the SCL, indicating a more compact structure. This pattern may arise from the (believed) intertidal origin of these sediments: the more compact structure is evidence of overconsolidation due to desiccation in the intertidal zone.

Profiles of yield stress and yield stress ratio show that between 3m depth and the base of the shell bed the value of the yield stress ratio can be as high as 2.5 or greater. Below about 3m it fluctuates at around 1.6, which conventionally indicates that the deposits here are lightly overconsolidated. This pattern may also have arisen from a number of inter-related processes. Subaerial exposure during the tidal cycle will cause desiccation and the development of suction stresses that may be expected to be greater at the higher levels in the tidal frame. The relatively constant value below this depth does not relate easily to normal concepts of overconsolidation: however, similar values have been reported from other 'normally consolidated' estuarine sediments and have been explained variously by secondary consolidation and cementation by both organic and inorganic materials.

The profile of undrained shear strength can be divided into three main sections that correspond to the main sequence of silty clays, the intertidal zone and the desiccated crust. The main
sequence has a general increasing trend with depth which is consistent with a normally consolidated deposit. Within the intertidal zone the values do not vary greatly with depth, a behaviour consistent with that of a lightly-overconsolidated material. In the crust the strength increases rapidly upwards, probably as a result of an increased effective stress due to partial saturation.
CHAPTER EIGHT

DISCUSSION:
DEPOSITIONAL ENVIRONMENT AND CHARACTER OF
THE CLARET FORMATION AT BOTHKENNAR

8.1 INTRODUCTION

8.2 A WATER DEPTH MODEL FOR THE CLARET FORMATION
AT BOTHKENNAR

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CHAPTER EIGHT

DISCUSSION

DEPOSITIONAL ENVIRONMENT AND CHARACTER OF
THE CLARET FORMATION AT BOTHKENNAR

8.1 INTRODUCTION

One of the objectives of this work has been to relate the physical character of the Bothkennar clay to its depositional environment and, if possible, to draw conclusions of general relevance to other soft clay deposits. This chapter brings several of these ideas together. The chapter first considers the conditions under which the sediments were deposited (the depositional model) and secondly considers the influence this has had on their character. Of central importance is an analysis of the water depth history at Bothkennar and its relationship to the sedimentary facies and their stratigraphic architecture. This results in an overall depositional model which is used as a framework within which to evaluate the geotechnical profiles described in Chapters Three and Seven. Finally, the results from Bothkennar are compared with those known from three other Holocene soft clay sites in order to illuminate some significant factors that control the engineering geology of such deposits more generally.

8.2 A WATER DEPTH MODEL FOR THE CLARET FORMATION AT BOTHKENNAR

The approach taken is first to develop a model of the changes in water depth at Bothkennar during the depositional period. This has been obtained by a comparison of a sea-level curve (modified from those available in the literature) with an analogous sea-bed curve, obtained by radiocarbon dating points on the Bothkennar profile. This section demands a rather complete treatment of the errors involved in both the sea-level and seabed data: in particular it reveals, and rectifies, a number of previously unrecognised problems with the most appropriate of the published sea-level data.

8.2.1 The Robinson (1993) Sea-level Curve

The proposed analysis requires an accurate curve of relative sea-level during the Holocene. It is fortunate that in this respect the Forth is one of the best-studied regions in Britain. As discussed in Chapter Two (section 2.4), this work allowed the publication of a relative sea-level curve for the western Forth in the vicinity of Stirling (Sissons & Brooks 1971; Robinson, 1993: cf. Figure 2.9) for the early to middle Holocene. In a major paper, Robinson (1993) examined the Holocene

1 Publication of these results was somewhat delayed, the fieldwork having been conducted in 1977-80.
development of the Forth valley and estuary from Flanders Moss to Grangemouth over the time interval from about 10,500 to 2,900 $^{14}$C yrs BP. Integral to that paper were sea-level curves for the western and eastern Forth valley. Her curve for the eastern area is the one appropriate to Bothkennar and is shown in Figure 8.1.

Robinson was able to define her sampling locations accurately in relation to visible and buried geomorphological features, using the detailed results obtained previously by Sissons and his co-workers (cf. Tables 2.2 and 2.3). In addition to the radiocarbon dates, she also presented a comprehensive discussion of the palaeoenvironments of her index points based on microfossil (pollen and diatom) analyses, since, by studying the changes in both pollen and diatom species above and below dated samples, it was possible to determine when and where freshwater/brackish transitions occurred and so relate the dated sampling position to the contemporaneous shoreline.

For all these reasons Robinson's (1993) curve for the eastern Forth is considered to be the best available for the Bothkennar area and has been used as a basis for the analysis presented here. Since the samples from Bothkennar were all dated as being younger than around 5,000 BP\(^2\) (section 8.2.3.1) only that part of the curve post-dating the mid-Holocene transgression need be considered here. This encompasses Robinson's index points 9 to 12, which are analysed in detail in the following subsections in order to identify possible sources of uncertainty in this part of her sea-level curve.

8.2.2 Corrections to the Robinson (1993) Sea-level Curve

Unfortunately, the Robinson (1993) curve suffers from three defects in the methodology used to locate the index points, both in elevation and with respect to their radiocarbon ages, as a result either of the omission of possible sources of error or of the procedure adopted in constructing the curve itself. They may, in consequence, have introduced small but significant errors, since the change in relative sea-level during the time interval under study is itself not large. It is therefore important to quantify them and, if possible, to correct for them. They have been acknowledged by Dr Robinson (pers. comm.) who has kindly supplied additional data to allow them to be addressed.

The three defects are:

- Inappropriate plotting of the index points with respect to the mean $^{14}$C date;

\(^2\) All ages are simply quoted as years BP, from which it is understood that these are uncalibrated $^{14}$C dates, corrected by a reservoir age of 405±40 years in the case of shell samples.
• Failure to correct the measured sample elevation for post-depositional compression (autocompaction);

• Possible inappropriate location of the mean sea level curve relative to the index point.

Each of these defects will be considered in turn for the relevant index points (9 to 12), the size of the error will be evaluated and a possible correction suggested.

8.2.2.1 Correction with respect to mean radiocarbon age
Robinson (1993) showed the positions of her index points using arrows whose length represented half of the tidal range, assuming a tidal range similar to that of the present day at Grangemouth (5.2m). Figure 8.1 shows the index points and sea-level curve as published by Robinson. The base of each arrow is at the measured altitude of each sample and is taken to be MLWST, the top of the arrow is at mean sea level (MSL). The sea-level curve itself was drawn half-way between MSL and MLWST.

However, if the co-ordinates of the index points on Figure 8.1 are compared with the mean 14C ages given in her paper, it is found that the published points were not actually plotted at the mean date. By redrawing the arrows at the actual dated position with a base width of two standard deviations (red arrows), it can be seen Figure 8.2(a) that the original (green) arrows were actually plotted at the extreme youngest position on the two standard deviation error bar. This was confirmed by Robinson (pers. comm.) although it was not stated in her paper.

The first modification to Robinson’s (1993) sea-level curve, therefore, has been to move the plotted sample positions to the left so that the red arrows (with a base width of two standard deviations) appear at the dated position, whilst remaining at the same altitude. This has the effect of lowering the sea-level curve and is represented by the red curve in Figure 8.2(a) 3.

8.2.2.2 Correction for post-depositional compression
Post-depositional selfweight compression (often termed autocompaction) introduces an error in the elevation of a sample point which depends on the thickness of the complete sequence, the relative position of the sample point within that sequence, the geotechnical properties of the sediment below the sample point and any history of later loading. The significance of this error clearly depends on the relative magnitude of the sea-level change. Thus it may be of greater significance in estimates of late-Holocene sea levels than in those of the Late Devensian or early

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3 It should be noted that the Robinson index point 8 has been removed from this and subsequent sea-level curves as it represents a shell sample incorporated into the raised beach and underlying a shell midden and its position within the tidal range is not well-constrained.
Figure 8.2(a) Replot of Figure 8.1 to correct for the position of the arrows with respect to radiometric age.
Holocene. Also, the significance of the error will be greater in those situations where the tidal range is small and/or the sample position is well constrained within the tidal cycle.

The process of autocompaction can be analysed quantitatively using standard geotechnical theory. It has previously been shown (Paul & Barras, 1998) that, for the simple case of self-weight compression, at mid-depth in a layer the correction required to recover the depositional elevation is around 5% - 10% of the layer thickness for compressible sediments and around 1% - 2% for incompressible ones. In a geological context this suggests that the effect will usually be more important for sites closer to the local depocentre, where thicker sequences of finer sediment are likely to occur, rather than at basin margins where coarser, thinner sequences are likely to predominate.

Robinson (1993) plotted her sea-level curves using the present-day elevations of the sample positions. However, since all of the samples were collected from below the ground surface the sediment below each sample would have suffered some compression due to the subsequent deposition of the overlying material. The sample elevations were not corrected for this compression by Robinson herself and so this has been done by the present writer, using the method previously published by Paul and Barras (1998). The results are shown in Table 8.1. Since there were no geotechnical data available for the sites in question, it was necessary to estimate these data using plausible geotechnical assumptions and by inference from published regional patterns. These were aided by unpublished lithological profiles produced by J.B. Sissons and kindly made available by Dr Robinson (pers. comm.). From these lithological profiles likely geotechnical profiles were deduced for each of the sample sites, using the standard Type Profiles of Gostelow and Browne (1986). The depths to the upper surface of the underlying glacial till, (taken as an incompressible basement) and to the incompressible Bothkennar Gravel Formation were also estimated from Gostelow and Browne (1986) supplemented by Sissons (1969).

The second correction, therefore, has been to decompact the sediment profile below Robinson’s sample positions to recover their original depositional elevation.

The revised positions of the index points (magenta arrows) and the effect of these adjustments on the sea-level curve are shown on Figure 8.2(b). It can be seen that the red curve (radiocarbon age correction) is raised above the original (green) curve, resulting in the magenta curve. The corrections are relatively small (between 0.4m and 0.8m, Table 8.1) due both to the relatively shallow depth of burial and, in the cases of profile types 4 and 5, to the coarser and thus less compressible nature of the deposits compared with Bothkennar.
Table 8.1 Autocompaction corrections for Robinson (1993) Index Points 9-12.

<table>
<thead>
<tr>
<th>Location</th>
<th>Sample number and age $^{14}$C yrs BP$^1$</th>
<th>Inferred$^2$ geotechnical profile</th>
<th>Elevation of top of borehole (m OD)$^1$</th>
<th>Sample elevation (m OD)$^1$</th>
<th>Correction (m)</th>
<th>Corrected sample elevation (m OD)</th>
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<tr>
<td>Point 9</td>
<td>SRR-1899</td>
<td>Type 5</td>
<td>+5.25</td>
<td>+4.4</td>
<td>+0.4</td>
<td>+4.8</td>
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<td>Mid Thorn NS 908 809</td>
<td>4830±50$^a$</td>
<td>4.75m PG silty clay</td>
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<td></td>
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<td>0.25m gravel</td>
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<tr>
<td></td>
<td>4235±65$^c$</td>
<td>5.0m LG sandy clay</td>
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<tr>
<td></td>
<td></td>
<td>Glacial till</td>
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<td>Point 10</td>
<td>SRR-1540</td>
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<td>+4.27</td>
<td>+2.92</td>
<td>+0.6</td>
<td>+3.52</td>
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<td>3415±65$^c$</td>
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<td>+2.62</td>
<td>+0.8</td>
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<td>3560±90$^b$</td>
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<td></td>
<td>3155±100$^c$</td>
<td>10.0m LG sandy clay</td>
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<td>Glacial till</td>
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<td>Point 12</td>
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<td>+0.4</td>
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<td>Glacial till</td>
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</table>


$^2$ Gostelow and Browne (1986); PG = Postglacial; LG = Late-glacial.

$^a$ Outer age. $^b$ Inner age. $^c$ Corrected for apparent age of seawater: 405±40 yrs BP (Harkness, 1983).
Figure 8.2(b)  Replot of Figures 8.1 and 8.2(a) incorporating radiometric age and compaction corrections.
8.2.2.3 Correction with respect to tidal range

As stated earlier, the up arrows on Figure 8.2(b) (index points 9 to 12) represent the present-day tidal range from Mean Low Water Spring Tide (MLWST) to Mean Sea Level (MSL), which at Grangemouth is approximately 2.6m (i.e. half the present-day range of 5.2m) if the tidal range is assumed to be symmetric (Robinson, 1993). On her published (1993) curve the index points were plotted at the bottom of the arrows (i.e. at MLWST) but could in principle lie anywhere within the tidal range, depending on the ecology of the organism concerned. This distinction is of particular importance if, as in this setting, the tidal range and relative sea-level change are of similar magnitudes. In practice, intertidal organisms tend to inhabit the zone below MSL to avoid prolonged exposure and so index points based on in situ shell material are likely to be located between MLWST and MSL.

Figure 8.3 shows three possible reconstructions (MSL1, MSL2 and MSL3) based on the possible positions of MSL with reference to the Robinson index points (Robinson, 1993). The up arrows at the index points have been extended downwards to encompass the whole tidal range and the possible position of the sea-level curve is represented by the shaded error box, which encompasses the two standard deviation $^{14}$C error on the time axis and the tidal range on the elevation-axis in a manner suggested by Robinson (pers. comm.; also discussed by Haggart, 1989).

The two extreme reconstructions (MSL1 and 3) respectively assume that no sample lay below MLWST, or above MHWST. MSL1 (the highest theoretical position for the contemporaneous MSL at 2.7m above the index point elevation) would project the curve to ~2.90m AOD at present day. This is about 2.55m above the present-day MSL and so would appear to be too high, since it would imply an unrealistically rapid reduction in the rate of fall of sea-level from 2,900 BP to the present, whereas it is believed that the rate had probably greatly reduced by about 2,000 years BP and so the change between 2,000 years BP and the present day would probably have been negligible (J.D. Peacock, pers. comm.). The reconstruction based on MSL1 is therefore rejected.

On the MSL3 model the curve projects to ~1.30m OD at the present day, which is 1.65m below the actual present-day MSL. This implies a sea-level rise in order to recover the current level. However, there is no direct evidence that sea-level has risen in this area during the past 2,900 years and this model also assumes that the samples were at or just below MHWST, which is again very unlikely. Therefore MSL3 does not seem a plausible reconstruction and is also rejected.

$^{4}$ More precisely, at Kincardine, MHWST lies at 2.95m AOD and MLWST at -2.35m OD (Admiralty Chart: River Forth Grangemouth to Stirling, No. 738, 1988 (corrected 1995)). Present day MSL at Kincardine is at 0.35m OD and so the spring tidal range is nearly symmetric (2.7m above MSL to 2.6m below).
Figure 8.3 Three models for Mean Sea Level for the Bothkennar area incorporating corrections for radiometric age, compaction and tidal range considerations.
Using model MSL2, which assumes that the samples were in situ at or just below the contemporaneous MSL (maintaining a smooth curve), the curve projects to around 0.35 m AOD at present day, the present-day value calculated for Kincardine. Thus model MSL2 seems a plausible reconstruction and it seems a reasonable assumption that the shell beds from which the index point samples were taken were likely to have formed somewhere near the contemporaneous MSL. This is also in reasonable agreement with a backward extrapolation of the sea-level curve from the present day MSL.

*This is the third correction to the Robinson curve and creates the best-estimate sea-level curve, which is henceforth taken as the standard sea-level curve for the Bothkennar area for the purposes of this Chapter.*

This (blue) curve is shown on Figure 8.4(a) along with the original Robinson (1993) curve and the corrected curves discussed above. Initially, it passes through the original Robinson sea-level curve (green) but diverges from it at about 3,600 BP. All of the index point samples fall between MSL and MLWST but at different distances from the MSL curve. This is considered plausible considering the preferred habitat of the sampled organisms. In theory the sea-level curve (which corresponds to contemporaneous MSL) could pass through any part of the shaded boxes although, as discussed earlier, it is unlikely that MSL was at the lower end of the possible range. Figure 8.4(a) also shows that the original (green) and corrected (magenta) Robinson curves both lie within the shaded boxes. However, since the projection of these curves does not agree with the present-day MSL they are also rejected.

Figure 8.4(b) shows the best estimate (MSL2) sea-level curve with the possible palaeotidal levels superimposed in order to establishes the contemporaneous MLWST and MHWST, in addition to the MSL. The tidal levels are based on the assumption that the palaeo-tidal range was similar to that of the present day (5.3 m at Kincardine). This follows from the suggestion (Austin, 1991) that there has been little change in the North Sea tidal regime during the Holocene and so the forcing at the mouth of the Firth of Forth would be similar to today. It is accepted that in practice changes in bathymetry (as the result, for example, of local sedimentation) may well have influenced the tidal regime near the head of the estuary and in particular that the progressive infilling of the estuary may have led to a consequent increase in the tidal range over time. Thus the palaeotidal range of Figure 8.4(b) must be considered a first approximation only.

### 8.2.3 Construction of a Sea-bed Curve for Bothkennar

The derivation of a water depth model requires the construction of a seabed curve so that the water depth (the difference between the sea-level and seabed curves) can be established. Such a
Figure 8.4(a)  Plot of the best-estimate sea-level curve for the Bothkennar area against the corrected curves.
Figure 8.4(b)  The best-estimate sea-level curve for the Bothkennar area.
curve can be based on dated and height-corrected index points analogous to those used for the sea-level curve itself.

8.2.3.1 Radiocarbon age determinations at Bothkennar

For the engineering geology programme at Bothkennar, shell samples from boreholes HW3, SH1\(^5\), D1 and D2 were collected for \(^{14}\)C dating by accelerator mass spectrometry (AMS) at the Oxford University Radiocarbon Accelerator Unit. These samples were selected (Peacock, pers. comm.) to cover critical stratigraphic locations in the Claret Formation (Figure 8.5) and are believed to have been in situ. The dates obtained have been reported in Paul et al. (1995) and are reproduced here in Table 8.2. Overall, the dates show that at Bothkennar only the lowermost sediments were deposited prior to about 5,000 BP and that the bulk of the Claret Formation was deposited between about 5,000 BP and 3,000 BP. Thus, although the earliest sediments were possibly laid down during the sea level rise of the main Holocene transgression, at Bothkennar most of the Claret Formation was apparently deposited after the peak of the transgression and thus during a period of falling sea level.

8.2.3.2 Seabed index points: elevation correction

The dates and present day elevations of the samples provide a set of seabed index points exactly analogous to those of a sea-level curve. As before, in order to establish a correct seabed curve, it is necessary to apply a compression correction similar to that used for the sea-level index points. This is somewhat more complex than that described earlier, since the correction must allow not only for autocompaction but also for the post-depositional surcharge by the lagoonal sediments of the eighteenth century reclamation. These corrections have been presented in detail previously by Paul and Barras (1998) and are reproduced here in Table 8.2. It may be noted that, for these points, the geotechnical data on which the correction was based were obtained by direct measurement of the sediment properties (taken from Nash et al., 1992b) instead of by analogy with generalised profiles and so the corrections are regarded as more reliable. It is also of note that the magnitude of the correction is greater than for Robinson's (1993) sea-level index points owing both to the greater thickness of underlying sediment and its more compressible nature.

The correction is, however, subject to some uncertainty since data are available only for the Claret Formation: the compressibility of the sediments from strata underlying the Bothkennar Gravel Formation (the Loanhead Beds of Browne et al., 1984) is not known. An assumed value similar to that of the lowest part of the Claret Formation has been used here: this will lead to a maximal estimate since the Loanhead Beds are generally less plastic than the latter and thus will

\(^5\) SH1 is a short (~0.4m) Sherbrooke sampled block from within the 'Soil Characterisation area' kindly made available by Dr. D.W. Hight.
Figure 8.5 Stratigraphical locations of dated samples from Bothkennar.
Table 8.2  Autocompaction corrections for ¹⁴C sample levels at Bothkennar.

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>Location</th>
<th>AMS Sample number and standardised³ age ¹⁴C yrs BP</th>
<th>Elevation of top of borehole (m OD)</th>
<th>Sample elevation (m OD)</th>
<th>Correction⁵ (m)</th>
<th>Corrected sample elevation (m OD)</th>
</tr>
</thead>
<tbody>
<tr>
<td>S1</td>
<td>Bothkennar HW3 NS 9204 8584</td>
<td>OxA-3389⁴ 5075±90</td>
<td>+3.1</td>
<td>-16.5</td>
<td>+0.5</td>
<td>-16.0</td>
</tr>
<tr>
<td>S2</td>
<td>Bothkennar D1 1032 4993¹</td>
<td>OxA-3388⁴ 3825±130</td>
<td>+3.1</td>
<td>-4.2</td>
<td>+2.1</td>
<td>-2.1</td>
</tr>
<tr>
<td>S3</td>
<td>Bothkennar SH1 1000 5060²</td>
<td>OxA-3507⁴ 3045±80</td>
<td>+3.2</td>
<td>+1.8</td>
<td>+1.2</td>
<td>+3.0</td>
</tr>
<tr>
<td>S4</td>
<td>Bothkennar D2 1227 5043¹</td>
<td>OxA-3387⁴ 2945±80</td>
<td>+2.6</td>
<td>-4.2</td>
<td>+2.0</td>
<td>-2.2</td>
</tr>
</tbody>
</table>

¹ Local co-ordinates, shown on Figure 3.8
² Local co-ordinates shown on Figure 3.8: sample from within 'characterisation study area', exact location not known.
³ Corrected for apparent age of seawater: 405±40 yrs BP (Harkness, 1983).
⁴ Source: Paul et al. (1995) Stratigraphical locations of samples are shown in Figure 8.5.
⁵ Source: Paul and Barras (1998).
be less compressible. Figure 8.6 shows separately the continuous correction curves obtained for the Claret Formation alone and those obtained when the underlying deposits are included. These latter data are reported in Table 8.2 and used in the correction of the seabed elevations. It can be seen from Table 8.2 that OxA 3389 (near the base of the sequence) and OxA 3507 (from the Cerastoderma bed) are now respectively about 0.5m and 1.2m below their depositional level whereas the mid-depth sample (OxA 3388) is about 2.1m below its depositional level. Figure 8.6 suggests that, as would be expected, the effect of the Loanhead Beds dominates the settlement of OxA 3389, is of minor importance to OxA 3388 and is negligible for OxA 3507. It can be noted in passing that the correction for OxA 3507 is consistent with the suggestion that the surface of the site now lies about one metre below the adjacent landward carse surface as a result of loading by the lagoonal deposits formed during the 18th Century reclamation works (Barras & Paul, 2000).

8.2.3.3 The seabed curve

Figure 8.7 shows the above radiocarbon dates from the Claret Formation at Bothkennar plotted against elevation after correction for the effect of postdepositional compression. The smooth (dashed) line through the seabed index points is a seabed curve which describes the changing elevation of the seabed against time as sedimentation proceeds and so allows estimates to be made of the sedimentation rate during the deposition of the Claret Formation. These points define a seabed path that shows the change in seabed elevation as the sediment accumulated. Since the statistical uncertainty in the dates (~80 yrs to ~130 yrs) is a sufficiently long interval to allow a significant fall in relative sea level, separate seabed curves have been constructed using the maximum and minimum age estimates and are also shown on Figure 8.7.

In reality, of course, the seabed elevation would not have increased along a smooth curve and aggradation would have been episodic. One such hypothetical path, based on the central estimate of Figure 8.7, is shown in Figure 8.8. It is not the only possibility, since the statistical uncertainty inherent in the radiocarbon ages allows a large family of such paths to be constructed: any path that lies within the boxes is feasible, at least in strictly geometric terms. The dotted boxes in the figure show the constraints on other possible seabed paths and, for comparison, the figure shows the lithofacies log from borehole HW3. It is very likely that a more realistic sedimentation path might well have followed a staircase pattern, as indicated by the alternation of the facies and evidence of minor erosional intervals. Such a path is also likely to generate fabric features of geotechnical importance, as discussed later in section 8.4.2.

8.2.4 Construction of a Water Depth Curve for Bothkennar

The elevation difference between the seabed path and sea level curve is an approximate measure of the water depth during deposition. Figure 8.9 shows the water depth curve
Figure 8.8 Autocompaction correction curve for boreholes HW3/ D1/ SH1. Lithofacies profile based on borehole HW3, (redrawn from Paul & Baras, 1998 with modifications).
Figure 8.7 Construction of a water depth curve (relative to MLWST) using the Bothkennar seabed curve and the model for the sea-level curve, as discussed in the text.
Figure 8.8 Reconstruction of a hypothetical sedimentation path based on a mean rate from radiocarbon dated samples and geological evidence of facies transitions.
calculated by simple difference between the sea-level and seabed curves. It is possible to construct two curves based respectively on the statistical range of the dates. The earlier $^{14}$C age gives a slightly greater water depth than the later age, owing to the continued isostatic uplift during the interval between them. It is then possible to deduce the relationship of the contemporaneous water depth to the present day profile in borehole HW3 (Figure 8.10) by first obtaining the water depths at the level of the $^{14}$C samples and then interpolating between them to obtain the continuous curves shown in Figure 8.10. From this figure it is possible to relate water depth to the facies succession.

The results suggest that at Bothkennar the water depth fell throughout the period of deposition of the Claret Formation. The rate of depth reduction appears to have been relatively constant for much of the period and then to have declined sharply towards its close, and also to have been largely controlled by the rate of sediment accumulation (central estimate: from $\sim 11.3$ mm yr$^{-1}$ between 5,000BP to 4,000BP, reducing to $\sim 7.5$ mm yr$^{-1}$ between 4,000BP to 3,000 BP) rather than by the rate of fall in sea-level due to continuing isostatic uplift ($\sim 4$ mm yr$^{-1}$ reducing to $\sim 1.5$ mm yr$^{-1}$ over the same periods: data from the corrected Robinson (1993) sea-level curve).

The intersection of the water depth curves with MLWST in Figure 8.9 and Figure 8.10 indicates the timing and stratigraphical depth of the onset of intertidal conditions. These appear to have been established around 3,300BP to 3,400 BP at a level now between 0m OD to $-$1m OD. The water depth reduced steadily from an initial value of around 20m near the base of the formation until at the level of the Bothkennar *Cerastoderma* bed conditions had become intertidal, as indicated by sample OxA-3507 which lies 2m above MLWST at this time and is thus just below mean sea level. This reconstruction accords with the present day ecology of *Cerastoderma edule* (mainly lower intertidal), and also with the observations of Sissons and Smith (1965a) that, at their landward edge, the modern tidal flats accumulate to within about 3 to 4 feet (0.9m – 1.2m) of the spring tide high water level.

It may be noted that this model, based on falling relative sea-level, differs significantly from the earlier proposal of Hawkins *et al.* (1989) that the whole of the deposit was formed during a period of almost stationary relative sea level. That proposal was made without the benefit of radiocarbon dates and presupposed that the sediments dated from around 7,000 BP. Further support for the alternative reconstruction proposed here is given by the $^{14}$C dates obtained by Robinson (1993) from other intertidal to immediately subtidal shell beds in the Grangemouth area, considered to be the stratigraphic equivalents of the *Cerastoderma* bed. The dates agree with that of OxA-3507 and their elevations (also shown on Figure 8.7) plot within the reconstructed intertidal range at the time of the *Cerastoderma* bed. This supports the contention that the suggested date of 3,300 – 3,400 BP may apply to the establishment of intertidal
Figure 8.9  Progressive reduction in water depth at Bothkennar.
The higher and lower curves are respectively constructed
from the older and younger ends of the age ranges of the
samples on which the seabed curves of Figure 8.7 are based.
Figure 6.10 Progressive reduction in water depth at Bothkennar compared with the sedimentary profile. See text for discussion.
conditions in the Grangemouth area more generally and also suggests that the palaeotidal levels
assumed for Bothkennar are broadly correct.

This reduction in water depth over time is also commensurate with the palaeogeographic
change at the site from an offshore (sublittoral) to coastal (tidal flat) position as sea level fell. At
the height of the Flandrian (Holocene) transgression (around 6,500 BP) the site lay about 3km
offshore from the contemporary coastline marked by the so-called main postglacial shoreline
(PG1 of Sissons; cf. Gostelow & Browne, 1986). About 2km west of the site there is a subdued
fall in the carse surface from an elevation of around 10m to 12m OD to around 7m to 8m OD or
lower. This change marks the position of the Holocene shoreline (probably equivalent to PG3 of
Sissons et al., 1966) that is tentatively dated at ~4,000 BP (Sissons, 1967, p.184; Smith et al., 1993)
and so is similar in age to the middle division of the Claret Formation at Bothkennar
(cf. OxA-3388, dated at 3825±130 yrs BP). This suggests that the site then lay around 2.5km from
the contemporaneous shoreline. During the time of deposition of the Claret Formation the
position of the local shoreline migrated eastwards, until its present position was reached at
about 3,000 yrs BP, as indicated by the date on the Cerastoderma bed at Bothkennar of
3045±80 yrs BP (OxA-3507) and the various dates obtained by Robinson (1993) from shells in
similar settings elsewhere in the eastern Forth area.

8.3 SUMMARY OF THE DEPOSITIONAL MODEL FOR THE CLARET FORMATION AT
BOTHKENNAR

The above evidence, together with that presented in Chapters Five and Six, allows the
construction of a depositional model for the Claret Formation at Bothkennar. The key features
of the model at Bothkennar are:

- The sediments were derived from glaciogenic sediments themselves derived from bedrock
  sources that contained a generally inactive clay mineral suite;

- Deposition occurred under subtidal conditions in a broad estuarine embayment, probably
  with a moderate tidal range, and wave action was probably limited throughout most of the
  period, other than during episodic storms, although it was probably more significant during
  the later phases of deposition;

- The sediments at 0.3m above the base of borehole HW3 (~16.5m OD) were deposited in
  ~20m water depth (below MLWST). The depth of water (relative to contemporaneous
  MLWST) reduced throughout the depositional period until conditions became intertidal at
  around -1m OD to 0m OD. The water depth during the deposition of the lowermost 0.3m
  of sediment is unknown, although sedimentological evidence indicates that current activity
  may have been significant;
• The sediments from about -1m OD to 0m OD up to about 1m AOD are inferred to have been deposited under intertidal conditions. It is also inferred that the severity of intertidal exposure increased upwards until the aggrading surface was colonised by Cerastoderma edule;

• The mean rate of deposition varied from around 11mm yr$^{-1}$ in the lower part of the sequence to around 7mm yr$^{-1}$ in the upper part;

• The sediment is variably bioturbated, as indicated by the density and style of mottling: the bioturbation is limited in the lowest parts of all the cores studied at Bothkennar and generally increases upwards until the intertidal sediments are encountered. At this level in the sediments there is an increased frequency of larger biogenic structures. Mottling appears to be associated with the finer-grained sediments: there is some evidence that it is suppressed by an increased modal size in the silt range;

• In the mid-sequence the salinity of the pore water is $\sim 30\%$. The salinity reduces from this value both upwards towards the ground surface and downwards towards the Bothkennar Gravel Formation. Above a level of around $-3$ m OD (6m bgl) the salinity decreases rapidly and above about 1m AOD the pore water is effectively non-saline;

• The sediments contain an organic component of around 2%-5% of dry weight of sediment. This is largely amorphous and appears to be the product of the estuarine plankton/benthos. It is not terrestrially derived and is apparently not composed of plant remains. Its exact nature is not known, although characterisation tests indicate both polysaccharide, nitrogenous and methanol-toluene extractable components.

8.4 GEOTECHNICAL IMPLICATIONS OF THE DEPOSITIONAL MODEL

The depositional model described above contains a number of elements that are likely to have influenced the geotechnical character of the Bothkennar clay. They can be divided into two broad groups:

• Those that influence the packing or bonding of individual soil particles and so contribute to the level of geotechnical structure;

• Those that create discontinuities in the sediment profile and so are likely to create similar discontinuities in the profiles of geotechnical parameters.

These two aspects of geotechnical character will be discussed in turn.
8.4.1 Geological Factors in the Level of Geotechnical Structure

The Bothkennar clay is, by definition, a geotechnically structured material, since at a given effective stress it possesses a greater resistance to deformation in the intact state than in the reconstituted state. Although the term 'geotechnical structure' does not presuppose a physical structure, it is nevertheless reasonable to suppose that there could be some sedimentological basis for this behaviour and, if so, that this in turn reflects the geological conditions of deposition.

The evidence of geotechnical structure presented in Chapters Three and Seven falls into two general categories:

- Evidence of the in situ level of structure, exemplified by, for example, the position of the void index in relation to intrinsic compression line (ICL);

- Evidence of behaviour at various phases of deformation, exemplified by the pre-yield stiffness the magnitude of the yield stress and the post-yield compressibility and brittleness.

Profiles of void index and liquidity index (Figures 7.24 and 7.25) show that, in its in situ state, the sediment has a variable level of structure. The geotechnical data from HW3 are summarised in Figure 8.11, together with the raw water content profile and the results of the water depth model. The geotechnical data show that below about 2.5m depth the sediment is structured relative to the ICL and there is a suggestion that the peak level of structure occurs around 8m depth (-5m OD) (cf. Hight et al., 1992a: figure 13; Figure 3.20 this Thesis). Above about 4m depth (-1m OD), the water content reduces steadily upwards until eventually the void index falls below the ICL itself. This reduction in water content above 4m depth possibly corresponds to the onset of shallow water to intertidal conditions, as evidenced by both the sedimentology of the deposits and the inferred water depth shown in Figure 8.11(a). If the peak level is indeed found at ~8m depth, it would appear that some reduction in structure is initiated prior to the actual intertidal exposure and it is speculated that this may be a consequence of wave shoaling and the dynamic loading that this would have imposed on the seabed.

The Bothkennar characterisation study (Chapter Three, section 3.5) revealed a significant difference between the pre-failure behaviours of the mottled and the bedded facies. Prior to yield, the mottled facies was found to be stiffer than the bedded facies. At yield, in the mottled facies the yield stress was higher and the strain at yield was lower than in the bedded facies. Following yield, the mottled facies was found to be more compressible than was the bedded facies. These two aspects are clearly inter-related, since the higher stiffness and yield stress in the mottled facies implies that the particles are not as free to move. This is also implied by the
Figure 8.11 Comparison of the sedimentary profile, inferred water depth and geotechnical subdivisions in borehole HW3.
higher void index, which indicates that, at given level of effective stress, a higher level of structure can be maintained in the mottled facies as compared with the bedded facies.

These observations can be explained by appealing to the sedimentary microfabric and to the presence of organic cements. In the case of the microfabric the important distinction lies in the role of the aggregated fabric and the cumulate fabric. It was argued in Chapter Five that the former is the ‘normal’ fabric produced by water column processes and that the latter reflects periods of reduced sediment input and reworking. If this is so it would be expected that the former would be more frequent in the bedded facies, which formed under conditions of more rapid sedimentation, and that the latter would be more frequent in the mottled facies. Although the numerical data (also reviewed in Chapter Five) are not sufficient to confirm this hypothesis, it appears intuitively reasonable and will be used as a framework for the discussion.

The inter-particle bonding is very probably organic in origin. It is also likely that only certain components of the organic fraction, notably mucopolysaccharides, are implicated. The action of such a cement, which would have a degree of mechanical flexibility and the capacity to reform to some extent during and after failure, can explain several aspects of geotechnical behaviour. These include the increased water content/void index (due to an open fabric stabilised by the cement), a well-defined yield point and increased intact undrained shear strength (both due to the direct effect of interparticle cementation). As discussed in Chapter Six, the cement is thought to be associated with both epipelagic diatoms and with the burrowing fauna in the mottled facies. Many such sediment dwellers produce mucopolysaccharides as part of their activities and are known to play a significant role in the cementation of modern estuarine muds. Evidence of similar cements can be seen directly in electron micrographs (in particular the fused sheet microfabric: cf. Chapter Five, Figure 5.20) and further examples have been illustrated by Paul et al. (1992b: vol.2).

The precise geotechnical roles of the aggregated and cumulate microfabrics is speculative. Individual aggregates are largely composed of clay particles in edge-edge or edge-face arrangements, sometimes with ‘welded’ joints. There is some evidence from other work (e.g. Tovey & Hounslow, 1994) that when a sediment deforms at the macroscopic level such aggregates move relative to one another without major internal distortion: the aggregates remain intact but move to take up the interaggregate voids which are of similar size. If so, such deformation will not be accompanied by a sharp yield point but will more resemble the failure of a granular material of moderate relative density. This is indeed what is seen in the bedded facies. The role of the cumulate microfabric is less clear: it may contain major voids which increase compressibility and these may be associated with increased bioturbation, which in turn may be associated with some types of cementation. Both these features are compatible with the
behaviour of the mottled facies during deformation and also would contribute to a greater fabric openness, expressed as an increase in void index.

8.4.2 The Development of Discontinuities within Geotechnical Profiles

Many profiles of geotechnical parameters, such as water content, density, void index, undrained shear strength and yield stress ratio contain discontinuities. Detailed study reveals that such discontinuities are normally associated with sedimentological evidence of events associated with changes (usually reductions or pauses) in the local accumulation rate. This evidence comprises:

- Intra-facies structures such as density variations within beds and changes in microstructure within beds;

- Intra-facies breaks and bedding surfaces, which separate individual beds and may be the result of hiatuses in deposition;

- Inter-facies transitions between macrofacies with dislocation of various parameters across the facies boundary;

- Relatively large erosional episodes evidenced by the decapitation of ripples and mottle cycles, the presence of laminae, often associated with underlying erosion surfaces or possibly winnowing of exposed surfaces. This erosional evidence becomes increasingly common at both top and base of sequence.

Since these sedimentological features arise from changes in the accumulation rate, they will be reflected in the seabed curve (or sedimentation path) introduced in section 8.2.3. Based on the lithological features observed in borehole HW3, Figure 8.8 suggests one possible such path that contains examples of all the above events, which are indicated by the letters [a] to [d]. The processes underlying each of these events and their possible relevance to the subsequent geotechnical properties of the sediment can be considered in turn:

- Periods of rapid sedimentation during which bioturbation was suppressed, indicated lithologically by the presence of bedded facies and often by an increased silt content.

This is illustrated by point [a] in Figure 8.8. Although the rate of deposition was rapid by geological standards, the sediment was probably not underconsolidated to a significant extent. Indeed, the use of Gibson's (1958) solutions for the average degree of consolidation in an accumulating body of sediment suggests that the average degree of consolidation would have exceeded 95% (Paul, pers. comm.). Thus, during pauses in deposition, the sediment would have remained largely stable and so would probably not have experienced substantial slumping and
resuspension. The stability of the sediment surface would have allowed the operation of other syn-depositional processes such as those described below.

- Periods of reduced sedimentation during which bioturbation occurred, now shown by the development of the mottled facies.

An example is illustrated by point [b] in Figure 8.8. Periods of reduced sedimentation allowed bioturbation to become significant and, on the microscopic scale, allowed the development of the cumulate microfabric, which is a small-scale condensed sequence. The transition between the bedded and mottled facies leads to geotechnical discontinuities: for example, in general (at a given depth) the mottled facies has a higher water content (by ~7%) than the bedded facies and the contact between them is often marked by a sharp change of water content (Figure 8.11(b)). Similarly, profiles of undrained shear strength (Figure 7.27(a)) can show a slight but perceptible strengthening at a bedded to mottled boundary. Similar features can be seen in profiles reported by other authors (Hawkins et al., 1991; Hight et al., 1992a; Nash et al., 1992a) and are thought to be responsible for discontinuities in continuous cone penetration profiles from the site (J.J.M.Powell, pers. comm.). Such periods of reduced sedimentation might obviously become the precursors to complete hiatuses (below) with which they thus share some common characteristics.

- Sedimentation hiatuses (or minor erosion) during which bedding surfaces became stabilised or were partially re-eroded.

An example is illustrated by point [c] in Figure 8.8. Macroscopic silt laminae and the associated granular microfabric provide evidence of the removal by re-erosion of the clay and silt fraction from an otherwise aggrading surface, followed by the re-deposition of medium to fine silt in the form of a thin lamina that may be transitional into the underlying bed or may rest on an eroded surface. Re-erosion of the clay can produce sedimentary structures such as rip-ups and mud-lumps: examples can be seen in the bedded (burrowed) subfacies near the top of the succession in HW3. In addition to bioturbation, physico-chemical processes were probably also operative during these sedimentation pauses and could, in principle, lead to minor geotechnical discontinuities. For example, small variations in water content, void index and undrained strength within otherwise uniform parts of cores may be the result of hardening processes which might include, in addition to simple mechanical rearrangement, fusion at clay particle contacts by the formation of authigenic clay minerals (cf. Burley et al., 1983) and point contact cementation by iron compounds (cf. Curtis, 1977).
Episodes of erosion and resedimentation.

An example is illustrated by point [d] in Figure 8.8. Such events are infrequent but clearly have major geotechnical significance should they occur. They are evidenced by the sporadic intrusion of laminated sediments into an otherwise bedded/mottled sequence, such as is seen between depths of 7.70m and 10.15m bgl in borehole HW3. In the south-east corner of the BSCRS such an episode, on a much larger scale, has completely removed the sediments of the Claret Formation and replaced them with the laminated, channelised, sediments of the Grangemouth Formation. It can be speculated that these individual features are in fact parts of a connected network of former supratidal and subtidal channels, similar to those seen today on most estuarine flats (Allen & Fulford, 1996; Dyer, 1998). If so, this has implications both for the consolidation time of the sediments at large-scale and also for the subsurface transmission of fluids (and contained substances) within a clay unit of supposedly low permeability.

8.4.3 Subdivision of the Sedimentological and Geotechnical Profiles at Bothkennar

The combination of depositional processes and post-depositional events described above generated profiles of various geotechnical parameters seen at the Bothkennar site. Since the depositional conditions have also generated the sedimentological profile, it is to be expected that there will be similarities in the vertical structures of the profiles. Some examples of these similarities have already been provided in Figure 8.11, which compares the detailed facies architecture in borehole HW3 with various measures of the geotechnical structure, as discussed in section 8.4.2 above. In order to discuss the relationship of the geotechnical profiles to the depositional conditions – specifically to the water depth model – it is helpful to recognise four informal lithological subdivisions of the Claret Formation. These have been described by Barras and Paul (1999) who named them the basal, lower, middle and upper divisions. It is not known for certain whether these subdivisions can be applied beyond the Bothkennar site itself, although the sedimentological evidence on which they are based can be recognised in other boreholes at Bothkennar (those reported in Hawkins et al., 1989; Hawkins et al., 1991; Paul et al., 1992a) and hence it is possible that they could be present in the Claret Formation over a wide area.

The two principal sedimentary facies (bedded vs mottled) within the Claret Formation are central to this classification. Their relative frequencies vary with depth and since this variation appears to be laterally consistent across the site it provides a basis for the subdivision of the formation. In the basal division the sediments of the Claret Formation are entirely bedded and commonly contain laminae of silty sand. They are not mottled and in this respect they differ from units of the bedded facies in the overlying sequence. The lower part of this succession, in which the bedded facies dominates the profile, is termed the lower division. Although some
sediments in this division may show mottling, it is usually insufficient to obliterate the primary bedding, which remains visible to visual inspection. The higher part of the sequence comprises the middle division in which the mottled facies is dominant. The mottling can be intense in this division, covering more than 70% of any vertical section, and shows the full variety of styles and sizes (mottle types I, II and III) described in section 5.2.3.

Above the middle division there is a zone where the bedded facies (burrowed subfacies) dominates the sequence. This zone is termed the upper division. As described in section 5.2.2 it is associated with a coarsening of the grain size and the presence of silt laminae and other erosional contacts between the individual beds. Mottling is less frequent and there are numerous macroscopic burrows, often truncated by bedding surfaces which also show primary sedimentary structures such as mud clasts (up to ~2mm - 3mm), flakes and rip-ups. The top of this division is marked by the *Cerastoderma* shell bed.

In each of these subdivisions the local water content profile has a distinctive character, which is reflected in other geotechnical parameters such as undrained strength and yield stress. These features are indicated in outline in Table 8.3, which indicates the overall geotechnical character of the stratigraphical units within the Claret Formation.

The lower division is generally silty and mottling is not common. This may be associated with a higher rate of sedimentation or increased reworking of the fine fraction due to current activity. The corresponding relatively low offset of water content values reflects the lack of bioturbation. For example, in HW3 there is a notable discontinuity in the water content profile at around -14m OD which corresponds in the core to the onset of a minor mottling episode. Similar breaks, less conspicuous, occur elsewhere and appear to be coincident with the tops of bedding units. In general, both the water content profile in this zone shows a zigzag, sometimes saw-tooth, pattern at a scale of tens of centimetres: A similar pattern, at a finer scale, can also be seen in high-resolution bulk density scans of the core (Chapter Five: section 5.3.2).

In the middle division, as the water depth reduced, there is an increased proportion of mottling, indicating bioturbation during periods of reduced sedimentation or current activity. At low levels in the division bioturbated units are less frequent than bedded units but become more frequent at higher levels. It is probable that sedimentation was intermittent throughout the deposition of these units and thus that several hiatuses are contained within them. These hiatuses have lead to discontinuities of varying significance: in particular, there is a major, widespread discontinuity (seen in other cores also: Hight *et al*., 1992a) at around -12m OD. This feature may be explained as the result of an intra-unit hiatus: the suggestion of Hight *et al*. (1992a) that it is the result of an earlier period of subaerial exposure seems unlikely in view of
<table>
<thead>
<tr>
<th>STRATIGRAPHICAL UNIT</th>
<th>LITHOLOGICAL CHARACTER AND CRITERIA FOR FIELD RECOGNITION</th>
<th>PROCESS AND ENVIRONMENT OF DEPOSITION</th>
<th>GENERALISED GEOTECHNICAL CHARACTERISTICS</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>CLARET FORMATION UPPER DIVISION</strong></td>
<td>Silty clay to clayey silt with distinct bedding surfaces, commonly marked by minor erosional features such as mudballs (up to ~2mm - 3mm), flakes and ripples. Surfaces sometimes separated by silt laminae, usually spaced several cm apart. Some mottling, macroscopic burrows common, terminated by well defined bedding planes and/or silt laminae. Base of division placed at the level above which macroscopic burrowing becomes evident. Top of division marked by a shell bed (principally containing <em>Cerastoderma edule</em>).</td>
<td>The upper division was probably formed under intertidal to immediately subtidal conditions. It is possible to detect an increased frequency of silt laminations and a reduction in mottling, especially in the northern area of the Site. These latter features suggest an energetic environment perhaps associated with subaerial exposure. There is a transitional passage between the upper and middle divisions in many profiles: the unit is better-developed in the north of the Site.</td>
<td>The division lies within the lightly weathered transition zone immediately below the crust. Increased wave disturbance, suction stresses, desiccation and redox changes have probably contributed to its geotechnical signature. Median grain size ~6μm - 10μm. Water content ~45% - 70%. Water content and void index profiles show an overall decrease relative to the middle division. There are corresponding increases in the undrained shear strength and yield stress ratio.</td>
</tr>
<tr>
<td><strong>CLARET FORMATION MIDDLE DIVISION</strong></td>
<td>Silty clay to clayey silt, bedding surfaces often partially or completely obliterated by bioturbation. Heavy mottling in a variety of patterns and densities. Mottling often cyclic from coarse low density upwards to fine, high density over intervals of 300mm or more. Base of division placed just above lowest level where medium density mottling (~40% - 50% cover) extends over more than 100mm interval.</td>
<td>The middle division was formed under subtidal conditions during which sedimentation rate reduced with time, allowing intense bioturbation. The division is thus dominated by the mottled facies in its upper part, whereas transitions between bedded and mottled facies become more frequent in its lower part suggesting more rapid sedimentation with less reworking. The division is locally channelled: the channels are occupied by finely laminated clayey silts.</td>
<td>Water content 55% - 75%, greater variability in upper part associated with more intense mottling. Transitions from mottled to bedded facies associated with breaks in water content and related profiles. A well-marked discontinuity can be recognised in almost all profiles of water content, yield stress and undrained strength at an approximate depth of ~6m to ~12m OD. In the lower part the median grain size is ~2μm - 4μm, but in the upper part increases steadily to 5μm - 8μm.</td>
</tr>
<tr>
<td><strong>CLARET FORMATION LOWER DIVISION</strong></td>
<td>Silty clay to clayey silt with distinct bedding surfaces. Surfaces often separated by silt laminae especially in lower part. Mottling infrequent in upper part (coarse: cover less than 30% over intervals less than 100mm) to absent in lower part. Base of division taken as level above which shell debris becomes very rare.</td>
<td>The lower division was formed under subtidal conditions during which sedimentation rates were rapid. It is dominated by the bedded facies: occasional mottling is sporadic, nearly always coarse and usually insufficient to obliterate the primary bedding. Thus we regard the Lower division as a coherent, bedded unit.</td>
<td>Water content 45% - 60% shows a steady reduction with depth: the gradient is less in the basal division and individual values have limited scatter about the mean line. Most profiles within this unit contain discontinuities in water content and related parameters: these are interpreted as local boundaries of individual beds. Median grain size ~4μm - 6μm at base: at higher levels median grain size reduces, minimum of ~3μm around the top of the division. This trend is possibly the result of a decrease in sedimentation rate and is associated with increased bioturbation in the uppermost part.</td>
</tr>
<tr>
<td><strong>CLARET FORMATION BASAL DIVISION</strong></td>
<td>Silty clay to clayey silt plus subsidiary sand. Entirely bedded facies with no mottling. Frequent laminae of medium to coarse silt and fine sand. Distinct bedding surfaces commonly separated by laminae of fine sand/coarse silt, often irregular or contorted. Shell fragments and disarticulated whole shells found throughout division. Base of division has a sharp contact with the underlying Bothkennar Gravel Formation.</td>
<td>Possibly formed during the Flandrian transgression during a period of rising sea-level. The faunal evidence suggests a condensed succession laid down in predominantly shallow, offshore marine conditions. The grading and sedimentary structures suggest considerable current activity, perhaps associated with tidal movements.</td>
<td>Water content ~35% - 40% and commonly shows a sharp decrease from the overlying lower division. Profiles typically show strong discontinuities near the base. Laminae impart an irregular signature to both the piezocene and undrained strength profiles. Median grain size ~4μm - 8μm.</td>
</tr>
</tbody>
</table>
the water depth model, which indicates subtidal conditions at this level in the profile. This seems to preclude an origin as a drying surface.

The upper division has a distinctive sedimentology which includes a reduction in mottling, local predominance of the bedded facies, sporadic silt laminae and, in particular, the presence of macroscopic (>1cm) burrows which usually overprint other biogenic structures. A distinct geotechnical pattern can be identified in this division. There is a fall in water content, equivalent to that first identified by Skempton (1948) in his geotechnical zone II (cf. Chapter Three, section 3.2.1), that is associated with a marked increase in the spread of the envelope of water content values. This suggests that the sediment contains a variable level of structure, which may have arisen from burrowing and reworking, combined with fluctuating conditions of effective stress and biochemical/chemical (cementation) environment. We note that other studies (Nash et al., 1992b; cf. Figure 7.26) have found that the yield stress ratio increases above this depth (from ~1.6 to >2.5) which is probably the result of an increased degree of both organic (possibly polysaccharide) cementation and intermittent drying.

These observations have a probable geological explanation. The upper division is believed to be intertidal in origin, both from the sedimentological evidence of increased silt lamination and macroscopic burrowing and from interpretation the water depth model. This implies that sediments located above about ~1m OD are likely to have been exposed between tides and, clearly, the higher above this level the sediment lies, the longer its exposure during each tidal cycle. This has a number of geotechnical implications, including: wave action and sediment sorting; periodic drying and thus the development of cyclic suction stresses; cementation due to redox changes and the precipitation of inorganic oxides and hydroxides; changes in the soil fauna and related changes in organic geochemistry. Together these factors are considered to be responsible for the distinctive geotechnical character of the upper division.

These subdivisions can also be identified in the cores from the previous University of Bristol investigations (Nash & Lloyd 1988a; 1989; Hawkins et al., 1989). The resulting picture of the site is shown schematically in Figure 8.12. The data show that the Claret Formation is everywhere separated from the overlying Grangemouth Formation by an erosion surface and is itself similarly separated from the underlying Bothkennar Gravel Formation. It was for this primary reason that the sediments of the then Claret beds of Browne et al. (1984) were raised to Formation status by Paul et al. (1995) as discussed in Chapter Two. Within the Claret Formation the individual divisions at the scale of the BSCRs have an apparently pseudo-tabular architecture. However, they are bounded by undulating surfaces and contain numerous minor surfaces of erosion or non-deposition. They are probably lenticular on the larger scale, since they represent a tidal mudflat stratigraphy, although their geometry has yet to be established in detail and so is shown only in idealised form in Figure 8.12. This model envisages the Claret
Figure 8.12: Geological block diagram of the Bothkennar area. Sources: Hawkins et al. (1986); Paul et al. (1992b).
and Grangemouth formations as individual, relatively large scale, cut and fill units which form part of the Holocene infill of the Forth valley. It is possible that this entire infill is constructed of many such units, each bounded by inter-formational erosion surfaces and having diachronous relationships with one another.

8.5 COMPARISONS WITH ANALOGOUS SOFT CLAY DEPOSITS

The aim of this section is to compare and contrast the geotechnical properties of the Bothkennar deposits with other Holocene marine soft clays which have analogous settings and post-depositional sea-level histories. These comparisons illustrate the effect of the environment of deposition and post-depositional processes on the geotechnical properties of fine-grained deposits as a result of the position of a sample, with respect to sea-level, since its deposition. The geotechnical character of the sediments can then be related to their lithology and to their environment of deposition, in particular to their history of local relative sea-level change, to the extent that such information is available.

The sites which have been chosen for comparison with Bothkennar are: Belfast in Northern Ireland; Brean on the Severn Estuary and Drammen in southern Norway (collectively referred to as the 'comparison sites' hereafter). Their locations in relation to the regional pattern of uplift or subsidence at the present day are shown in Figure 8.13. Table 8.4 presents an overview of their Holocene geological setting, together with the geochemical and geotechnical character of the profiles reported in the literature.

The choice of these sites was based on a desire, firstly, to examine a range of depositional settings and post-depositional histories of relative sea-level movement. These differences have lead to inter-site variations in mineralogical and geochemical properties which in turn determine the geotechnical behaviour of the deposits. For practical reasons, the particular comparison sites were chosen for the availability of data: the material on which the city of Belfast was built was investigated when it became clear that there were subsidence problems within the city due to foundation conditions; foundation problems in the town of Drammen were investigated in detail by the Norwegian Geotechnical Institute; Brean was one of the candidates to become the national soft clay research site and was investigated at the same time as Bothkennar.

Although the data are incomplete for some of the comparison sites, it has proven possible overall to assemble a coherent picture of the geological and geochemical controls that have governed the geotechnical character of the deposits at each site and so it is possible to place Bothkennar in an overall context.
Figure 8.13 The locations of the Bothkennar Soft Clay Research Site and the three comparison sites at Belfast, Brean and Drammen. The isobases show the rate of land uplift and subsidence (in mm) (based on Devoy, 1987: figure 10.6 [Europe and Scandinavia] and Shennan, 1986: figure 9 [United Kingdom]). Dashed and dotted isobases are less certain.
Table 8.4  Details of the four comparison soft clay sites.

<table>
<thead>
<tr>
<th>SITE</th>
<th>BOTHKENNAR</th>
<th>BELFAST(^1)</th>
<th>BREAN(^2)</th>
<th>DRAMMEN(^3) (plastic clay)</th>
</tr>
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<tbody>
<tr>
<td><strong>Geological comparators</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Depositional setting</td>
<td>Estuarine</td>
<td>Estuarine depth not stated</td>
<td>Estuarine intertidal flats and salt marsh</td>
<td>Temperate fjord</td>
</tr>
<tr>
<td>Modern spring tidal range</td>
<td>5.6m</td>
<td>3.1m</td>
<td>~12m</td>
<td>n/a</td>
</tr>
<tr>
<td>Age of deposits</td>
<td>5000 to 3000 BP</td>
<td>Basal peat ~9,000 BP</td>
<td>6,000 to 4,000 BP [OD peat]</td>
<td>~3,000 BP</td>
</tr>
<tr>
<td>Relative sea-level change since deposition</td>
<td>Uplift 3m - 8m</td>
<td>Little change ~0m with some oscillation Previous fall, now static</td>
<td>Subsidence &lt;~5m Continuing sea-level rise ~2mm a(^{-1})</td>
<td>Uplift ~25m Continuing sea-level fall ~3mm a(^{-1})</td>
</tr>
</tbody>
</table>

| **Geochemical Comparators**              |                                 |                     |                   |                               |
| Mineralogy of clay fraction             | Illite, chlorite, kaolinite, quartz and feldspar flour | Illite, chlorite, swelling chlorite, kaolinite | Illite, chlorite, illite-smectite, kaolinite, quartz | Illite, chlorite, quartz and feldspar flour. |
| Typical organic content                 | 3% - 5% (marine)                | 3% - 5% (possibly marine) | >5% (terrestrial) | 1%                            |
| Pore fluid salinity                     | 30% mid-depth reducing <5% in top 5m | c.14% reducing >5% in top 3m | Not stated probably close to estuarine | 25% initial <1% after leaching |

| **Geotechnical comparators**             |                                 |                     |                   |                               |
| Percent clay size                       | 40% - 50% below 2m bgl          | 40% - 50% 2m-8m bgl | 30% - 35%         | 20% - 30%                     |
| Liquid limit                            | 54% - 87%                       | 75% - 92%           | 40% - 60%         | 55%                           |
| Plastic limit                           | 26% - 43%                       | 27% - 35%           | 20% - 25%         | 25%                           |
| Plasticity index                        | 20% - 53%                       | 48% - 57%           | 15% - 30%         | 30%                           |
| Maximum liquidity index                 | ~1.2                            | ~1.4                | ~0.5 except in silts at base | ~1.0 (original) ~2+ (leached) |
| Typical activity                        | 0.9 - 1.3                       | 1.2 - 2.0           | 0.7 - 1.0         | 0.7 - 1.0                     |
| Typical yield stress ratio              | 1.5 - >2                        | 1.0 - >2.0          | 1.1 - >2.5 \(^4\) | 1.6                           |
| Typical sensitivity                     | 5 - 10                          | 5 - 10              | Not stated        | 8 - 10 (original) 200+ leached |

\(^1\) Crooks & Graham (1976); \(^2\) Hawkins et al. (1989); \(^3\) Bjerrum (1967); \(^4\) Narbett (1992).
8.5.1 Belfast

Following the retreat of the last major (Midlandian) ice sheet from Ulster (estimated to be around 18,000 years BP (Wilson, 1972)), Northern Ireland has been influenced by the general pattern of isostatic recovery across northern Britain. Uplift was centred on Scotland (probably in the area of Rannoch Moor) and Belfast lies on the margin of uplift close to the hinge-line. At the present day sea-level is falling at 2.4mm yr\(^{-1}\) at Malin Head, is rising at 0.3mm yr\(^{-1}\) around Dublin and is stable at Belfast (Devoy, 1987). The sea-level curve for the south-eastern Ulster coast at Down, south of Belfast, is shown in Figure 8.14(a) and indicates the transgression that marks the start of the Littletonian\(^7\) period (Wilson, 1972; Taylor et al., 1986; Lambeck, 1996) following the eustatic sea-level rise (of about 10m to 20m) from the melting of the Laurentian and Fennoscandian ice sheets.

The general location of the city of Belfast [GR 350 750] [54°36'N, 5°55'W] on Belfast Lough is shown on Figure 8.14(b). The Holocene estuarine deposits of the Belfast region of Northern Ireland occur within the Lagan Valley and Belfast Lough which connects to the North Channel of the Irish Sea. In the Lagan estuary the Littletonian transgression is marked by the covering of thin peats (dated at 9,100±200 years BP (Wilson, 1972)) by unconsolidated sands, muds and silts. The Holocene deposits in the Belfast area lie in the broad trough formed by the Lagan River valley and Belfast Lough. Figure 8.14(b) shows the extent of these deposits, which consist of narrow bands of marine raised beach deposits and present-day marine beach deposits along the shores of the Belfast Lough. These widen out into extensive estuarine alluvium measuring about 3km in width.

In the Belfast area itself these deposits consist of three distinctive horizons (Wilson, 1972; Crooks & Graham, 1976; Gregory & Bell, 1991). The lowest is a brownish-blue sandy clay with abundant mollusc shells, foraminifera tests and roots and leaves of grass wrack, and has been interpreted as an indication of intertidal conditions at the commencement of the Littletonian transgression. This was followed by warm, low-salinity open water conditions in water of about 5m depth. The upper horizon consists of a blue-grey clay with fewer in number but greater variety of shells than the lower horizons and was attributed to increasingly saline, cooler conditions in water depths of about 9m. These clay deposits are known locally as ‘sleech’ (Gregory & Bell, 1991) and are generally up to 15m in thickness under Belfast city and 20m nearer the Lough, although on the flanks of the Lough they reach a maximum of about 6m in thickness (Crooks & Graham, 1976).

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\(^6\) In Ireland ‘Midlandian’ or Midland General Glaciation (the local terminology for Ireland) correlates with the British Stage ‘Devensian’.

\(^7\) Littletonian or ‘Post-glacial period’ in Ireland correlates with the Flandrian (Holocene) British Stage (Wilson, 1972: table 4) now referred to as the Holocene.
Figure 8.14  (a) The relative sea-level curve for County Down, east-central Northern Ireland (after Lambeck, 1996: figure 13d).
(b) Drift geology of the Belfast region showing the location of Belfast Lough and the Kinnegar site (map source: Geological Survey of Northern Ireland, 1991).
The geotechnical character of the estuarine deposits has been described by Crooks (1973), Crooks and Graham (1976), Bell (1977), and Gregory and Bell (1991). Although the estuarine deposits have been reported to be up to 20m thick, none of the boreholes described in the literature reaches below about 8m. The log of one of these from Kinnegar (hereafter referred to as Kinnegar b/h No.1), on the southern flank of the Belfast Lough, is illustrated in Figure 8.15 (Crooks & Graham, 1976: figure 3). Here the main sedimentary sequence which lies between 2.0 and 8m depth consists of 6m of soft, grey, lightly fissured, organic silty clay with a moderately high sensitivity. The top 1m consists of grey sand, with grey silty sand between 1.0 and 2.0m. About 1.5m of basal sands were found below 8m depth (Crooks & Graham, 1976).

Figure 8.15 also shows profiles of the key geotechnical properties (Crooks & Graham, 1976: figure 2). In the main sequence, the water content between 2m and 3m increases from about 60% to 80%. Below this there is a general trend of decreasing values with depth but with breaks in the profile. The profile of exchangeable cations shows that sodium, potassium and magnesium all behave in a similar fashion and that there is a sharp decrease in all these cations above about 2m-3m depth. The calcium content appears not to vary with depth and may reflect the presence of disseminated shell debris.

The liquid limit varies with depth, with a corresponding change in the plasticity index. The maximum values ($W_L \sim 110\%$; $I_p \sim 70\%$) occur between 3m and 4m depth: there is a decrease in both parameters from about 4m to the base of the borehole and a sharp decrease above 3m depth. The clay fraction between 4.0m and 8.0m is between 30% and 40% and X-ray diffraction studies (Crooks, 1973) have shown that it consists of illite, chlorite, kaolinite and swelling chlorite, together with quartz and feldspar flour. The activity (Figure 8.15) generally lies in the range 1-2, with an outlying value of 3 at a depth of 2m, and it may be noted that these are high values for this clay mineral suite.

In most cases the water content was below the liquid limit, indicating a liquidity index of $<1$. This differs from values reported by Gregory and Bell (1991) for a borehole at the same location (hereafter referred to as Kinnegar b/h No. 2) (Figure 8.16) where the values of liquid limit between 2 and 3m were found to be as high as $\sim 90\%$ with the water contents up to 93%. In that borehole the water content was close to or above the liquid limit (liquidity index often $>1$). This was reported by Gregory and Bell (1991) to be the result of the organic matter which is about 5% below 2.0m with averages of $\sim 3\%$ in the main sedimentary sequence. Although the organic content was similar in the two boreholes, the clay content of $\sim 50\%$ in Kinnegar b/h No. 2 (sand:silt:clay ratio: 10:40:50) was about 10% higher than in Kinnegar b/h No. 1 (sand:silt:clay ratio: 2:60:40) and this is thought to be the most likely reason for the higher liquidity index in borehole No. 2.
Figure 8.15 Borehole log and geotechnical profiles from Kinnegar borehole No. 1. Belfast (source: Crooks & Graham, 1976).
Figure 8.16 Geotechnical profiles from Kinnegar borehole No. 2, Belfast (source: Gregory & Bell, 1991).

Figure 8.17 Correlation of plasticity index with exchangeable sodium, Kinnegar borehole No. 1, Belfast (source: Grooks & Graham, 1970).
At Kinnegar there is a positive correlation between the liquid limit and the concentration of adsorbed sodium (Figure 8.17) although, as discussed below, this relationship may not be straightforward. The organic content (loss on ignition method\(^8\)) was found to be about 4% in the soft clay deposits but only about 1% in the upper sandy layers. Measurements of undrained shear strength (Figure 8.16) show considerable scatter, probably as a result of the sampling procedures adopted at the time of the work. It can be seen, however, that below 4.0m the profile shows a generally increasing trend expected of a normally-consolidated clay material. The sensitivity at the Kinnegar site was reported to range from 5 to 15 with a mean value of 8 (Gregory & Bell, 1991), although the highest value shown in Figure 8.15 appears to be about 10. These values (≤15) were considered to be relatively high for a clay deposit in this area and this has been attributed to the leaching of the magnesium cation (Bell, 1977; Gregory & Bell, 1991).

### 8.5.2 Brean (Severn Estuary)

The Severn Estuary is located in the south-western region of England and forms the innermost part of the Bristol Channel which is open to the Atlantic Ocean (Figure 8.18(a)). The Severn Estuary is flood dominated and is second only to the Bay of Fundy in Nova Scotia with respect to its tidal range (Allen 1990, 1991): Bristol has a mean spring tidal range of 12.2m and Cardiff 11.2m (Hydrographer of the Navy, 1999).

In southern England the surface has undergone downwarping for at least the last 4,000 years (Shennan, 1989) resulting in a relative sea-level rise. The highest estimated rates of subsidence are for the Thames Estuary and Norfolk (up to 2mm yr\(^{-1}\)) (Figure 8.13), which contrasts with northern England and Scotland where uplift has continued throughout the Holocene. In the Bristol Channel area the sea-level data are poorly resolved (Figure 8.18(b)) and the sea-level curve is a composite obtained from a wide range of sites in the Severn estuary (Shennan, 1989).

In spite of these uncertainties it seems likely that there has been a net isostatic subsidence at a rate of about 0.24 ± 0.19mm yr\(^{-1}\) (Shennan, 1989) and a consequent sea-level rise over the last 8,500 years.

The extensive estuarine alluvial deposits which are found along the flanks of the Severn Estuary include both intertidal mudflats and salt marsh which cover a wide area due to the high tidal range. Allen (1992) described the alluvial deposits of the inner Bristol Channel and Severn Estuary in terms of a 'standard' geological sequence for the area (Figure 8.18(c)). It can be seen that the succession is dominated by the Wentlooge Formation (WF) which is informally divided into the lower, middle and upper divisions. The lower division consists of thin gravels, sands,

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\(^8\) Head (1992) states that the ignition method is "suitable for sandy soils which contain little or no clay" and it is, therefore, unsuitable for clay soils: it gives high values of organic content as it removes the water of crystallisation. The potassium dichromate method of determination is recommended provided that precautions are taken to eliminate the effect of high chlorine or sulphide contents.
Figure 8.18 Distribution (a), sea-level curve (b) and stratigraphy (c) of the Holocene deposits in the Severn estuary. (Source: Allen, 1992).
organic-rich palaeosols and rooted peats which grade up into silty sands. These in turn are overlain by bluish to greenish grey sandy to clayey silts. The middle division consists of peats (dating from 6,500 to 2,500 years BP) intercalated with organic-rich beds and sandy to clayey silts. The upper Wentlooge Formation is represented by bluish to greenish grey mottled silts which change to a green-brown colour at the top of the formation and are topped by salt marsh deposits at the surface. Deposition of these Holocene alluvial deposits has largely been curtailed due to reclamation which commenced as early as the Roman period (Allen & Fulford, 1986) and continues up to the present day (Allen, 1986).

The detailed architecture of part of the alluvium is known from Brean on the Somerset Levels [NGR ST 308962] (Hawkins et al., 1989). The borehole log from the site (down to 14.0m) is shown in Figure 8.19. The lowermost two metres comprise stiff clayey silt (clay content <20%), which is overlain by a unit of firm miaceous clayey silt (clay content 30% - 40%) with occasional sand, grey brown mottles and traces of rootlets. This unit contains peats at about 6m bgl. These deposits have been interpreted by Hawkins (1984) as being deposited in a low energy environment which favoured the deposition of estuarine muds, possibly an upper intertidal environment as indicated by the presence of phragmites roots and stems. Their deposition appears to correspond to a decrease in the rate of sea-level rise in this area (Hawkins, 1984). There are occasional horizons of thin (1mm) silt laminae which were deposited in a higher energy environment, possibly during storm conditions.

The peats occur in two layers (between ~5.5m and 6.5m bgl) separated by silty clay containing phragmites stems. They are believed to represent salt marsh vegetation which formed at approximately high water spring tide. They probably correlate with the Ordnance Datum (OD) peat horizon found widely in the lowlands bordering on the Severn Estuary, which has been dated between about 6,000 to 4,000 yrs BP and, in particular in the 'levels' north of the Mendip Hills, has been dated between 6100±120 yrs BP and 4145±100 yrs BP (Hawkins, 1971). The sequence is completed at surface by a desiccated crust which extends down to a depth of about 2.0m and consists of firm, brown, silty clay with frequent rootlets and becomes gleyed at the base.

The minerals present in the estuarine alluvial deposits in the Severn Estuary were reported by Narbett (1992) who used XRD analyses to examine samples from a number of sites. He found the mineral suite to be of fairly constant composition and generally only showed variations in the relative abundance of the different minerals. The main minerals present in order of decreasing abundance were quartz, clay minerals (illite [dominant], expanding clays, kaolinite and chlorite), calcite, dolomite, feldspar and pyrite. The expanding clays were present in the form of mixed-layer clays consisting of illite-smectite. The organic content of the sediments was
Figure 8.19  Geotechnical profiles and site location for Brean (Somerset, England).
(Source: Hawkins et al., 1989).
measured by means of the ignition method (at 425°C for 19 hours). Outside the peat horizons the organic content was generally between about 3% and 6%. The lower, thicker peat horizon contains 85% organics whilst the upper peat layer (between 5.48 and 5.68m) contained only 25%, reflecting an influx of sediment which did not occur when the lower peat layer was active.

The geotechnical profile at Brean is also shown in Figure 8.19. Between the base of the desiccated crust (~2m bgl) and the top of the upper peat the water content is mostly around 30%, which corresponds to a liquidity index of <0.5. In this interval the plasticity index fell from >25% to ~12%, owing mainly to a reduction in the liquid limit. Below the lower peat layer the water content increased from about 35% to about 50% at about 8.5m and then decreased to about 35% at the base of the borehole (14m bgl). In this lower interval there was a corresponding change in the liquidity index from around 0.5 in the clay unit to near unity in the basal clayey silt unit. The plasticity index reduces consistently in this interval from about 40% to <20%, again due to a reduction in the liquid limit.

Neither the remoulded shear strength nor the geochemical profiles of these deposits were reported by Hawkins et al. (1989). Sensitivity values, therefore, could not be directly calculated for this site. However, Cook and Roy (1984) estimated the sensitivity of deposits from the Somerset alluvium to be in the region of 8 or 9 at their liquid limit. This implies that, below a depth of about 10.0m, the sensitivity of the deposits at Brean tested by Hawkins et al. (1989) were probably close to these values as the in situ water content was close to the liquid limit. Above this depth the in situ water content found by Hawkins et al. (1989) was well below the liquid limit and the sensitivity would therefore probably have been lower than 8 or 9.

8.5.3 Drammen (Oslofjord, Norway)

The mountainous nature of the terrain of Norway results in the majority of the country's population living on a narrow strip of low-lying land along the coastal belt which makes up only about 30% of the land area of the country (Bjerrum, 1967). These low-lying sediments are generally soft and often are quick clays. Owing to the compressible and unstable foundation conditions that these clays provide, the Norwegian Geotechnical Institute was established 1953 in order to carry out research and to advise civil engineers on design and construction on these sediments. Early research was concentrated on the foundations of buildings in the town of Drammen, situated about 40km south-west of Oslo, where Dammenselvet (the Drammen river) flows into the Drammensfjord, a branch of the Oslofjord (Figure 8.20).

The Late-glacial and early Holocene chronology of the Oslofjord area has been established by a series of well-dated recessional moraines from the southern margin of the Fennoscandian ice sheet (Sørensen, 1979). Drammen is located between the lines of the Ski Moraine
Figure 8.20 Location map of Drammen (Norway) in relation to the Late-glacial and Holocene moraine system and the sea-level curve for the area.
(Source: Sorensen, 1979.)
(10,200 $^{14}$C yrs BP) and the Aker Moraine (9,800 $^{14}$C yrs BP) which indicates that, to a first approximation, the area was deglaciated around 10,000 yrs ago, as stated by Bjerrum (1967). As the ice withdrew, unloading of the crust took place and isostatic rebound occurred, which raised the marine deposits above sea level and allowed fresh water to replace saline pore fluids. In the Drammen area, the uplift since ice retreated from the valley has amounted to about 205m relative to present sea-level. This is confirmed by the sea-level curve for the Ski area (Figure 8.20: from Sørensen, 1979: figure 2) which shows that here there has been about 220m of uplift since 10,000 BP. Isostatic uplift continues to this day throughout the inner Oslofjord region at a rate of about 3mm yr$^{-1}$ at Drammen (Bjerrum, 1967).

The Late-glacial to Holocene sequence in the Drammendalen (Drammen valley) is illustrated schematically in Figure 8.21. Immediately above bedrock is found a unit broadly termed by Bjerrum the ‘lean clay’, a clayey silt with an admixture of sand dispersed throughout the matrix. Bjerrum (1967) ascribes this structure to the ready availability of coarser particles (derived from the meltwater input to Drammendalen) which were sedimented simultaneously under saline conditions with clay flocs of similar settling velocity. This led to an homogeneous, open microstructure with a relatively low plasticity. As relative sea-level fell and the climate ameliorated, deglacierisation allowed only silt and clay to reach the Drammen area and thus the later deposits, termed the ‘plastic clay’, have a higher plasticity and a higher organic content than the underlying lean clay. These deposits are themselves overlain by fluvial deposits dated to about 3,000 BP (Bjerrum, 1967).

The clays at Drammen are, in geological usage, normally consolidated, in that no load has been removed from them since deposition. Profiles of geotechnical properties found in two boreholes A and B are given in Figure 8.21. In the lean clay the water content profiles and Atterberg limits do not differ greatly. In the plastic clay, however, there is a marked difference between the two boreholes. The water content in borehole A was at, or just above, the liquid limit, whereas in borehole B at a corresponding depth the water content was about 10% higher than the liquid limit. This is the result of a reduction in the liquid limit and not an increase in water content, in fact it can be seen that the water content in borehole B is lower than that in borehole A in that part of the core. The undrained shear strength also showed marked differences between the two boreholes: in borehole B, in the region of the plastic clay, values were obtained in the region of about 10kPa, which is some 20kPa to 30 kPa less than at the same depth in borehole A. The corresponding sensitivities were 200 to 300 (borehole B) and around 8 (borehole A). Thus the plastic clays in borehole B are well into the ‘quick’ category whereas those in A are not quick.

In most respects they exhibit several properties often considered characteristic of normal consolidation; in particular, below the desiccated crust their water content shows a steady (though punctuated) reduction and the ratio of undrained strength to effective stress is constant.
Figure 8.21 Late-Glacial and Holocene stratigraphy and geotechnical profiles at Drammen (source: Bjerrum, 1967).
and obeys the general relationship with plasticity index proposed by Skempton (1948). They do, however, show two anomalous features. In boring B, the water content substantially exceeds the liquid limit over parts of the profile and in the plastic clay, the one-dimensional yield stress exceeds the in situ vertical effective stress by a constant factor of about 1.6.

Isostatic uplift has changed the hydrogeology of the area (Bjerrum, 1967; Moum et al., 1971), since differential uplift between the floor and sides of the valley has created an upward pore pressure gradient and has resulted in artesian conditions. The clay deposits, therefore, have been subjected to an upward flow of fresh water with consequent leaching of the saline pore water. This change in pore water chemistry has had profound effects on the geotechnical character of part of the sedimentary profile and is considered the basic mechanism by which the clays has been brought to a quick condition.

Figure 8.22 (from Moum et al., 1971) shows the geochemical profiles for two boreholes that penetrated the quick clay with a clay-size fraction consisting of illite and chlorite, plus substantial quartz and feldspar flour. The geotechnical profiles show that there are distinct quick clay and soft, sensitive (but not quick) clay sections. Their development appears to have been controlled by the pore water chemistry. The quick clay units have a low Ca$^{2+}$ concentration (~10mg/l) but a relatively high Na$^+$ concentration (65 to 75 mg/l). By contrast, the ‘soft clay’ has a much reduced concentration of Na$^+$ (10 to 25 mg/l) and an increased Ca$^{2+}$ concentration (~5mg/l - 25 mg/l). The high Na$^+$ to Ca$^{2+}$ ratio in quick clay unit agrees with the findings of Quigley (1980) on the Canadian sensitive clays, who showed that in these clays a high ratio of adsorbed monovalent cation relative to the divalent cations is one of the factors which produced a low remoulded shear strength and, therefore, a high sensitivity.

There is also one further difference between the quick and non-quick plastic clays and that is the organic content of the two units. The total organic content of the plastic (non-quick) clays was measured at between about 1% on the south side of the Drammen River (Bjerrum, 1967) whereas on the north side of the river Moum et al., (1971) reported values of between 0.28% and 0.50% in the quick plastic clays. The nature of this organic material is not stated, although it appears to be amorphous and Bjerrum (1967) has ascribed its presence to the climate amelioration which allowed an increase in biological activity.

8.6 DISCUSSION OF THE COMPARISON SITES IN RELATION TO BOTHKENNAR

As noted earlier, Table 8.4 assembles the data on these sites under three headings: geological data on the depositional setting, age and sea-level history; geochemical data on clay mineralogy, organic content and pore water composition; geotechnical data on particle size, Atterberg limits, yield stress ratio and sensitivity. In total they cover the significant factors described in the
Figure 8.22 Geochemical profiles at Drammen. (Source: Moum et al., 1971.)

Sensitivity, pore water chemistry and K/Na ratio (adsorbed) for boring I

Sensitivity, pore water chemistry and K/Na ratio (adsorbed) for boring II
literature on these deposits and together provide a coherent picture of the geological and geochemical controls that have governed the geotechnical character of the deposits in question.

8.6.1 Comparison of Belfast, Brean and Drammen with Bothkennar

Table 8.4 shows that Bothkennar and Belfast have broadly similar depositional settings. Both are estuarine and neither have undergone sufficient uplift since deposition to be removed from the tidal frame. The upper part of both profiles is believed to be intertidal whereas the lower part of both is subtidal. The deposits are also of similar age (mid- to late-Holocene) and were therefore laid down in relatively warm water. They also have a similar organic content which in both cases was probably derived from estuarine marine biota. The mineralogy of the clay-sized fraction is also similar, being in both cases largely illite-chlorite with some kaolinite. The clay fraction in the Bothkennar sediment also contains inert quartz and feldspar flour, possibly in greater proportion than that at Belfast, whereas at Belfast swelling chlorite was reported (Crooks & Graham, 1976). Finally, neither sediment has suffered a major change in pore fluid composition, other than in the uppermost few metres where fresh water has replaced the originally saline pore water.

Table 8.4 also compares a number of geotechnical characteristics between the sites. Bothkennar and Belfast are very similar in many respects. In both cases the clay-size content is around 40% - 50%, the maximum liquid limit around 90% and the maximum plasticity index around 50% - 60%. At both sites the liquid limit is positively correlated with the pore water salinity. The activity of the clay fraction is around 0.9 - 1.3 at Bothkennar and is said (Crooks & Graham, 1976) to vary between about 1.2 and 2.0 at Belfast except at about 2m depth where it reaches ~3.0. In both cases these values seem excessive for the clay mineralogy involved and the action of the organic component may be implicated. At Bothkennar the liquidity index reaches a maximum between 2.4 and 3.3m bgl, generally declining both downwards (with a local peak at 7.8m) and upwards from that level. At Belfast (Kinnegar b/h No. 1) the liquidity index reaches a maximum at about 3.5m and declines downwards (with a peak at ~7.0m), but generally has a value <1. Kinnegar b/h No. 2, however, had a liquidity index >1. In both cases the maximum value is similar: around 1.4 (Kinnegar b/h 2) and 0.9 (Kinnegar b/h 1) at Belfast and 1.2 at Bothkennar. The yield stress ratios and sensitivities are also similar at 1.5 - 2.0 and 5 - 10 respectively. Thus Bothkennar and Belfast appear to share a geotechnical similarity as well as a geological similarity.

The situation at Brean is rather different. The sediments are intertidal throughout: there has been a history of subsidence throughout the depositional period and the estuary also has a substantial tidal range that encompasses much of the sediment profile. The organic content is largely of terrestrial origin and there are peat horizons within the sequence. The clay-sized
fraction is again an illite-chlorite assemblage, although the illites are stated (Narbett, 1993) to be interlayered with smectite. The pore water chemistry is not stated but leaching seems unlikely in view of the subsidence history. The deposits at Brean show a number of geotechnical differences from those at Bothkennar and Belfast. They are somewhat less plastic (plasticity index: \(-15-30\%\)) and have a lower activity (0.7-1.0), which is more commensurate with their clay mineralogy. Significantly, the liquidity index is much lower in the main clay unit, indicating a more compact structure due possibly to intertidal exposure as suggested by Hawkins et al. (1989). Neither the yield stress ratio nor sensitivity was reported from Brean.

The situation at Drammen is different again. There are two very important differences from Bothkennar: the greater uplift since deposition and the relative paucity of organic material. The sediments formed initially under cold-climate conditions in an actively glacierised fjord (the lean clay): at a later stage conditions ameliorated to some degree and a small organic component was introduced (the plastic clay). Uplift has also raised the sediments above the tidal frame and in some areas this has allowed the original saline pore water to be replaced by fresh water to an extent not seen at either Belfast or Bothkennar. Geotechnically, the lean clay is not directly comparable with the Claret Formation, being similar in origin, composition and properties to the lower glaciomarine clays of the Forth (cf. Chapter Three). In the case of the plastic clay, the clay mineral suite is similar to that at Bothkennar, although the activity of the clay fraction is somewhat lower (0.7 - 1.0). This is probably still a little greater than expected from the mineralogy. At their original salinity, the deposits themselves are less plastic than those at Bothkennar but have a similar liquidity index: after leaching, the plasticity is reduced and the liquidity index raised. This has increased the sensitivity from around 8-10 to 200 or more. Conversely, at the original salinity the yield stress ratio is around 1.6 but is reduced by leaching.

8.6.2 A Framework for Comparisons

The discussion above indicates that it is possible to identify a limited number of key factors which have determined the level of geotechnical structure within the sediments and their overall profile architecture. These factors and the nature of their influence are summarised as follows:

1. The position of deposition relative to the local exposure and tidal frame

It seems reasonable to suppose that the tidal and local exposure conditions under which the sediment was deposited will exert an important influence on the particle packing and hence on the level of geotechnical structure. However, the exact relationship appears to be complex and not clear-cut in practice. Consideration of the comparison sites suggests that those clays which were deposited largely subtidally and have not suffered major uplift (Bothkennar and Belfast) have a maximum liquidity index close to unity in the uppermost part of their profiles. There is
also an increased spread of values at this level. In the case of Bothkennar the position in the profile at which this commences is around the level believed to correspond to low intertidal to immediately subtidal conditions. This wider range of liquidity index then persists at higher levels in the tidal frame. In the case of Bothkennar the void index profile also follows such a pattern, although the maximum structural level is reached at a profile position somewhat prior to the onset of intertidal conditions. This can be seen most clearly at a depth of ~7m in the composite profile published by Hight et al. (1992b) and reproduced in this Thesis as Figure 3.20.

At Brean, where deposition is thought to have been entirely under intertidal conditions in a more exposed setting, and the sediments are somewhat siltier than at Bothkennar, the liquidity index does not exceed about 0.5. This suggests that a higher level of geotechnical structure may result from deposition under conditions of limited wave action in sediments which are likely to have undergone some biogenic stabilisation. This is the situation in the relatively sheltered location at Bothkennar. By contrast, conditions at Brean, which occupies a more exposed location on the Severn estuary, seem to have caused some compaction of the sediment packing leading to a lower level of structure.

It may be noted that at Bothkennar the yield stress ratio is higher (~2.5) in the upper, intertidal part of the profile than in the lower, subtidal part (~1.6). At Brean, where the whole profile may be intertidal, the yield stress ratio is around 2.5 in the upper part and falls to just above unity at depth. This indicates that the YSR can act as a proxy measure of geotechnical structure although this may not be as straightforward as might otherwise be supposed.

Below the intertidal zone the yield stress ratio is around 1.4 - 1.6 throughout both the profiles at Bothkennar and Belfast and in the plastic clay at Drammen. In this last example, Bjerrum (1967) argued that this was the result of ageing of the deposit. The lean clay did not show this effect and thus it may require some minimum plasticity in the clay, perhaps as a result of the organic component, which is absent in the lean clay. This may be related to the results of recent modelling by Tovey and Paul (pers. comm.) which suggests that during continuing deposition the pore water pressure at depth in a sediment column does not fully dissipate but is 'pumped up' by successive load increments at the surface. Once deposition ceases, it then will dissipate in accordance with normal Terzaghi consolidation theory; however, Tovey and Paul further hypothesise that the structure might by then have become locked, possibly by an organic cement, in which event although the pore water pressure will dissipate, the void ratio will not reduce by the expected amount. This could have the consequence that the compression curve would show a step when the cement gave way, which might then be interpreted as a raised yield stress and hence the yield stress ratio would appear greater, although why a consistent value of around 1.5 should emerge remains unclear.
2. The change in relative sea-level with respect to the sediment profile and the consequent change in position relative to the tidal frame.

Sea-level change determines the gross facies architecture of the profile. For example, at Bothkennar there has been a continuous fall in sea-level which has preserved the bedded/mottled facies pattern within a dominantly fine grained sequence. By comparison, at Brean, where sea level rose, facies changes involve greater variation in grading and the sediments were subject to cut and fill episodes (indeed, the whole Severn architecture is based around this). Rising sea levels also promoted the intercalation of peat deposits which are characteristic of many southern UK estuaries.

Changes of sea-level also control the history of inter-tidal exposure and groundwater hydrology. The exposure history (the severity and timing of intertidal desiccation episodes) will have determined the stress history and will have caused overconsolidation and a rise in yield stress ratio. This is clearly important if the sediment has moved through the tidal frame, as it must have done if it has become a terrestrially exposed surface. This leads not only to a visible desiccated crust but also to a less visible underlying zone (Skempton's zone II) with its characteristic profile of water content. Changes in groundwater hydrology also lead to sensitivity changes and development of quick clays in soils of susceptible mineralogy, as illustrated at Drammen.

3. The dominant mineralogy of the clay-sized fraction, whether rock flour, inactive or active clay minerals.

This factor is intrinsic to the site in question, since it reflects the lithology of the source rocks. The comparison sites were selected to be consistent in this respect, although Brean perhaps contains less rock flour. This factor is reflected in the activity of the clay-sized fraction and in particular in the agreement between the activity and the mineralogy: in the cases of all the sites except Brean there is the suggestion that the clay fraction is more active than would be expected from the mineralogy alone. Additionally, the production of a quick clay condition seems to be limited to sites which lack active clay minerals: this is so at Drammen and also in the Canadian Leda clay of the St. Lawrence basin (Quigley, 1980; Torrance, 1983; 1990; 1995; 1998;).

4. The presence of amorphous organic material.

The presence of amorphous organic material causes significant changes to the geotechnical behaviour in both the intact and remoulded states. This has already been discussed at length in Chapter Seven for the case of Bothkennar, where the material is believed to be of estuarine [marine] origin. The position at Belfast is not so clear, since the nature of the organic material is not described, although the general setting would indicate at least some marine component.
Certainly the reported values of activity from the Belfast deposits appear difficult to explain in terms of the mineralogical composition of the clay-sized fraction alone. This is also the case in the plastic clay at Drammen, where an instructive contrast is presented with the underlying, less active, lean clay (as is also the case in the Forth). In the case of the Brean deposits the activity appears less incompatible with the mineralogy, which includes some swelling minerals.

5. The change (if any) in the ionic composition and concentration in the porewater since deposition.

The general role of changes in pore water salinity has been discussed above. The essential feature is the change in plasticity with concentration and the exchange of cations by leaching. The latter promotes the development of a quick clay condition, although it has been observed (Moum et al. 1971; Torrance, 1974, 1975, 1979, 1983, 1988, 1995, 1998) that this association is not always straightforward. For example, a distinction must be made between an increase in intact strength (possibly due to cementation) and a reduction in remoulded strength: both will raise the sensitivity. Also the presence of divalent cations in quite small concentrations can be sufficient to inhibit the development of a quick clay condition, even at low overall salinities.

Drammen is the only one of the comparison sites at which a quick clay condition is found. More generally, quick clays do not occur in the UK, despite the similarities between the source mineralogies of Scotland, Scandinavia and eastern Canada. It is instructive to ask why this should be so. From this work two possible reasons emerge: amorphous organic material may have some role in maintaining the remoulded strength and/or there has been insufficient rapid, uplift to raise the lower, subtidal part of the profile into the freshwater zone. At both Bothkennar and Belfast the upper, freshwater layer occupies only the intertidal part of the profile.

8.7 SUMMARY

The Claret Formation at Bothkennar records a period of falling sea level during which the conditions of deposition changed from subtidal to intertidal, as indicated by the facies succession and the gross sedimentology of the deposits. The water depth history can be quantified by a comparison of the sea-level curve with a seabed curve constructed in an analogous manner. In both cases corrections have been made to the index points, including those for autocompaction using a geotechnical method developed for this study.

There is some evidence that at Bothkennar the Claret Formation can be subdivided into four stratigraphic units, although their regional extent is unknown. These stratigraphical units have a geotechnical significance since they are based on fabric features which impose controls on the
geotechnical properties. For this reason, it is possible to define an engineering stratigraphy at Bothkennar which closely follows the geological stratigraphy at the site.

The detailed sedimentology and SEM microfabrics indicate that the depositional history has generated features of possible geotechnical importance. These are encompassed by the concept of geotechnical structure, which arises from the spatial fabric of the sediment and thus may differ between geological facies defined at various scales. Furthermore, the level of geotechnical structure can vary with position in the profile, leading to the conclusion that it is governed by events such as rate and continuity of deposition, development of intraskeletal and secondary surfaces (hiatuses) and secondary ageing of the material.

Geotechnical profiles from Bothkennar typically show the three zones first described by Skempton (1948). Comparison of the geotechnical profile with the water depth model suggests that zone I developed under subtidal conditions and that zone III developed under intertidal conditions. The status of zone II is less certain, but it is seems probable that it developed under very shallow subtidal to marginally intertidal conditions and is thus the product of the transition between the two.

Comparison of the results from Bothkennar with those published from other sites at Belfast, Brean (Severn) and Drammen (Oslofjord) indicates that it is possible to identify a limited number of key factors which have determined the level of geotechnical structure within the sediments and their overall profile architecture. These are: the initial depositional level; subsequent changes of sea level; the composition of the clay-sized fraction; the presence of amorphous organic material; the ionic composition of the pore water.
CHAPTER NINE

CONCLUSIONS

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CHAPTER NINE

CONCLUSIONS

9.1 SUMMARY OF OBJECTIVES

This programme had four principal objectives that followed from the previous investigations both at the Bothkennar Soft Clay Research Site itself and in the Forth area more generally. These were:

- To develop a depositional model for the sediments based on the Holocene evolution of the Bothkennar area and so to erect a local stratigraphy with engineering geological relevance;

- To describe the sediments of the Claret Formation at Bothkennar in terms of their composition, macrofabric and microfabric and to explain these by reference to the depositional model;

- To describe selected geotechnical properties of these sediments in their intact and remoulded states, and to explain these by reference to their source materials, the geological processes by which they had accumulated and to their subsequent post-depositional history;

- To compare the geotechnical character of the Bothkennar sediments with those of other soft clays of Holocene age from analogous areas, to identify similarities or differences and suggest the reasons for these.

9.2 STRATIGRAPHICAL SEQUENCE AND CONDITIONS OF DEPOSITION

At Bothkennar, the Claret Formation can be subdivided into four stratigraphic units on the basis of sedimentary and biogenic structures. These units have been termed the upper, middle, lower and basal divisions. Published evidence suggests that on the larger scale the Claret Formation may be diachronous and associated both with rising or falling sea level, depending on the local age of the deposit and the position of the contemporaneous depocentre relative to the basin margins.

At Bothkennar the Claret Formation records a period of falling sea level during which the conditions of deposition changed from subtidal to intertidal. This is indicated by the facies succession and the gross sedimentology of the deposits. The detailed sedimentology and SEM microfabrics indicate that deposition was not continuous and that interruption has generated features of possible geotechnical importance.
The water depth history can be quantified by a comparison of the sea-level curve with a seabed curve constructed in an analogous manner. In both cases corrections have been made to the index points for autocompaction using a geotechnical method developed for this study.

The sediments were derived from glacigenic sediments, themselves derived from bedrock sources that contained a generally inactive clay mineral suite.

Deposition occurred under tidal conditions in a broad estuarine embayment, probably with a moderate tidal range; wave action was probably limited throughout most of the period, other than episodic storms, although it was probably more significant during the earliest and the later phases of deposition.

Other than at the immediate base of the succession, sedimentation occurred between around 5,000 and 3,000 $^{14}$C yrs BP. The water depth reduced throughout this period from $\sim$20m below contemporaneous MLWST to intertidal. The age of the basal sediment is unknown, although it post-dates the Bothkennar Gravel Formation ($\sim$11,000 yrs $^{14}$C BP) and sedimentological evidence indicates that current activity may have been significant, thus indicating relatively shallow conditions during deposition.

The mean rate of deposition varied from around 11mm yr$^{-1}$ in the lower part of the sequence to around 7mm yr$^{-1}$ in the upper part.

The sediment is variably bioturbated, as indicated by the density and style of mottling: the bioturbation is limited in the lowest parts of all the cores studied at Bothkennar and generally increases upwards until the intertidal sediments are encountered. At this level in the sediments there is an increased frequency of larger biogenic structures. Mottling appears to be associated with the finer-grained sediments: there is some evidence that it is suppressed by an increased modal size in the silt range.

9.3 SEDIMENT FABRIC

On the basis of primary sedimentary structures and the nature and extent of bioturbation, the sediments of the Claret Formation can be divided into three principal macrofacies: the bedded facies; the mottled facies; and the laminated facies. Biogenic structures are found in all three facies. The microfabric of the sediment can be categorised into three principal types and there is a limited correspondence between the type of macrofacies and the relative proportions of the microfabrics in samples taken from it.

The bedded facies is believed to be the product of primary sedimentation under subtidal to intertidal conditions. As the name suggests, the principal sedimentary feature is a centimetre-scale bedding (best visible on X-radiographs), within which individual beds often
show a well-defined density profile from the base (higher density) to the top (lower density). Beds may be separated by visible erosion surfaces indicated by siltier laminae and ripple-marks. There exists a basal subfacies in which these surfaces and their associated laminae are better developed and more frequent. There exists also a burrowed subfacies, restricted to the top of the sequence, in which macroscopic burrows are common, together with evidence of minor erosion such as mud-flakes and frequent laminae. Frequent mud-clasts, possibly molluscan faecal pellets, are also common. This subfacies is believed to be the product of immediately subtidal to intertidal conditions.

The mottled facies is a bioturbated equivalent of the bedded facies and intermediate forms are common. The mottling occurs in three distinct styles, here termed types I, II and III, which have been used to define three subfacies. These subfacies often succeed one another in a cyclic manner, from Type I at the base to Type III at the top, although many cycles are incomplete. It is likely that a 'cycle' represents the depth structure of the infaunal community. X-radiographs and densimetric profiling have shown that the effect of bioturbation is to homogenise the sediment to a variable extent, although there is not always a clear correlation with the density of mottling.

The sediments of the laminated facies occur as the infill to channels eroded within the sequence. This facies is characterised by millimetre-scale silt-clay laminae and in bulk is noticeably more silty than the bedded facies. Individual laminae are normally a few millimetres thick and their contacts commonly show a variety of evidence of local erosion. They are very clearly visible on X-radiographs and the siltier laminae appear as sharp peaks on density profiles. The facies also contains units of more clayey sediment, sometimes mottled. The facies is interpreted as a deposit formed within tidal channels cut into the main body of the Claret Formation.

Biogenic features occur throughout all the facies. In addition to the mottling, a variety of burrow structures are also found, ranging from larger shafts and U-shaped burrows down to fine anastomosing structures. There appears to be evidence both of mutual association and mutual exclusion and the relative frequency of the structures varies with position in the sediment profile. Comparison of the various fossil features with the present invertebrate fauna of the Forth mudflats suggests many of the organisms possibly responsible for them.

Examination of the sediments under the SEM has allowed the microstructure to be categorised into three types, termed the aggregated, the granular and the cumulate microfabrics. They are considered respectively to have originated by the accumulation of flocculated/pelletised particles; by the selective winnowing of fines to leave a siltier unit; and by the biological reworking of the sediment under conditions of reduced sedimentation. Although the results are not unequivocal, it is suggested that the aggregated microfabric may be relatively more frequent
in the mottled facies and in the bedded facies away from silty partings; the granular in the
bedded facies; the cumulate microfabric is less frequent overall but is most common in the
mottled facies.

9.4 SEDIMENT COMPOSITION AND GEOCHEMISTRY

The sediments of the Claret Formation are composed of a mixed mineral suite that comprises
quartz, feldspar, mica (illite, biotite and muscovite) and chlorite, with other ferromagnesian
minerals, sulphides and oxides as minor constituents.

These minerals are all typical of the mixed rock suite from which the Claret Formation has been
ultimately derived. Quartz is ubiquitous in the sedimentary and igneous rocks of the area,
except in the olivine basalt lavas. The Dalradian metasediments (grits and mica schists) north of
the Highland Boundary Fault and the sedimentary Old Red Sandstone strata of the Midland
Valley are possible sources of muscovite, biotite and Fe-chlorite. Layered minerals are present in
the sedimentary Carboniferous strata and include both muscovite and biotite. Plagioclase and
alkali feldspars are both common constituents of the igneous and sedimentary rocks found
widely in central Scotland. The widespread Devonian and Carboniferous lavas are a possible
source of ferromagnesian minerals including olivine and pyroxene, whilst amphibole is found in
the metamorphic rocks of the highland area.

The sediments are almost entirely silty clays or clayey silts with a sand content usually less than
5%. In general terms there is little difference in overall grading between the bedded and
mottled facies, despite their obviously different appearance. If the burrowed subfacies is
excluded then the mean values of modal size are statistically different (at the 95% level or
greater) between the mottled and the bedded facies. However, the mean values of their median
size and their clay content are not statistically different. The variances for all three parameters
are also not statistically different.

The grading profiles can be divided into four general depth zones (base to 14.7m, to 10.6m, to
4.1m and then to the Cerastoderma bed) based on values of the median size, the modal size and
the percentage of clay and the width of the envelope which contains the profiles. It is argued
that these zones have a genetic origin that relates to the changing water depth and tidal
conditions during the depositional period.

The cumulative grading curves for all facies (and subfacies) are contained within a narrow
envelope, except in the case of the burrowed subfacies, where the spread is much greater. The
overall shape of the curve arises from the polymodal distribution of particle size, a probable
consequence of their original glaciogenic origin, allied to differences in the dominant modes
between the samples. The relative size of these modes, however, varies considerably between
the samples. By contrast, differences in the shape of their cumulative grading curves are less obvious and the value of their median diameter ($d_{50}$) also varies much less.

The organic material in the Bothkennar clay is derived largely from estuarine organisms: plant tissues such as leaves, stems and root fibres are found only rarely. The material occurs mainly as mucal sheets or amorphous coatings which cement individual soil particles into larger aggregates or pellets. Total organic material comprises between about 2% and 4% by weight. At the average sediment accumulation rate at Bothkennar this value equates to an organic carbon accumulation rate typical of modern estuaries. The variability in the percentage of total organic material can be explained by variations in the sedimentation rate and does not necessarily imply major changes in organic productivity.

The composition of the organic component can be characterised in terms of three gross parameters: monosaccharide residues, Kjeldahl nitrogen and a methanol-toluene extractable fraction. The weight percentage of monosaccharide residue varies with depth by a factor of around two whereas organic nitrogen is relatively constant. The average monosaccharide content is typical of many inshore sediments. The methanol-toluene extractable content is more variable and generally correlates with the total organic material.

The C:N ratio is relatively low and indicates a marine rather than a terrestrial origin for the material. If the weight percentages of monosaccharide residues, Kjeldahl nitrogen and methanol-toluene extractable material are plotted on a ternary diagram, the compositions fall along a line on which the nitrogen and monosaccharides maintain an almost constant weight ratio. The reason for this value is unclear, although it is possible that it arises from some particular (unidentified) amino-carbon structural material.

Profiles of acidity (pH) and redox potential (Eh) have been obtained in the upper part of the sequence. Although close to neutral throughout the profile above 6m, the pH shows a slight but perceptible increase with depth from slightly acid at surface to slightly alkaline at depth. Below about 2m depth, the groundwater alkalinity is constant at around pH 8, probably due to the presence of buffered, saline water and possibly also to the presence of comminuted shell debris. The redox potential at this depth is around zero volts, which is probably a result of the organic content, whose decomposition has created an oxygen deficiency. Above the water table, in the unsaturated, aerated percolation zone, the acidity falls to around pH 6 and the redox potential rises to around 0.5V.

Iron is quantitatively the dominant acid-extractable cation and below the crust comprises about 4% of the dry soil weight (by acid digestion). This proportion shows complex variations with depth. Manganese and titanium show a more constant pattern with depth, although manganese also correlates with iron. Titanium occurs principally as the detrital mineral rutile
and so, generally, it is not well-correlated with the occurrence of iron. The DCB extractable iron is relatively more variable and may be associated with Eh-pH conditions in the upper sediment profile that are appropriate for the formation of amorphous iron oxides and hydroxides.

The full depth profiles of three major exchangeable cations (sodium, potassium and magnesium) are generally similar in form and overall conform to an approximately parabolic shape. By contrast, calcium is approximately constant with depth. A more detailed sodium profile above a depth of about 6m shows two distinct zones. There is a steady decrease in sodium content from 6m depth to around 2m depth, which may be due to the infusion of fresh water from precipitation onto the ground surface and perhaps also from the adjacent, slightly higher, caselands to the west. This zone of infusion appears to be broadly coincident with the iron pyrite oxidation zone and may thus reflect the input of oxygenated waters from the surface. The higher zone, from 2m depth to surface, contains almost no sodium and is interpreted as a fully desalinated zone, developed largely by vertical leaching in response to precipitation and improved soil drainage.

The ratios of the major cations against sodium show that although the pore water is likely to have originated as trapped estuarine water during deposition, it has undergone subsequent modification leading to the relative loss of sodium and enhancement in potassium. It is proposed that these effects are respectively the results long term diffusion and the breakdown of potassium-bearing minerals such as feldspars and micas.

9.5 GEOTECHNICAL PROPERTIES

In its natural state Bothkennar clay is a soil of medium to high plasticity. The liquid limit varies considerably with depth, having a maximum value of >80% at around 7m to 10m depth and reducing to a minimum of ~50% both above and below this level. The plastic limit shows much less variation. The index properties differ little between the bedded and mottled facies. There is no significant inter-facies difference in the variance or the mean of any parameter.

In its natural state the Bothkennar clay is more plastic than expected from its clay mineralogy. The plasticity is considerably reduced both by air-drying or treatment with hydrogen peroxide. The liquid limits for the air dried and treated samples showed a marked decrease from their corresponding value from natural water content, whereas the plastic limits show only a relatively small change. This reduction is interpreted as the result of altering (air drying) or almost completely destroying (peroxide) the organic component of the sediment.

Although the Atterberg limits correlate only poorly with the total organic content, when the organic material is characterised in terms of separate monosaccharide, nitrogen and methanol-toluene extractable components, the liquid limit shows a statistically significant positive correlation with the first two of these components and the plastic limit shows a statistically
significant positive correlation with the nitrogen component. Surprisingly, the strength of the correlations is nearly always increased by treatment with hydrogen peroxide, perhaps due to selective removal of some part of these components.

There is a strong, positive correlation between the index properties and the concentration of each of the cations sodium, potassium and magnesium. In the cases of the plastic limit and the activity, this correlation is increased after treatment with hydrogen peroxide. It is decreased in the case of the plasticity index and shows no clear pattern in the case of the liquid limit. The plasticity index and activity are negatively correlated with calcium after peroxide treatment.

It is proposed that at Bothkennar the soil plasticity is controlled by the grading, adsorbed cations and by organic components. This model can be termed the grading-salinity-organics model. It is proposed that the baseline plasticity of a soil is determined by the mineralogy of its clay-sized fraction and that this is expressed by the clay-fraction activity. From this baseline the plasticity will develop owing to the effect of each of the components.

The grading component causes the envelope of values to spread around the baseline (due to grading within the clay-sized fraction) and to shift downwards due to the presence of non-clay minerals in the finer fraction. The salinity and organic components both cause enhancement of the plasticity relative to the baseline. On the activity chart this is seen as a movement of the plotted positions upwards by some variable amount, the potential movement due to the organics being the greater. On the plasticity chart this enhancement is seen as a limited expansion of the baseline area along the direction of the A-line due to salinity alone and as a larger expansion plus a shift along the A-line when both organics and salinity are involved.

By comparison of the measured bulk density with the measured water content, the mean particle specific gravity was found to be 2.65 and the mean saturation ratio to be 0.989. Both these values appear to be reasonable in view respectively of the mineral composition and the evidence of gas.

In gross form there is a general increase in water content from the water table down to about 4m depth, below which the profile envelope is approximately unchanging down to about 7m depth. From this depth down to 17m the general trend is one of falling water content, there is then a slight increase between 17m and 19m depth, after which values fall off markedly to the base of the profile. This profile does not conform to that expected for a uniform, normally consolidated clay, which suggests that the stress history was not one of simple selfweight consolidation throughout the profile and that longer-term changes in the depositional environment may have been involved.

Profiles of liquidity index show much less variability than corresponding water content profiles. This suggests that the local variability in water content is dominated by changes in index
properties and that the underlying stress-controlled change in normalised water content is more simple. The trend of the liquidity index below about 7m - 8m depth indicates simple selfweight compression, with a reduction from a value close to unity at around 7.5m to a value around 0.5 at 20m depth. Between around 7.0m and 2.5m the liquidity index remains almost constant: there is no reduction in liquidity index with decreasing depth as is seen in the water content profiles. This implies that any imposed intertidal stress history has not strongly overprinted the selfweight trend below a depth of around two metres.

The void index profile follows the sedimentation compression line (SCL) fairly closely downwards from a depth of about three metres, although there is a marked departure at ~8m depth. Similar departures are also seen in other, published, profiles. This suggests that below this depth the sediment is both geotechnically structured (since it follows the SCL) and is normally consolidated, presumably as a result of selfweight compression. Above a depth of three metres there is some indication of a more open structure, and at still shallower depths the profile then falls increasingly to the left of the SCL, indicating a more compact structure. In detail, the profiles can show a more complex structure with the appearance of repetitive subunits, each with an internally consistent pattern. However, the points remain close to the SCL and so the sediment cannot be said to be strongly overconsolidated.

Profiles of yield stress and yield stress ratio show that between 3m depth and the base of the shell bed the value of the yield stress ratio can be as high as 2.5. Below about 3m it is constant at around 1.6, which conventionally indicates that the deposits here are lightly overconsolidated. This pattern may have arisen from a number of inter-related processes. Subaerial exposure during the tidal cycle will cause desiccation and the development of suction stresses that may be expected to be greater at the higher levels in the tidal frame. The relatively constant value below this depth does not relate easily to normal concepts of overconsolidation: however, similar values have been reported from other 'normally consolidated' estuarine sediments and have been explained variously by secondary consolidation and cementation by both organic and inorganic materials.

Below about 4m depth the undisturbed undrained shear strength increases linearly with depth, with some scatter due to facies variations. There is a generally increasing trend from about 18 kPa at 4m depth to 50 kPa at 20m. The sensitivity was typically between 5 and 10. Above this depth, until the water table is reached, the values do not vary greatly, which is consistent with a lightly-overconsolidated material. In the crust the strength increases rapidly upwards to nearly 100kPa at surface, probably as a result of an increased effective stress due to partial saturation.

In the fully remoulded state the undrained shear strength is determined by the liquidity index and is independent of facies. The removal of organic material does not alter this basic relationship above $I_c = 0.99$. Below this value, the untreated soil retains some additional
strength which suggests that the organic component acts as a cement that may either partially survive remoulding or has the ability to partially reform afterwards. In the intact state there is also a relationship between liquidity index and undrained strength although there is considerably more scatter in this case.

9.6 RELATIONSHIP OF GEOTECHNICAL CHARACTER TO GEOLOGICAL HISTORY

In the remoulded state the geotechnical character is controlled by composition in a broad sense. This includes particle mineralogy and grading, cation content and organic materials. In the intact state the principal concept is that of geotechnical structure, which is demonstrated by a resistance to deformation above that seen under the same stress in the remoulded state. Geotechnical structure arises from the spatial fabric of the sediment and thus may differ between geological facies defined at various scales. Furthermore, the level of geotechnical structure can vary with position in the profile, leading to the conclusion that it is governed by geological events such as rate and continuity of deposition, development of intradepositional surfaces (hiatuses) and secondary ageing of the material.

Geotechnical profiles from Bothkennar typically show the three zones first described by Skempton (1948). In the lowest (zone I) the sediment is normally consolidated and the level of structure can be correlated with individual features in the sedimentological profile, such as bedded/mottled facies transitions, hiatus surfaces and erosional events (laminae, channel fill). In the middle zone (zone II) this pattern becomes unclear and the level of structure is more variable, with evidence of increased yield stress and a wider scatter of values in the water content and other packing parameters. In the uppermost zone (zone III) the sediment becomes increasingly overconsolidated and there is a fall in water content upwards towards the unsaturated zone (the desiccated crust). In zones I and II there is increasing sedimentological evidence of more energetic conditions (coarsening grain size, frequency of laminae) and of subaerial exposure. There is also an increased frequency of macroscopic biogenic structures such as burrows.

Comparison of the geotechnical profile with the water depth model suggests that zone I developed under subtidal conditions and that zone III developed under intertidal conditions. The status of zone II is less certain, but it seems probable that it developed under very shallow subtidal to marginally intertidal conditions and is thus the product of the transition between the two.

9.7 COMPARISON WITH OTHER HOLOCENE CLAYS

Comparison of the results from Bothkennar with those published from other sites at Belfast, Brean (Severn) and Drammen (Oslofjord) indicates that it is possible to identify a limited number of key factors which have determined the level of geotechnical structure within the
Chapter 9  Conclusions

sediments and their overall profile architecture. These are: the initial depositional level; subsequent changes of sea level; the composition of the clay-sized fraction; the presence of amorphous organic material; the ionic composition of the pore water.

Those clays which were deposited largely subtidally have a maximum liquidity index around or in excess of unity. The depth at which this occurs varies but, in the case of Bothkennar, corresponds to that calculated at the onset of intertidal conditions in the profile. At this level there is also a noticeable increase in the spread of values and, hence, in the level of geotechnical structure. At Brean, where deposition is thought to have been entirely under intertidal conditions and the exposure possibly greater, the liquidity index does not exceed about 0.5.

The history of relative sea-level change determined the gross facies architecture of the profile. For example, at Bothkennar there has been a simple fall of sea level which has allowed changes in local sedimentation rate to generate the bedded/mottled facies pattern up the sequence and the onset of intertidal conditions as the water depth reduced. By comparison, at Brean, where sea level rose, facies changes involve greater variation in grading and the sediments were subject to cut and fill episodes. Rising sea levels also promoted the intercalation of peat deposits which are characteristic of many southern UK estuaries.

Changes of relative sea level also controlled the history of intertidal exposure and groundwater hydrology. The exposure history (the severity and timing of intertidal desiccation episodes) will have determined the stress history and will have caused overconsolidation and a rise in yield stress ratio. This is clearly important if the sediment has moved through the tidal frame, as it must have done if it has become a terrestrial exposed surface. This leads not only to a visible desiccated crust but also to a less visible underlying zone (Skempton's zone II) with its characteristic profile of water content. Changes in groundwater hydrology also lead to sensitivity changes and development of quick clays in soils of susceptible mineralogy.

The dominant mineralogy of the clay-sized fraction, whether rock flour, inactive or active clay minerals is intrinsic to the site in question, since it reflects the lithology of the source rocks. It is reflected in the activity of the clay-sized fraction and in particular in the agreement between the activity and the mineralogy: in the cases of all the sites except Brean there is the suggestion that the clay fraction is more active than would be expected from the mineralogy alone.

The presence of amorphous organic material causes significant changes to the geotechnical behaviour in both the intact and remoulded states. In the case of Bothkennar the material is believed to be of estuarine origin. The position at Belfast is not so clear, since the nature of the organic material is not described, although the general setting would indicate at least some marine component. The reported values of activity from the Belfast deposits appear difficult to explain in terms of the mineralogical composition of the clay-sized fraction alone. This is also
the case in the plastic clay at Drammen, where an instructive contrast is presented with the underlying, less active, lean clay (as is also the case in the Forth). In the case of the Brean deposits the activity appears less incompatible with the mineralogy, which includes some swelling minerals.

Changes in the ionic composition and concentration in the porewater since deposition alter the plasticity of the sediment and can also allow the development of a quick clay condition. However, Drammen is the only one of the comparison sites at which such a condition is found and, more generally, quick clays do not occur in the UK, despite the similarities between the source mineralogies of Scotland, Northern Ireland and Scandinavia. From this work two possible reasons emerge: amorphous organic material may have some role in maintaining the remoulded strength and/or there has been insufficient rapid, uplift to raise the lower, subtidal part of the profile into the freshwater zone. At both Bothkennar and Belfast the upper, freshwater layer occupies only the intertidal part of the profile.

9.8 CRITIQUE AND RECOMMENDATIONS FOR FURTHER RESEARCH

Inevitably this work has suffered from various problems of both methodology and practicability, as does most research. The writer is fortunate that sufficient time and funding have been available to remedy many of these during the course of the programme. The work has also raised questions that remain to be fully answered and has opened new avenues that remain to be explored. Again, this is normal for any piece of scientific research and, indeed, is a desirable outcome.

There are three areas that it is felt would be worthy of attention in any future programme of this type at Bothkennar or elsewhere:

- To remedy particular gaps in the work programme – to obtain a complete X-ray record of the full profile; to obtain a full profiles of organic components and to characterise them in more detail; to obtain certain key measurements at closer spacing (index properties, geochemical analyses) and to match them better with the local sedimentology; to obtain a larger dataset on the microfabrics/macrofacies relationship for statistical analysis; to investigate the micropalaeontology of the deposits.

- To explore interesting new concepts relating to the Bothkennar sediments – to develop the grading-salinity-organics model by controlled laboratory tests and to elucidate the mechanisms involved; to understand the sedimentological basis for the relationship of macrofacies to microfabrics; to explore further the possible relationship of size modes to the mineralogy of the particles; to refine further the fundamental concepts that underly the relationship of geotechnical structure to sedimentological structure.
To further explore the control exerted by depositional history on the development of geotechnical profiles – this would require one or more comparative studies based on estuarine clay sites elsewhere. Such studies would use the methodologies developed in this work (particle/organic/cation analysis, fabric densimetry, X-radiography, very detailed sedimentological logging) and would match these to equally detailed logging of the geotechnical properties that define structuration (water content, index properties, undrained shear strength, yield stress). The settings chosen would, preferably, already have been the subject of detailed studies on their sea-level and sedimentological histories, as was the case at Bothkennar, thus providing an existing geological framework for the new work.

9.9 CONCLUDING STATEMENT

This work has attempted a detailed correlation of the sedimentology, geotechnical properties and depositional history for a Holocene estuarine clay whose overall depositional setting was well understood. It has shown that the geotechnical character can be explained by reference to the composition, geochemistry and sediment fabric that arise in this setting. A limited comparison with the literature on similar estuarine deposits suggests that this work can provide an initial framework within which to interpret the results of similar studies. The demonstration of such a framework is felt to be a major achievement of this research and one which will be refined in due course by studies on other estuarine clays with geological histories different from that at Bothkennar.
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Sedimentology and depositional history of the Claret Formation (‘carse clay’) at Bothkennar, near Grangemouth

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Synopsis

The estuarine sediments of Flandrian age which comprise the Claret Formation in the Bothkennar area can be divided into bedded, mottled and laminated facies on the basis of primary sedimentary structures, frequency of silty laminae and the nature and extent of bioturbation. The facies can be further subdivided by the style of bioturbation. The facies are very similar in overall particle size distribution and mineralogy. Biogenic traces occur in all the facies and have been tentatively matched to organisms present in the modern fauna of the Firth of Forth. The succession of facies has been controlled by the fall of relative sea-level since the Flandrian maximum and records an emergent sequence from sub-tidal (probable water depth around 20 m) to inter-tidal. In the Bothkennar area four local subdivisions can be recognized in the Claret Formation, which reflect changes in sedimentary environment consequent on the reduction in water depth.

Introduction

The low-lying ground around the head of the Forth estuary is underlain by generally fine-grained estuarine deposits (‘carse clays’) deposited during and after the cal Flandrian sea level maximum. Previous descriptions of these sediments (Sissons 1969, 1970; Francis et al. 1970; Axton and Ross 1983; Browne et al. 1984; Gostelow and Browne 1986; Browne 1987) have shown them to vary from silty clays near the local depocentre to silty sands at the basin margin. In the Grangemouth area, recent stratigraphical proposals (Browne et al. 1993; Paul et al. 1995) have designated them as the Claret Formation (formerly larent Beds of Browne et al. 1984) and Grangemouth Formation. Their deposition commenced around 8000–6500 BP at the head of the Forth estuary (Sissons and Brooks 1971) and near Grangemouth was largely complete by 3000 BP (Paul et al. 1995).

This paper provides a detailed description of the Claret Formation at the EPSRC (Engineering and Physical Sciences Research Council) Bothkennar Soft Clay Research Site (BSCR) on the reclaimed coastal mudflats [NS 771] approximately one kilometre south of the Kincardine Bridge (Fig. 1). The site was established by the former ERC for large-scale geotechnical testing and our present work arises from the ongoing geological investigations and at that time to guide subsequent geotechnical research (Paul et al. 1992a).

In this area the sediments of the Claret Formation are clayey clays which form the ‘soft clay’ unit for which the site was chosen and which has been the focus of geotechnical interest (Hawkins et al. 1989; High et al. 1992). The surface exposure of the Claret Formation is very poor throughout the area and our work is based on the examination of fresh, 100 mm diameter, continuous-flight piston cores (HW series: Fig 1), Sherbrooke sample (SH1: Fig 1) the reexamination of earlier (Hawkins et al. 1989) 60 mm diameter, Delft stocking cores (D series: Fig 1) by Bristol University. On the HW series of cores high resolution, non-destructive X-ray densitometry was carried out using a 160 kV scanning system (Been 1981; Paul 1996); photography was carried out on fresh subsamples cleaned by osmotic knife; X-radiographs were made on thin slabs (nominal 17 mm thickness) using a 55 kV system. Particular attention was given to borehole I, which penetrated the full local succession, plus HW8, which provided a comparative record of the units. The results provide a very detailed picture of sedimentary facies, whose architecture can be related to changes in the palooenvironment as deduced from carbon-14 dating and published sea-level curves. They enable us to propose a subdivision of the Claret Formation in the Bothkennar area which we believe may be extendable more widely around the upper Forth estuary.

Geological framework

The Flandrian succession in the Bothkennar area is outlined in Table 1. The Bothkennar research site (Fig. 1) at an elevation of c. 3 m OD and is adjacent to the estuary on land which was reclaimed from the tidal flat around 1784 (Cadell 1929). The western margin of the site is marked by a very minor cliff which formed in the eighteenth century coastline. About 2 km west of the site is a subdued fall in the carse surface from an elevation of around 10 m OD to around 7 m OD or less. This change marks the position of a Flandrian shoal (probably equivalent to PG3 of Sissons et al. 1966) t
tentatively dated at c. 4000 BP (Sissons 1967, p. 184) and so is similar in age to the middle part of the Claret Formation at Bothkennar, which was deposited before about 3800 BP (Paul et al. 1995). This suggests that the site then lay around 2.5 km from the contemporary shoreline, which agrees with our finding (below) that the contemporaneous water depth was initially around 20 m below LWOST. During the time of deposition of the Claret Formation the water depth reduced and the position of the local shoreline migrated eastwards, until its present position was reached around 3000 BP.

Figure 2 shows the overall succession at the Bothkennar research site. The end late Devensian Bothkennar Gravel Formation, which is present throughout this area (Sissons 1969; Browne et al. 1984; Peacock 1998), occurs below the site at around −12 m to −9 m OD. It is succeeded locally by about 10 m−20 m of micaceous silty clays (Fig. 2) which we place within the Claret Formation on the basis of lithology, radiocarbon age and stratigraphical position relative to the Bothkennar Gravel Formation. The Claret Formation is here terminated by a major erosion surface, colonized by a shell bed principally containing the common cockle (*Cerastoderma edule*), which appears widespread over much of the study area and its environs (Robinson 1993).

Above this surface lie thin (~1 m), stratified clayey silts containing lenses of detrital shell material which we have assigned to the Skinflats Member of the Grangemouth Formation (Paul et al. 1995; Table 1). They represent the inter-tidal to supra-tidal beach which formed above the erosion surface and become banked against the local cliff-line. Above these in turn are unstratified sediments believed to have accumulated as lagoonal deposits following the late 18th century reclamation work, which we assign to the Saltgreens Member of the Grangemouth Formation. Their present-day thickness is around one metre, although contemporary records (Udny 1831) show that their maximum initial thickness was around nine feet (2.7 m). They have thus suffered substantial post-depositional compaction, probably due in large part to the introduction of artificial drainage. Near-surface desiccation and oxidation have subsequently formed a hardened crust that now extends almost to the depth of the shell bed and so involves both the Saltgreens and Skinflats members.

**Sedimentary Facies**

**Overview**

The sediments of the Claret Formation can be divided principally into a bedded and a mottled facies on the basis of their primary fabric and the extent and nature of subsequent bioturbation. Within the mottled facies it is possible to recognize three distinct subfacies based on the style of mottling. We are also able to recognize a finely laminated facies of more restricted extent. The log from borehole HW3 (Fig. 3) shows how the lower part of the succession is dominated by the bedded facies and the upper part by the mottled facies. The log also indicates the existence of an upward fining trend in the lower part succeeded by an upward coarsening which is particularly evident near the top of the sequence.
## Stratigraphy of the Flandrian deposits at the Bothkennar soft clay research site

<table>
<thead>
<tr>
<th>Stratum</th>
<th>Lithology</th>
<th>Interpretation</th>
<th>Type section</th>
<th>Earlier nomenclature</th>
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</thead>
<tbody>
<tr>
<td><strong>Grangemouth Formation</strong></td>
<td>Very dark greyish-brown clayey silt usually lacking visible stratification</td>
<td>Lagoonal deposits formed by artificial impounding during modern foreshore reclamation</td>
<td>Bothkennar HW3 Borehole [NS 9206 8858] between ground surface and 0.7 m depth where it lies on sediments of the Skinflats Member.</td>
<td>Not named</td>
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<tr>
<td><strong>Skinflats Member</strong></td>
<td>Dark grey clayey silt becoming dark greyish-brown upwards with lenses and thin layers of marine shells, chiefly <em>Cerastoderma edule.</em></td>
<td>Inter-tidal and supratidal estuarine beach deposits</td>
<td>Bothkennar HW3 Borehole [NS 9206 8858] between 0.7 m–1.8 m depth where it lies with an erosional base on strata of the Claret Formation and is overlain by the artificially induced alluvium of the Saltgreens Member.</td>
<td>Not named</td>
</tr>
<tr>
<td><strong>Grangemouth Docks Member</strong></td>
<td>Black to dark grey silty clay and clayey silt, often finely laminated. Thin sandy bands. Local bioturbation.</td>
<td>Deposits within a tidal channel</td>
<td>Grangemouth Docks No. 114 Borehole [NS 9466 8382] between 0.75 m–9.82 m depth. Here it lies between fill above and Claret Formation below (Browne <em>et al.</em> 1984 figs. 1 and 2).</td>
<td>Grangemouth Beds</td>
</tr>
<tr>
<td><strong>Claret Formation</strong></td>
<td>Black to very dark grey clayey silt and silty clay, sandier towards base, becoming dark grey at top. Bioturbation common.</td>
<td>Offshore shallow sub-tidal marine to intertidal estuarine deposits</td>
<td>Bothkennar HW3 Borehole [NS 9206 8858] between 1.8 m–20.0 m depth, overlain by the Skinflats Member of the Grangemouth Formation and underlain by the Bothkennar Gravel Formation.</td>
<td>Claret Beds</td>
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<tr>
<td><strong>Bothkennar Gravel Formation</strong></td>
<td>Fine to coarse gravel with sand and clayey silt. Locally cobbles.</td>
<td>Inter-tidal to shallow water marine deposits</td>
<td>Bothkennar No. 3 Borehole [NS 9186 8645] between 11.60 m–15.00 m depth (Browne <em>et al.</em> 1984, figs 1 and 2) where it is overlain by the Claret Formation and underlain by till.</td>
<td>Bothkennar Gravel</td>
</tr>
</tbody>
</table>

*not formally defined (see text for details)*

1Paul *et al.* 1995  
2Browne *et al.* 1984  
3Believed to be wrongly assigned

The sediments are generally silty clays to clayey silts whose particle size distribution shows relatively little variation between the bedded and mottled facies (Fig. 4). Frequency curves from both facies exhibit a limited number of recurrent modal classes (at around 14–15 μm, 8–9 μm, 4–5 μm and 1–2 μm): the relative proportions of these classes vary from sample to sample and so produce minor variations in the shape of the overall distribution curve and in derived parameters such as median size.

The bulk mineral composition was determined by X-ray powder diffraction at about 80 positions in boreholes HW3 (Paul, Wood and Peacock 1992), HW5 and HW7. The principal minerals in the assemblages are quartz, alkali feldspar, plagioclase, biotite/phlogopite (1M), muscovite (2M1), illite (1M) and chlorite (lib): within the Claret Formation the mineral content of these assemblages varies little with either depth or facies type (Paul *et al.* 1992a). There is, however, some variation in the relative strengths of the individual peaks, which possibly arises from the relative variations in modal size in the
silt and clay range mentioned above. Similar, equally uniform, assemblages have been recorded from Claret Formation sediments in the Stirling area (Norcliffe, pers. comm. 1997). We believe that this composition and areal uniformity is a consequence of their derivation from the reworked finer fraction of the regional till sheet of central Scotland. A more detailed account of these findings is in preparation.

Bedded facies

This facies (Fig. 5) comprises black to very dark gray (5Y2.5/1–5Y4/1) silty clay deposits whose individual beds are well defined on X-radiographs and range in thickness from a few millimetres to about ten centimetres, often separated by silty partings on which small scale ripples are occasionally visible. The upper surfaces of the ripples are frequently coated by a micaceous silt drapage which terminates in a sharp upper boundary. The lower contacts are usually sharp, with limited evidence of erosion of the underlying unit. There are occasional thicker laminae (3 mm–10 mm) of medium to coarse silt which may have been deposited under temporarily more energetic conditions. We include in this facies those horizons where limited bioturbation (evidenced by motting) has occurred but primary sedimentary structures remain clearly visible. We have found empirically that this usually corresponds to a relatively low density of motting (perhaps around 5% or less areal coverage on a vertical surface). Near the base of the Claret Formation individual beds are frequently separated by laminae of coarse silt or fine sand and the sediment contains both whole valves and disseminated fragments of species such as Spisula subtruncata and Turritella communis (Peacock, pers. comm. 1998).

Within individual beds the density structure can be resolved by X-ray densitometry (Paul et al. 1996). A detailed comparison of the density profile with the X-radiograph (Figs. 5b, 5c) shows individual inter-bed contacts and silt laminae, and also reveals a 'saw tooth' pattern that arises from an upward gradation from higher density clayey silt to lower density silty clay within each unit. SEM observations (Paul et al. 1992b, 138) have shown that in the lower, denser zone many particles are individual mineral grains of fine to medium silt size, whereas in the upper, less dense zone many 'particles' are silt-sized aggregations of clay platelets (flocs or biogenic aggregates) which are probably hydraulically equivalent to the silt grains of the lower zone. Similar features have been seen in artificially sedimented beds studied by X-ray densitometry (Edge and Sills 1989) and it has been shown that the ability of a clay layer to support the next, overlying silt layer is developed within a few hours (i.e. within a single tidal cycle), presumably by secondary processes such as particle readjustment.
Fig. 3. Facies profile of the Claret Formation for borehole HW3. The bedded facies, which is predominant in the lower part of the profile, is replaced by the mottled facies in the upper part but returns at the top of the succession. The profile of median grain size ($d_{50}$) shows a generally upwards-fining trend until about mid-depth and an upwards-coarsening trend above this level.

Near the top of the Claret Formation the bedded facies occurs only infrequently. However, here it contains rip-up clay flakes and layers of pellets (probably biogenic) which are a few millimetres in size and locally abundant, occurring along distinct horizons above sharp, cross-cutting, erosive boundaries. These distinctive features are not seen in the bedded units lower in the succession and suggest possible sub-aerial exposure and current conditions sufficient to cause local erosion of the sediments.

Discussion. We interpret the bedded facies as the product of sedimentation under quiet, tidally dominated conditions such as those in the Forth estuary at the present day. Individual beds consist of a lower, silty layer composed of individual mineral grains, overlain by a hydraulically similar layer of less dense clay particle aggregates. The slightly coarser sediments at the base of the Claret Formation suggest a succession laid down in predominantly shallow, offshore marine conditions that suffered greater current activity than did the later deposits.

The highest sediments of the bedded facies (above

Fig. 4. Comparison of particle size in the bedded and the mottled facies in borehole HW3.

Fig. 5. Example of the bedded facies (HW7: 4.94 m–5.17 m bgl): (a) Photograph of fresh surface after cleaning with osmotic knife; (b) X-radiograph (positive print); (c) Densimeter trace, showing examples of saw-tooth pattern produced within individual beds (i)–(iv).
about ~3 m OD) show evidence of very shallow sub-tidal to inter-tidal conditions. This includes the flakes and pellets described above, the presence of burrows probably associated with low inter-tidal to sub-tidal species (cf. below) and the completion of the sequence by an erosion surface colonised by intertidal species such as Cerastoderma. These findings also are in agreement with our water depth model (described below) which suggests the onset of inter-tidal conditions from around this depth in the succession.

Mottled facies

The mottled facies (Fig. 6) is composed of black to very dark gray (5Y2.5/1–2.5Y5/1) silty clay in which primary bedding and laminations are poorly defined or absent due to bioturbation. The expression of this bioturbation is a lighter motting (gray to olive gray, 5Y5/1–5Y5/2), easily visible on a freshly cut surface (Fig. 6a). We are able to define three subfacies (Types I, II and III) on the basis of the size and style of the motting. On exposure to the atmosphere, oxidation of monosulphide in the surrounding, darker material soon occurs and within about 30 minutes the motting is no longer visible.

X-radiographs (Fig. 6b) reveal that motting is associated with internal reworking of the sediments, which normally causes the partial or complete loss of primary sedimentary structures. The extent of reworking is indicated by the densimeter trace: in this example (Fig. 6c) the density profile in the mottled facies is almost uniform. The change from the bedded to the mottled density signature occurs over a short distance, which corresponds to the visible increase in the density of motting seen in Figure 6a. In other examples the loss of structure is less complete and the correlation between the intensity of surface motting and the degree of internal reworking appears variable. However, a simple relationship might not necessarily be expected. The visible motting is a chemical alteration of the surrounding sediment, whose extent depends on the local redox environment and rate of alteration. It thus does not always provide a simple trace which can be used to determine the proportion of the actual biogenic disturbance. Also, some reworking appears restricted to individual layers and so, although it creates surface motting, does not remove completely the density signatures of the different layers.

Subfacies. In the Type I subfacies the mottles are relatively large (10 mm–15 mm) and usually appear elliptical or irregular in vertical section. They occur singly or in small clusters which cover up to about 30% of the surface. Individual units of this subfacies are relatively thin (up to around 50 mm). In the Type II subfacies the motting is finer (2 mm–5 mm) and individual mottles are mainly elliptical with some curved, hooked or irregular shapes present. They usually occur in dense patterns that can cover up to about 60% of a vertical surface. In the Type III subfacies the motting is elongated (5 mm–10 mm) but narrow (1 mm–2 mm) and forms a fine, threadlike, sub-horizontal pattern which is usually densely packed, often covering more than 70% of a vertical surface.

The subfacies frequently succeed one another in a

FIG. 6. Example of the mottled facies ((a)–(c) HW7: 4.87 m–4.96 m bgl): (a) Photograph of fresh surface after cleaning with osmotic knife; (b) X-radiograph (positive print); (c) Densimeter trace, showing (i) signature of underlying bedded facies (ii) transition and (iii) total loss of structure in the mottled facies. (d) (HW3: 5.85 m–6.05 m). Example of cyclic succession of mottled subfacies over 200 mm. Regular, cyclic pattern (Fig. 6d). Complete cycles may occupy up to 500 mm, although many examples of shorter,
in complete cycles are seen. Within a cycle, the transitions between subfacies are gradual and combinations of mottle Tyres I/II and Types II/III have been observed. A cycle usually commences with a short (20 mm–50 mm) section of sparse Type I, I/II or II mottles, followed by a gradational transition to dense Type II motling. The cycle is usually dominated by Type II motting; Type III mottles occupy the uppermost few centimetres at most and may be absent. Cycles usually terminate upwards against a more silty bed or laminae, above which a new cycle may often commence.

*Discussion.* We attribute the motling to burrowing, probably by polychaetes and oligochaetes, both of which groups are known to be common in the Forth mudflats at the present day (see below). Although polychaete burrows can vary greatly in shape, size and configuration, they usually have very smooth walls with a mucal lining (Howard and Frey 1973) and we have found that the central tubes which occur occasionally in some mottles also have similar smooth walls and mucal coatings. We note that although it has been suggested that the mottles were caused by plant root systems (Hawkins *et al.* 1991), this explanation appears less likely to us in view of the water depths involved (below) and the general absence of plant material associated with the mottles.

The succession of the subfacies through Types I to Type III mottles is believed to reflect the depth structure of the infaunal community responsible for the motting (see below). Thus the larger sporadic Type I mottles represent organisms able to exist in larger, well irrigated, burrows below the oxygenated surface layer, whereas the smallest, dense, mainly horizontal (Type III) mottles represent organisms in very small burrows and thus confined to the near surface layers. The bulk of the organisms (responsible for Type II mottles) exist at intermediate depth. If the community is disturbed by an influx of sediment, it becomes reestablished at a higher level, so creating an apparent cycle of motting.

**Laminated facies**

In the southeast of the site (Fig. 1) the upper part of the Claret Formation is cut out by a unit largely comprising sediments of the laminated facies (Fig. 7) referred to the Grangemouth Docks Member of the Grangemouth Formation (Paul *et al.* 1995). This unit occupies a local channel and rests on the underlying sediments with a strongly erosive contact (Hawkins *et al.* 1989). Elsewhere at Bothkennar, similar sediments occur within the Claret Formation as a minor unit of limited thickness (~1 m) in boreholes HW1, HW3, and D1 (Fig. 2). The lower boundary has a strongly erosive lower contact with the underlying sediments (motled facies) but passes upwards into the sediments of the motted facies with a gradational contact. The unit is absent from nearby boreholes D5 and D6 (Fig. 1) and is also believed to represent a channel fill of limited extent.

The facies comprises very numerous black to dark gray (5Y2.5/1–5Y4/1) clayey silt layers (Figs. 7a, b), usually around 5 mm–10 mm thickness, separated by thin laminae (typically 1 mm–4 mm thickness) of clean, medium to fine silt. The silt laminae have sharp bases with evidence of erosion of the underlying clay layer. Individual laminae are often lenticular in form and may contain minor, symmetrical ripples of height ~3 mm and length ~30 mm. Each succeeding silty clay bed usually has a graded base and fines upwards to be terminated by an undulating, eroded upper contact. Within the unit there are two sporadic layers (~20 mm) of fine sand which rest on the underlying beds with a strongly erosive contact. The thicker silty clay horizons occasionally show local evidence of bioturbation, usually in the form of limited Type I or II motting essentially similar to that described above. The degree of disturbance to the primary sedimentary structure varies; where the motting is relatively sparse disturbance is minimal.

On the densimeter profile (Fig. 7c) the thicker silt laminae normally appear as strong peaks, whereas the thinner laminae are below the nominal resolution (~2 mm) of the instrument. The magnitude of the peaks are generally higher than in the bedded facies and the majority of the larger peaks are more nearly symmetrical, which suggests that the density is uniform across an individual lamina. Conversely, the intervening silty clay beds show the low-relief saw-tooth pattern typical of graded units like that seen within the bedded facies.

*Discussion.* The finely laminated silt-clay alternation, together with the symmetrical ripples and lenticular form of the beds, suggests that the sediments of the laminated facies were deposited in a tidal setting involving transitions from bedload to suspended load transport during the tidal cycle (Reineck and Wunderlich 1968; Klein 1971). The erosive contacts and apparently linear subsurface form of
the unit suggest the fill of a tidal channel analogous to modern day counterparts nearby.

Biogenic structures

The sediments in all the facies exhibit a variety of biogenic structures. These include not only the pervasive bioturbation which is associated with surface motting, but also U-shaped burrows; funnel-shaped structures of varying size; large, irregular shaft-like structures and various anastomosing systems of fine burrows with multiple side branches and end chambers. Several of these structures tend to be associated with particular facies and to vary in frequency with depth in the succession.

Examples of these structures and their relationships to the host sediment are shown in Figure 8(a–d). Funnel structures (b1: Fig. 8a, b) are 5 mm to 30+ mm in le 39th and show incremental internal layering which may be evidence of escape through an aggrading sediment bed. U-shaped burrows (b2: Fig. 8a, b) are usually infilled with fine to medium silt which shows little internal structure and terminate against a bedding surface, often marked by a silty lamina, and penetrate other bedding surfaces with little disruption. Anastomosing burrow systems (b3: Fig. 8c, d) lack any infilling and appear to be a late development which cross-cuts other biogenic structures, particularly the pervasive bioturbation.

These structures cannot be attributed unambiguously to particular organisms. However, comparison with the modern infaunal ecology of the Forth estuary (Elliot and Kingston 1987; Kingston pers. comm. 1995; McClusky 1987; Moore 1987) enables us to make some tentative deductions, based simply on the most common candidates reported by the above authors. On this basis it seems likely that the U-shaped burrows are probably produced by the amphipod Corophium. Funnel-shaped structures occur over a range of sizes which suggests that more than one type of organism may be responsible: candidates include the small gastropod Hydrobia and a number of the larger polychaete worms. The large, irregular shafts are believed to be sections of deep burrows produced by bivalves, one example of which (probably Macoma) was found at the base of a shaft. The pervasive bioturbation is attributed both to polychaetes (e.g. Nereis, Polydora) and meiofaunal oligochaetes (e.g. Tubificoides) and is associated with visible motting as discussed earlier. Anastomosing burrow systems are very characteristic of decapods such as Upogebia and Callianassa (Howard and Frey 1973).

These structures show an inter-relationship that appears to be determined by depth and particle size. Figure 9 shows in detail, for borehole HW7, the nature and extent of biogenic disturbance in relation to the sedimentary profile, using the methodology of Taylor and Goldring (1993). We find that discrete burrows are associated with slightly silty layers and are most frequent in the uppermost 1 m–2 m below the Cerasiasterina bed; at lower depths discrete burrows are more sporadic. Pervasive bioturbation is confined to the more clayey layers, where it obliterates all other structure except the anastomosing burrow systems. There appear to be two separate associations (A and B in Fig. 9), comprising in (A) the discrete funnel and U-shaped burrows and in (B) the anastomosing burrows and pervasive bioturbation. This suggests either a temporal succession, in which the discrete-burrow association is replaced by pervasive bioturbation, itself later to be overprinted by the anastomosing-burrow system, or that instead there has been mutual exclusion between the two associations, controlled by particle size or by interspecies interference such as that observed between Corophium and the polychaete Nereis (Olafsson and Persson 1986).

Discussion: Water depth and facies architecture

The succession at Bothkennar can be understood in terms of a model for the local water depth. Figure 10 shows
part of the eastern Forth valley sea-level curve of Robinson (1993, fig. 22), and also the inferred palaeo-tidal levels based on the assumption that the palaeo-tidal range was similar to that of the present day. Figure 10 also shows radiocarbon dates from the Claret Formation at Bothkennar (Robinson 1993; Paul et al. 1995) plotted against elevation, after correlation for the effect of post-depositional auto-compaction (cf. Paul and Barras 1998). Since the OxA series of dates were obtained on shells considered to be in their life positions, their positions define a seabed path (analogous to a sea-level curve) that shows the change in seabed elevation as the sediment accumulated.

The elevation difference between the seabed path and sea level curve is an approximate measure of the water depth during deposition. Figure 10 shows the mean seabed path: other possible paths could also be constructed to allow for the episodic sedimentation likely in a tidal estuary. A more realistic path might thus follow a staircase pattern around the mean line, constrained by the geological evidence of the bedded to mottled facies transition.
We calculate that the combined uncertainties in the sea-level curve, depositional seabed path and subsequent seabed auto-compaction together probably cause, at most, an underestimate of 5 m to an overestimate of 2 m in the water depth.

Figure 11 shows the relationship of the water depth to the detailed succession in borehole HW3. It suggests that the water depth at Bothkennar fell throughout the period of deposition of the Claret Formation, reducing from around 20 m depth relative to LWOST at the base of the formation to inter-tidal (2 m–3 m above LWOST) at the time of the Cerastoderma bed. The rate of depth reduction appears to have been relatively constant for much of the period and then to have declined sharply towards its close. It also seems to have been largely controlled by the rate of sediment accumulation (from ~11 mm/a between 5000 BP to 4000 BP, reducing to ~7.5 mm/a between 4000 BP to 3000 BP: Paul and Barras 1998), rather than by the rate of fall in sea-level due to continuing isostatic uplift (~4 mm/a reducing to ~1.5 mm/a over the same periods; Robinson 1993). There is a general increase in bioturbation higher in the sequence which accords with the fall in mean sedimentation rate during the depositional period. We suggest that the relatively high proportion of the bedded facies lower in the sequence indicates a period of sustained, possibly even rapid, deposition, whereas the general increase in bioturbation higher in the sequence indicates periods of reduced sedimentation associated with a continued reduction in water depth.

Upwards from about ~2 m OD the number of discrete burrows increases sharply (cf. Fig 10) as does the frequency of silt laminae and evidence of probable sub-aerial exposure such as mud flakes, which accords with the onset of nearshore, shallow sub-tidal to inter-tidal conditions suggested by the water depth model. The sequence is completed by an erosion surface upon which lies the Cerastoderma bed, which seems from the model to have formed at around 2 m–3 m above the low water spring tide level. This accords with the present day ecology of Cerastoderma edule (chiefly inter-tidal) and also with the observations of Sissons and Smith (1965) that the modern tidal flat accumulate to within about 1–2 m of the spring tide high water level. Above this surface lie the sandy silts and intercalated layers of shell detritus of the Skinflats Member,
presumed to be a beach deposit derived from the uppermost sediments of the underlying Claret Formation.

The relative frequency of the bedded and mottled facies, together with criteria such as the frequent presence of burrows, silt laminae, mud flakes and pellets, allows us to identify four local subdivisions of the Claret Formation at Bothkennar which we have called the basal, lower, middle and upper divisions. These divisions are detailed in Figure 11. Their names are not formal stratigraphical proposals, since we are not yet able to demonstrate their general applicability outside the study area, however, they do occur in all the cores we have so far studied from Bothkennar and can also be recognized from descriptions in the literature (Hawkins et al. 1989; 1991). For this reason, and since they appear to be genetically related to the water depth model and so reflect the changing palaeoenvironment, we consider it likely that they or their correlatives will occur generally throughout the Claret Formation in the Forth valley.

Conclusions

The Flandrian succession at Bothkennar records a period of reducing water depth in an estuarine embayment. The silty clays which comprise the Claret Formation in the Bothkennar area can be divided into bedded,
mottled and laminated facies on the basis of primary sedimentary structures, frequency of silty laminae and the nature and extent of bioturbation. The mineralogical composition of all the facies is very similar and consists predominantly of quartz, feldspar, mica, illite and chlorite. There are only minor variations in particle size distribution between the facies: in general the mottled facies is slightly finer grained and the bedded units higher in the sequence show a greater variability than those from lower levels. Throughout the sequence there is evidence of biogenic activity, much of which is analogous to that produced by the present day infauna of the Forth estuary.

The facies succession has been controlled by the fall of relative sea-level since the Flandrian maximum and records an emergent sequence from sub-tidal (probable water depth around 20 m) to inter-tidal. This pattern enables the Clarat Formation to be divided locally into four divisions whose character reflects the changes in sedimentary environment consequent on the reduction in water depth. The lowest, basal, division records increased current activity. The overlying lower division is largely composed of the bedded facies and records rapid sedimentation in a sub-tidal setting perhaps 2–3 km offshore. The succeeding middle division contains an increasing proportion of mottled sediments, indicating a reduction in sedimentation. The highest, upper, division contains structures indicative of higher energy, nearshore conditions and passes upwards into an inter-tidal erosion surface overlain by shell detritus and fine beach sediments.

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References


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TECHNICAL NOTE

Role of organic material in the plasticity of Bothkennar clay

M. A. PAUL* and B. F. BARRAS*

KEYWORDS: chemical properties; clays; organic soils; plasticity; sedimentation; soil classification.

INTRODUCTION
Investigations at the EPSRC Bothkennar research site (Hawkins et al., 1989; Hight et al., 1992; Nash et al., 1992a; Paul et al., 1992a) have shown that these estuarine clay sediments combine moderately high Atterberg limits ($W_L \sim 47\%$ to $\sim 85\%$, $W_P \sim 25\%$ to $\sim 41\%$) and a relatively high clay fraction activity ($\sim 0.7$ to $\sim 1.5$) with a suite of inactive minerals (principally chlorite, illite, mica, feldspar and quartz flour) in the clay-sized fraction. This unusual combination has been attributed to the action of organic material and it has been shown (Paul et al., 1992a) that treatment with hydrogen peroxide reduces the activity to a value ($\sim 0.3$ to $\sim 0.6$) more commensurate with the clay fraction mineralogy.

The purpose of this note is to consider the nature and geochemistry of the organic material in the Bothkennar clay and to report further the effect that it has on the Atterberg limits. We find that total organic content is a poor predictor of the liquid and plastic limits and that it is more helpful to examine separately the role of the major geochemical components. The results are consistent with a model based on the adsorption of large organic molecules onto soil particles and the formation of flexible organic cements. We believe that this model can explain other unusual geotechnical properties of the Bothkennar clay and, since it arises from normal estuarine biology, may be of general application to other such soils.

METHODS
Continuous piston samples were collected from 2.43 m to 4.81 m depth (the upper part of the Claret Formation: Paul et al., 1995) from a borehole (HW9) positioned adjacent ($\sim 2$ m) to the site of an earlier borehole (HW3: Paul et al., 1992a). The stratigraphy in the two holes is presumed to be very similar. Subsamples for geochemical testing were removed in an inert (argon) atmosphere and stored at $-20^\circ$C until required. The organic material was characterized using potassium dichromate oxidation to determine the total organic content, sulphuric acid–phenol colorimetry to determine the content of monosaccharide residues, Kjeldahl reduction to determine organic nitrogen and Soxhlet (toluene–methanol azo trope) extraction to determine the "lipid" content. The last three of these are general proxy measures for soil polysaccharides and other carbohydrates; protein, peptide and peptidoglycan residues; and oils, fats, waxes and a large variety of ring compounds, respectively.

Atterberg limits were determined by testing both from the natural water content without prior drying and after treatment with 20 vol hydrogen peroxide. Procedures were otherwise in accordance with BS 1377 (British Standards Institution, 1990). The electron micrographs were taken from those reported previously from borehole HW3 (Paul et al., 1992b).

ORGANIC MATERIAL IN THE BOTHKENNAR CLAY
Occurrence and origin
Figure 1 shows examples of the variety of organic materials and structures to be found in intact samples from the Bothkennar clay. These include mucal structural linings (Fig. 1(a)), amorphous coatings (Fig. 1(b)), agglomerates (Fig. 1(c)) and faecal pellets (Fig. 1(d)). In all of these examples individual soil particles appear to be cemented in to large structures by the organic material. We stress that although the term 'organic' can suggest a fibrous or peaty material of terrestrial plant origin, we have seen plant remains only very occasionally. We believe instead that the major proportion of the organic material at Bothkennar, as is the case in most Holocene estuarine sediments, originates from planktonic (water column) or benthic (bottom-dwelling) organisms.

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Discussion on this technical note closes 5 November 1999; for further details see p. ii.

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The former are responsible for 'marine snow': a general term for detritus composed of inorganic aggregates, organic mucal agglomerates and zooplankton faecal pellets (Syvitski, 1992), evidence of which can be seen frequently in electron micrographs of the Bothkennar clay (Paul et al., 1992b). Studies of the modern benthic biota of the intertidal mudflats near Bothkennar (Moore, 1987; McLuskey, 1987) have revealed that it is dominated by worms (polychaetes and nematodes), crustaceans and molluscs. Many species in these groups produce mucopolysaccharides which they use as structural materials in burrow systems (Howard & Frey, 1973). The intertidal mudflats also support an epipelagic diatom flora, which is known elsewhere to bind sediment surfaces by mucopolysaccharides exuded as a means of locomotion during the tidal cycle (Holland et al., 1974; Frostick & McCave, 1979; Paterson et al., 1990).

Following deposition, these compounds will have degraded in a variety of ways: there will have been a relative loss of nitrogen, proteins and sugars will have been partially consumed by bacterial action and ill-specified 'complexes' will have formed, perhaps using clay surfaces as catalytic templates (Hedges, 1977; Syvitski & Murray, 1981; Harvey et al., 1983). In its final state in the sediment the organic material will thus consist of a mixture of identifiable structural components, amorphous materials and secondary degradation products.
**Geochemical composition**

The total organic material (Table 1) comprises between 2% and 4% by weight of the samples tested, which is typical for the upper part of the Bothkennar sequence (Hawkins *et al.*, 1989; Hight *et al.*, 1992; Paul *et al.*, 1992a). Given an average sediment accumulation rate at Bothkennar of about 10 mm per year (Paul *et al.*, 1995), this value equates to an organic carbon accumulation rate of the order of 200 g/m² per year, a value typical of modern estuaries (Wollast, 1991). The variability in the percentage of total organic material can be explained by variations in the sedimentation rate deduced from 14C dating (Paul & Barras, 1998) and does not necessarily imply major changes in organic productivity. This is consistent with SEM images that suggest occasional condensed sequences associated with an increase in reworked biogenic debris (Paul *et al.*, 1992b).

The weight percentage of monosaccharide residue varies with depth by a factor of around two, whereas organic nitrogen is relatively constant (Table 1 and Fig. 2). The average monosaccharide content (calculated as glucose units) equates to ~27 μmoles/g, which is typical of many inshore sediments (Degens & Mopper, 1975). The lipid content is more variable and generally correlates with the total organic material. These three components together account for only about 60% to 75% of the total organic material by weight and we presume that the remainder comprises materials, possibly of high molecular weight, which are resistant to the procedures employed in this work.

The results indicate an overall C:N ratio in the range 12 to 16, a relatively low value that indicates a marine rather than a terrestrial origin for the material (Table 22.1 of Tyson (1996)) since terrigenous organic material, which is often dominated by cellulose and its products, typically has a C:N ratio up to 100 or higher. When the weight percentages of monosaccharide residues, Kjeldahl nitrogen and toluene–methanol-extractable lipids are plotted on a ternary (three-component) diagram (Fig. 3), in which the apices represent 100% of each component, the compositions fall along a line on which the nitrogen and monosaccharides maintain an almost constant weight ratio of 1:4.37 ± 0.63. If the sugar is a hexose (six carbon atoms), this corresponds to a consistent nitrogen:carbon atomic ratio close to 1:2 (1:2.04 ± 0.30). The reason for this value is unclear, although it is possible that it arises from some particular (unidentified) amino-carbon structural material, such as a mucopolysaccharide (from diatoms) or an amino-sugar (e.g. in chitin from crustaceans, polychaetes and zooplankton). We note, in parenthesis, that amino-sugars are known to form up to 0-2% by weight of some marine sediments (Degens & Mopper, 1975).

**EFFECT ON INDEX PROPERTIES**

In this study (Table 2 and Fig. 4) we have found no significant statistical relationship between the total weight of organic material (measured by the dichromate method) and either the liquid or the plastic limits. Examination of our data from earlier work (Paul *et al.*, 1992a,b; Paul & Barras, 1993) leads to a similar result. Thus total organic content appears to be a poor predictor of the Atterberg limits in this soil, in contrast with peaty soils of higher organic content (Skempton & Petley, 1970). After treatment with hydrogen peroxide we find that, as expected, both Atterberg limits are reduced by the removal of the organic material and that the reduction is greater in the liquid limit than in the plastic limit. The magnitude of this reduction is, however, variable and is not statistically correlated with the percentage of total organic material, which suggests that not all the weight of organic material is effective in changing the Atterberg limits.

However, when individual geochemical comp-

---

**Table 1. Geochemical characterization: borehole HW9**

<table>
<thead>
<tr>
<th>Depth: m</th>
<th>Total organic content: %</th>
<th>Weight: mg per gram of dry soil</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Monosaccharide residues</td>
</tr>
<tr>
<td>2.43</td>
<td>3.9</td>
<td>3.2</td>
</tr>
<tr>
<td>2.61</td>
<td>3.3</td>
<td>5.5</td>
</tr>
<tr>
<td>3.01</td>
<td>2.6</td>
<td>5.6</td>
</tr>
<tr>
<td>3.19</td>
<td>2.4</td>
<td>3.9</td>
</tr>
<tr>
<td>3.41</td>
<td>3.2</td>
<td>4.9</td>
</tr>
<tr>
<td>3.69</td>
<td>3.9</td>
<td>5.9</td>
</tr>
<tr>
<td>4.09</td>
<td>3.1</td>
<td>4.5</td>
</tr>
<tr>
<td>4.26</td>
<td>3.1</td>
<td>6.0</td>
</tr>
<tr>
<td>4.63</td>
<td>3.0</td>
<td>6.0</td>
</tr>
<tr>
<td>4.81</td>
<td>3.7</td>
<td>6.4</td>
</tr>
</tbody>
</table>
Fig. 2. Profile of total organic material and individual geochemical components in Bothkennar borehole HW9.

Fig. 3. Composition of the organic fraction expressed as weight percentages of the three measured components (ms, monosaccharide).

clay activity may thus not yield an obvious outcome.

Since the components act in combination, the relationships shown in Fig. 5 are projections of the multidimensional data onto single variates and true maxima may occur out of this plane. A full statistical description clearly requires a multiple-regression model (which may be non-linear) and possibly a reduction in the number of independent variables by principal-component analysis. We have not considered this aspect further in this note since we feel such an approach would benefit from a larger data set than is available here. For this reason we treat the statistical significance of the linear-regression lines (Table 3) with caution: we note, for example, that inclusion or exclusion of the single point (a) on the toluene–methanol curve (Fig. 5) changes the significance level of that regression substantially.

DISCUSSION

We suggest that the organic material increases the Atterberg limits of Bothkennar clay by two principal mechanisms. Firstly, direct observation of intact samples shows that cementation (Fig. 1) occurs between soil particles when organic material becomes adsorbed onto several adjacent particles simultaneously and so forms flexible bridges, creating three-dimensional void spaces which maintain their integrity at high natural water contents. We hypothesize that some of these bridges may survive the remoulding processes and that others may reform when the organic material comes into renewed contact with the soil particles. Thus the remoulded soil can maintain a higher water content.
at a given undrained shear strength and, since the index tests are a form of undrained shear strength test, this causes an increase in the Atterberg limits. Secondly, the relationship of the Atterberg limits to identifiable geochemical components, rather than to the weight of organic material as a whole, suggests the direct adsorption of water by way of specific organic sites. This process does not involve all the organic material, some of which thus simply provides inert bulk, and so explains the lack of correlation between the total organic content and the Atterberg limits. Both mechanisms are ultimately controlled by adsorption of the organic molecules onto the clay structure and so we surmise that, over a wider range of weight fractions than is found in the Bothkennar clay, the plots shown in Fig. 5 would conform to the shape of an adsorption isotherm (Fig. 6), although in detail the curves for various components would probably differ, since they are likely to be governed by binding mechanisms specific to the functional groups involved (Mortland, 1970).

It is likely that these mechanisms are also responsible for some of the other abnormal geotechnical properties of the Bothkennar clay which have been reported in the literature. We note that the yield stress ratio throughout the deposit is consistently higher (~1.6 and above) than is expected from its known consolidation history (Nash et al., 1992b) and we suggest that organic cementation may be responsible, an explanation also proposed by Hight et al. (1992) to explain differences between the yield envelopes of samples from the different sedimentary facies at Bothkennar. It

### Table 2. Geotechnical characterization: borehole HW9

<table>
<thead>
<tr>
<th>Depth: m</th>
<th>Clay: %</th>
<th>From natural water content</th>
<th>After hydrogen peroxide treatment</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>( W_L: % )</td>
<td>( W_F: % )</td>
</tr>
<tr>
<td>2.43</td>
<td>50</td>
<td>49</td>
<td>28</td>
</tr>
<tr>
<td>2.61</td>
<td>38</td>
<td>47</td>
<td>23</td>
</tr>
<tr>
<td>3.61</td>
<td>44</td>
<td>55</td>
<td>29</td>
</tr>
<tr>
<td>3.19</td>
<td>45</td>
<td>48</td>
<td>24</td>
</tr>
<tr>
<td>3.41</td>
<td>41</td>
<td>60</td>
<td>26</td>
</tr>
<tr>
<td>3.69</td>
<td>41</td>
<td>61</td>
<td>28</td>
</tr>
<tr>
<td>4.09</td>
<td>41</td>
<td>57</td>
<td>27</td>
</tr>
<tr>
<td>4.26</td>
<td>46</td>
<td>65</td>
<td>31</td>
</tr>
<tr>
<td>4.63</td>
<td>44</td>
<td>66</td>
<td>32</td>
</tr>
<tr>
<td>4.81</td>
<td>42</td>
<td>61</td>
<td>27</td>
</tr>
</tbody>
</table>

**Fig. 4. Relationship of Atterberg limits to total organic material in Bothkennar borehole HW9**
We anticipate that the results of this work may be applicable to other estuarine clays deposited under intertidal to shallow subtidal conditions, since they also are likely to contain an organic fraction dominated by organisms such as epipelagic diatom flora, polychaetes and other worms and burrowing crustaceans, which are variously implicated in the production of mucopolysaccharides, peptides, peptidoglycans and lipids. The relationship of the organic content to the environment and geological age of such deposits is beyond our present scope but we observe that a useful engineering classification of soft clay deposits can be based on such criteria.

CONCLUSIONS

This work has shown that the organic material in the Bothkennar clay is derived largely from estuarine organisms and that plant tissues such as leaves, stems and fibres are found only rarely. The material occurs mainly as mucilaginous or amorphous coatings which cement individual soil particles into larger aggregates or pellets. In its natural state the Bothkennar clay is more plastic than expected from its clay mineralogy but removal of the organic material reduces both of the Atterberg limits to values more commensurate with the clay

Table 3. Statistical correlation of Atterberg limits with organic components: borehole HW9

<table>
<thead>
<tr>
<th>Organic component</th>
<th>Liquid limit</th>
<th>Plastic limit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total organic material</td>
<td>Not significant at 95% level</td>
<td>Not significant at 95% level</td>
</tr>
<tr>
<td>Monosaccharide</td>
<td>$r = 0.6794$ ($n = 10$)</td>
<td>Not significant at 95% level</td>
</tr>
<tr>
<td></td>
<td>Significance level &gt; 95%</td>
<td></td>
</tr>
<tr>
<td>Kjeldahl nitrogen</td>
<td>$r = 0.7403$ ($n = 10$)</td>
<td>$r = 0.7647$ ($n = 10$)</td>
</tr>
<tr>
<td></td>
<td>Significance level &gt; 98%</td>
<td>Significance level &gt; 98%</td>
</tr>
<tr>
<td></td>
<td>$r = 0.7955$ ($n = 9$)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Significance level &gt; 98%*</td>
<td>Significance level &gt; 98%*</td>
</tr>
</tbody>
</table>

* Excluding value (a) in Fig. 5. Not significant at 95% level if (a) is included.
mineralogy. Although the Atterberg limits correlate only poorly with the total organic content, when the organic material is characterized in terms of separate monosaccharide, nitrogen and toluene-methanol-extractable components, the liquid limit shows a statistically significant positive correlation with each of these components and the plastic limit shows a statistically significant positive correlation with the nitrogen and lipid components. We believe that this follows from the presence of large organic molecules bound to the mineral particles and so is governed by an underlying adsorption mechanism. If correct, this model will apply to other estuarine clays of similar origin and provides an explanation for certain other geotechnical properties of these soils.

ACKNOWLEDGEMENTS
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REFERENCES


Rapid Communication

A Geotechnical correction for post-depositional sediment compression: examples from the Forth valley, Scotland

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ABSTRACT: Fine-grained sediments usually suffer post-depositional compression (compaction), which reduces bed thickness and lowers the elevations of sample positions and marker horizons. This reduction has been calculated for two general cases using geotechnical theory, and a correction (decompaction factor) is presented that can be applied to many field situations. Its use is illustrated by two examples from the clay sediments of the Clare Formation of the Forth valley. In the first, the correction is calculated for 14C samples from the Bothkennar research site near Grangemouth, and the effect on the reconstruction of water depths is discussed. In the second example, corrections are obtained for published sea-level index points from former shorelines in the Forth valley, using a method for a posteriori estimation of geotechnical data. General conclusions are drawn between the size of the correction and the sample position relative to the local depocentre. Some applications to palaeotidal modelling are also suggested.

KEYWORDS: sediment compression; geotechnical; elevation correction; index point correction; decompaction; Forth valley, Scotland.

Introduction

It is well known that fine-grained sediments undergo volume reduction (compaction) after deposition. As a result, the modern-day elevation of datum points within a sediment profile, such as sample positions, marker horizons, palaeo-surfaces and so forth, may be lower than their original elevation; this may be significant in the construction of sea-level curves, palaeotidal models or calculation of sedimentation rates. In this paper geotechnical theory is used to determine the correction that should be made to restore present-day elevations to their original values, and we discuss applications of this correction to the estuarine sediments of the Clare Formation of the Forth valley in east-central Scotland.

Basic principles and calculations

When a fine-grained sediment is first deposited, it usually contains a large proportion of pore fluid: during burial, this fluid is expelled over a period of time and the porosity of the sediment is reduced. During this expulsion the effective stress carried by the particles will increase until the volume of the sediment is in equilibrium with this stress. In geological usage, this process of volume reduction is usually termed compaction and the corresponding correction decompaction; however, the term is a broad one and can also include additional thermal and mineralogical changes at depth (Audet, 1995). These are not addressed in this paper, which considers only the volume decrease induced by an increase in effective stress.

This volume reduction has both a time dependence and a stress dependence, which arise from distinct processes and are thus treated separately by geotechnical theory. In geotechnical usage the reduction in equilibrium volume due to an increase in effective stress is termed compression, whereas consolidation is the time-dependent expulsion of water while the sediment structure comes into equilibrium at a constant applied load. In the case of the latter, Gibson

* Correspondence to: Professor Michael A. Paul, Department of Civil and Offshore Engineering, Heriot-Watt University, Edinburgh EH14 4AS, Scotland. Email: M.A.Paul@hw.ac.uk
(1958) has shown that the degree of consolidation is determined by the coefficient of consolidation and the rate of sedimentation: most natural sediments, including those discussed below have been deposited at a relatively slow rate and are therefore almost fully consolidated at any stage throughout the process of deposition. This means that it is not usually necessary to consider the time dependence of the compaction process when establishing the magnitude of the correction, unless the rate of sedimentation is exceptionally rapid (as may occur in some deltaic environments: cf. Coleman and Prior, 1978).

Consider sample S (Fig. 1a), which was deposited at an original elevation \( H_0 \) under an initial effective stress \( P_0 \). Below this sample lay a thickness of sediment \( Z_a \) which raised the effective stress to a level \( P_1 \) and compressed the underlying sediment to a thickness \( Z_c \), so reducing the elevation of the sample to \( H_1 \). The vertical compression is \( \Delta H \) and it this which must be calculated to recover the initial elevation. The response of the sediment to this change in loading is shown in Fig. 1b. From an initial position at \( S_0 \), the sample point moves down a compression line to position \( S_1 \); the magnitude of the reduction \( \Delta H \) depends on the ratio (due to the log scale) of the final to the initial effective stresses and on the slope of the compression line. The latter (the compression factor) is correlated empirically with the liquid limit of the sediment (Skempton, 1944): the greater the liquid limit, the steeper the slope.

Using this simple model, we have calculated a correction for sediment beds of various thickness and compressibility using a one-dimensional numerical model with 21 nodes. The calculations were performed on a desktop PC using Microsoft® Excel. For simple self-weight compression (Fig. 2a) the correction scales with the thickness of the bed and can thus be shown in terms of dimensionless variables. The maximum can be 5–10% of the bed thickness near the centre of the bed, because at this point there is an optimum combination of post-depositional load increment and underlying sediment thickness (\( Z_b \) and \( Z_c \) in Fig. 1). At shallower depths the correction is less because the surcharge load is less, whereas at greater depths it is also less because the thickness of underlying compressible sediment is less. Using Fig. 2a the expected self-weight compression at any depth in any given bed can readily be obtained if the liquid limit of the sediment is known, or can be estimated.

A common geological event is the deposition of a relatively incompressible bed above an original compressible layer (Fig. 2b), in which the additional load is modelled as a surcharge, the effect of which is added to the existing self-weight compression. We show for illustration the specific example of a 20 kPa increase (equivalent to ca. 2 m of new sediment) on a bed of 20 m thickness, which is broadly similar to the example that we describe below. The effect of the surcharge is greatest in the upper part of the deposit and declines logarithmically with depth. Its contribution to the total correction is shown (Fig. 2b) by the vectors: near the surface it is about equal to the self-weight compression, whereas at mid depth it contributes only about 5% of the total. The effect is not independent of scale and so a dimensionless figure cannot be produced: each case must be calculated individually. The details of such calculations are beyond the scope of this contribution but follow normal engineering settlement theory as described in most texts on soil mechanics. Details of the implementation in Excel can be obtained from the authors.

![Figure 1](image-url)

**Figure 1** The model situation that is assumed in the analysis. (a) A sample S lies initially at an elevation \( Z_b \) above datum level. The sample is buried beneath a further depth \( Z_a \) of new sediment. This causes compression of the sediment below the sample and reduces its elevation by \( \Delta H \) to \( Z_c \). (b) The geotechnical response of the sediment. At elevation \( H_0 \) the effective stress on the sample is \( P_0 \). Burial to a depth \( Z \) raises the effective stress on the sample to \( P_1 \). In response the underlying sediment compresses along the line from \( S_0 \) to \( S_1 \) and so lowers the elevation of the sample to \( H_1 \).
Applications to the Flandrian sediments of the Forth valley

Palaeoenvironments at the EPSRC Soft Clay Research Site, Bothkennar

Studies of the engineering geology at this research site (Paul et al., 1992) near Grangemouth have led to practical applications of these theoretical models. An integral part of our work has been a detailed study of the Flandrian geological setting of the Bothkennar site in order to establish rates of deposition, water depth changes and patterns of sedimentation during the period from 5000 yr BP to the present. Using $^{14}$C dated marine shells, measured lithostratigraphical profiles and published sea-level data (Robinson, 1993), we have modelled the facies succession and palaeoenvironments at Bothkennar (Paul et al., 1995). Central to the method has been the need to know the accurate original depositional elevations, both of our own samples and of those in the published literature. This has provided the rationale for the corrections described here.

The Flandrian succession is around 20 m thick at Bothkennar (Fig. 3). The deposits largely comprise compressible silty clays of liquid limit 60–80% (Hight et al., 1992; Nash et al., 1992; Paul et al., 1992), which can be subdivided into repetitive facies on the basis of primary sedimentary structures and trace fossil assemblages. The succession records an emergent estuarine sequence, deposited between about 5000 and 3000 yr BP, initially in a subtidal setting (water depth ca. 20 m) that became intertidal as sea-level fell and the estuary silted up. The name Claret Formation has been used for this succession (Browne et al., 1984, 1993; Paul et al., 1995). The Claret Formation is succeeded by sediments of the Grangemouth Formation (Paul et al., 1995), which at Bothkennar comprise a shell bed (principally Cerastoderma) and lenses of shell fragments, followed by 1–2 m of clayey silt which dates from artificial reclamation in the late eighteenth century (Udny, 1831) and thus has surcharged the underlying deposits. Below the Claret Formation lies about 3 m of incompressible sandy gravel (the Bothkennar Gravel Formation), which in turn overlies 5–10 m of sandy glaciomarine muds of low compressibility (the Loanhead Formation), which in turn lie on till.
The post-depositional settlement throughout the Claret Formation has been calculated from the above model (Fig. 3), as have the corrections applied to individual samples (Table 1). The correction at mid-depth is between 2 and 2.5 m, ca. 10% of the bed thickness, although there are noticeable fluctuations from the theoretical curve, which result from the facies successions. This has had a significant effect on the elevation of sample OxA-3388 (Table 1). At the top of the profile the correction results largely from the superposition of the Grangemouth Formation, which accounts for about 1 m of the 1.2 m correction to the elevation of OxA-3507. The effect of the underlying Loanhead Formation increases with depth, and at the base accounts, not surprisingly, for nearly all the correction to the position of sample OxA-3389. This emphasises the importance of these lower beds to the correction, despite the intervention of the incompressible Bothkennar Gravel Formation and some possible overconsolidation of the Loanhead Formation due to the erosion of the Main Late-glacial Shoreline Platform upon which the Bothkennar Gravel Formation rests.

Flandrian sea-level data around Grangemouth

The construction of a sea-level curve relies on accurate determination of the elevation of sample points (index points), for which the relation to the contemporaneous tidal cycle is understood. Elevation errors are especially significant when studying Flandrian sea-level change, because the magnitude of the change can be similar to the magnitude of the errors. A series of \(^1^4\)C dated marine shell samples from the eastern Forth valley was used by Robinson (1993) in the construction of a Flandrian sea-level curve for the area, which we have combined with our own measurements of sea-bed elevation at Bothkennar in order to reconstruct former water depths at the site (Paul et al., 1995). Robinson's index points were not corrected for post-depositional compression and so we have corrected their elevations to take this into account. In so doing we have developed a method to deal retrospectively with these and similar samples, for which the geotechnical setting was not recorded at the time of collection.

There was very little borehole data available for the sample sites. No published profiles of lithology, voids ratio or \textit{in situ} stress were available, nor was there any direct information on soil properties. It has thus been necessary to estimate these data using reasonable geotechnical assumptions and inferences from published regional patterns, which were augmented by unpublished profiles kindly made available by Dr Robinson (pers. comm.).

Appropriate lithological and geotechnical profiles were deduced for each of the (four) sample sites (Table 1) from an earlier regional engineering geological study of the upper Forth valley (Gostelow and Browne, 1986). The depths to the upper surface of the underlying till, (taken as an incompressible basement) and to the incompressible Bothkennar Gravel Formation were also estimated from Gostelow and Browne (1986), supplemented by Sissons (1969).

The corrections we have calculated for these sites (Table 1) are relatively small (less than 1 m) owing to both the relatively shallow depth of burial and, in the cases of profile types 4 and 5, to the coarser and thus more incompressible nature of the deposits compared with Bothkennar. We argue below that this reflects a contrast between a basin-margin and depocentre setting, which we believe to have general importance.
Table 1  Compression corrections for sample positions in the Forth Valley

<table>
<thead>
<tr>
<th>Location</th>
<th>Sample number and date in 14C yr BP</th>
<th>Known* or inferred(^{b}) geotechnical profile</th>
<th>Elevation of top of borehole (m OD)</th>
<th>Sample depth (m bgI)</th>
<th>Correction (m)</th>
<th>Corrected sample elevation (m OD)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bothkennar HW3(^{c}) NS 921 859</td>
<td>OxA-3389 5075 ± 90</td>
<td>See Fig. 3</td>
<td>+3.1</td>
<td>19.6</td>
<td>+0.5</td>
<td>-16.0</td>
</tr>
<tr>
<td>Bothkennar D1(^{c}) NS 921 859</td>
<td>OxA-3388 3825 ± 130</td>
<td>See Fig. 3</td>
<td>+3.1</td>
<td>7.3</td>
<td>+2.1</td>
<td>-2.1</td>
</tr>
<tr>
<td>Bothkennar SH1(^{c}) NS 921 859</td>
<td>OxA-3507 3045 ± 80</td>
<td>See Fig. 3</td>
<td>+3.2</td>
<td>1.4</td>
<td>+1.2</td>
<td>+3.0</td>
</tr>
<tr>
<td>Mid Thorne(^{d}) NS 908 809</td>
<td>SRR-1899 4830 ± 50(^{e}) 4640 ± 50(^{f}) Type 5(^{b}) 4.75 m PG silty clay 0.25 m gravel 5 m LG sandy clay Till</td>
<td>+5.25</td>
<td>0.85</td>
<td>+0.4</td>
<td>+4.8</td>
<td></td>
</tr>
<tr>
<td>Brackenlees(^{d}) NS 913 850</td>
<td>SRR-1540 3860 ± 50(^{e}) 3820 ± 50(^{f}) Type 3(^{b}) 9.0 m PG silty clay 1 m gravel Till</td>
<td>+4.27</td>
<td>1.35</td>
<td>+0.6</td>
<td>+3.52</td>
<td></td>
</tr>
<tr>
<td>Newton Mains(^{d}) NS 918 835</td>
<td>SRR-1897 8090 ± 280(^{e}) 3560 ± 90(^{f}) Type 6(^{b}) 9.0 m PG silty clay 1 m gravel 10 m LG sandy clay Till</td>
<td>+4.30</td>
<td>1.68</td>
<td>+0.8</td>
<td>+3.42</td>
<td></td>
</tr>
<tr>
<td>East Kerse Mains(^{d}) NS 967 804</td>
<td>SRR-1898 3340 ± 80(^{e}) 3310 ± 100(^{f}) Type 4(^{b}) 4.75 PG silty clay 0.25 m gravel 5.0 m LG sandy clay Till</td>
<td>+3.75</td>
<td>1.83</td>
<td>+0.4</td>
<td>+2.32</td>
<td></td>
</tr>
</tbody>
</table>

*Paul et al., 1992. \(^{b}\)Goselow and Browne, 1986. \(^{c}\)Paul et al. 1995. \(^{d}\)Robinson, 1993. \(^{e}\)Outer age. \(^{f}\)Inner age.

Discussion and conclusions

Post-depositional compaction due to geotechnical compression alone introduces an error in the elevation of a sample point that depends on the thickness of the sequence, the geotechnical properties of the sediment and any history of post-depositional loading.

We find that for the simple case of self-weight compression, the mid-depth correction is around 5–10% of the layer thickness for compressible sediments and around 1–2% for incompressible ones. In a geological context this suggests that the effect usually will be more important for sites closer to the local depocentre, where thicker sequences of finer sediment are likely to occur, rather than at basin margins where coarser, thinner sequences are likely to predominate. The results from the Forth valley indicate that here the required correction may be around 2–5 m at sites around a local depocentre, but is probably less than 1 m for sites closer to the former shoreline.

The significance of the correction clearly depends on the use that will be made of the data, together with the relative magnitude of other likely errors. Thus it may be of greater significance in estimates of late Flandrian sea-levels than in those of the Late Devensian or early Flandrian. Also, the effect on index-point elevation will be more significant in situations where the tidal range is small and/or the sample position is well constrained within the tidal cycle.

There is an obvious importance for estimates of sedimentation rate that are based on dated profile thicknesses. In the case of the Claret Formation at Bothkennar, incorporation of the correction changes the rate between the base and mid-depth of the profile from 9.8 mm yr\(^{-1}\) to 11.3 mm yr\(^{-1}\) (OxA-3389 to OxA-3388: cf. Table 1 and Fig. 3) and between the mid-depth and the top of the profile from 6.4 mm yr\(^{-1}\) to 7.5 mm yr\(^{-1}\) (OxA-3388 to OxA-3507). Thus the corrected values reveal a slowing of sedimentation at this location towards the top of the succession as the sequence becomes emergent.

The correction can also be important in the estimation of palaeotidal ranges and patterns. Hinton (1995, 1996) has shown that, in coastal areas, changes in water depth of less than 10 m may be of significance in determining tidal behaviour and has commented (pers. comm., 1997) that in many simulations water depths are less well constrained than the geometry of former coastlines. Thus a compaction correction may be particularly significant for palaeotidal models of embayments which are rapidly sitting up (such as the upper Forth and the Wash) and where palaeobathymetry is based on removal of isopachytes of these later deposits.

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References


Investigation of the fabric of engineering soil using high resolution X-ray densimetry

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INTRODUCTION
Natural sediments are complex materials whose fabric usually influences their engineering behaviour. High resolution X-ray densimetry (Been, 1981) allows the non-destructive investigation of this fabric which is of particular value for cores or test specimens that cannot be dissected for fabric studies. Although originally developed for the study of early consolidation in very soft clays, the method has proven valuable in the examination of deep ocean sediments (Edge & Sills, 1989), in the geoaoustic study of continental shelf and slope sediments (Paul & Talbot, 1991) and in fabric studies at the Bothkennar research site (Paul, Peacock & Wood, 1992).

In this paper two case studies are presented which illustrate this work and which together have involved the examination of a total of almost 100m of core samples. The examples are drawn from the coastal and offshore areas of Scotland (Figure 1). The sediments involved are, in the first example, Late Devensian sandy silts and muds which have accumulated by various slope processes on the continental margin west of the Shetland Islands and which have allowed a comparison of the density signatures between gross facies types and, in the second example, estuarine mudflats of Holocene age at the Bothkennar research site near the head of the Firth of Forth in which densimetry has aided the recognition and very detailed description of the engineering facies.

Although drawn from very different environments, these sediments are all fine grained, soft and normally to lightly overconsolidated. In addition, they share a number of characteristics which are suited to the X-ray method. They generally lack non-horizontal fissures and other inclined structures so that the horizontal X-ray beam integrates densities across fairly uniform planes and resolves contrasts well. In all cases but one they have preserved a detailed record of primary bedding and associated boundary structures such as ripples, laminae and inter-penetration of beds which have a clear genetic significance: in some places they are also affected to a greater or lesser extent by bioturbation which has modified or destroyed the visible primary fabric. Thus they each possess features which may be expected to give distinctive density signatures and are excellent candidates for study by this technique.

X-RAYS IN CORE ANALYSIS
When X-rays are passed through soil, the beam is attenuated by the different soil components and different densities within the soil. This attenuation can be used to examine an undisturbed core or sample both qualitatively and quantitatively.

The simplest application of X-ray technology is in the production of X-radiographs. In a typical system, the complete or longitudinally sectioned core, or a thin slab, is laid on or close
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Figure 1
Locations of sites

Figure 2
Schematic view of X-ray densimeter

Support for vertical movement
Core under examination
X-ray tube
X-ray detector assembly
Calibration samples
Spacing between collimated slits in front of X-ray tube and detector is 280 mm
to a length of film and exposed to X-rays. The higher the transmission of X-rays, the darker will be the film or negative. With this technique, it is possible to observe inclusions such as individual clasts or large shells, and features such as cracks. The use of a comparatively thin section of the core allows better resolution of lateral non-uniformities and a more precise identification of the thickness of non-horizontal layers, but the core itself is damaged by being sectioned. This technique provides a qualitative measure of soil variability in the core. Attempts to convert the film contrast to a quantitative measure of density have failed due to the difficulty of controlling the development process. Other problems with precise interpretation of the film arise as a result of parallax, which depends on the focal length of the X-ray tube and the distances between X-ray source, core and film.

For the radiographic results reported in this paper, the cores were sectioned longitudinally to a thickness of about 17 mm. The X-ray was run at 55 kV and the film negatives have been printed as positives, so that the lighter regions correspond to lower densities.

An alternative method of monitoring the X-ray attenuation uses a detector crystal such as sodium iodide, coupled with a photomultiplier assembly (Been & Sills, 1981). Figure 2 shows a photograph of the system. A finely collimated beam of X-rays at 160 kV is traversed down the core and the output from the photomultiplier is logged. Some variation in count rate occurs due to fluctuations in the intensity of output of the X-rays, so that a suitable integration time for the counting must be set, along with a suitable speed of traverse of the X-ray head. The values chosen for general use leads to a spatial resolution of the order of ±1 mm. The count rate N obtained by such a system can be related directly to density ρ through the relationship $N = N_0 e^{-\rho d}$, where $N_0$ and $d$ are constants whose values depend on the energy and intensity levels of the X-rays and on the sediment mineralogy. Thus a conversion of count rate to density requires two independent calibration results. These are supplied by calibration samples of known density made up in the same core liner material that is used for the core being analysed.

The accuracy of conversion from count rate to density depends on the uniformity of the core liner as well as on the mineralogical similarity between the calibration samples and the sediment in the core. A realistic accuracy for the density measurement of the cores to be described in this paper is of the order of ±0.01 Mg m$^{-3}$ although the potential accuracy of the system as applied to long term settling column experiments in the laboratory is of the order of ±0.002 Mg m$^{-3}$. The technique is thus well suited to the investigation of a variety of sedimentary structures which involve variations in the packing of the soil framework or gross variations in local density, such as inclusions, primary bedding, grading variations and fabric disturbance (whether natural or as a result of sampling).

CASE STUDIES

Slope deposits on the West Shetland slope

This case study illustrates the identification of gross fabric differences within and between core samples. Plio-Quaternary sediments from the West Shetland slope consist of sediments of both glacial and marine origin; their relative proportions depend on the location and timing of deposition. In this area a stratigraphy has been defined based on seismic sequences, i.e. on the basis of acoustic signature and stratigraphic position (Stoker, Harland, Morton & Graham, 1989; Stoker, Harland & Graham, 1991). X-ray densimetry scans have been carried out on selected shallow vibrocores. Particular attention has been paid to the distinction between acoustically transparent mass flow units and laterally equivalent acoustically layered units since they possess differing density signatures which reflect differences in the depositional processes.
BGS vibrocore 60-06/37, from the acoustically structureless Morrison Sequence, is lithologically uniform in appearance. Figure 3 shows the section from 4.20m to 4.70m below seabed. There is little or no fabric variation with depth (Fig 3a), a uniformity also reflected in geotechnical properties such as water content (Paul, Talbot & Stoker, 1993). On the corresponding density trace (Figure 3(b)) only the larger dropstones and a crack (an artefact) at 4.53m depth are distinguishable from an otherwise uniform background due to the sandy mud. Many dropstones visible in the core section appear too small to show up in the trace, since the sediment itself has a relatively high, almost constant, density of 2.1 Mg m⁻³ and they contribute little additional mass to the overall cross section of the core. This uniform featureless trace is characteristic of nearly all of this core.

BGS vibrocore 60-05/51 is from the acoustically well-layered Faeroe-Shetland Channel Sequence. Figure 4(a) shows the visible fabric features from 3.90m to 4.80m depth. These include: dropstones (e.g. at 4.40m depth); burrows infilled by silt or fine sand and marked by colour changes in the sediment; other general bioturbation is delimited by a fine line on the diagram. Figure 4(b) shows the corresponding density trace from this section of core; the features annotated are described below. The general appearance of the signature reflects the lithofacies changes and the complexity of the fabric; we observe that since many fabric features do not occupy the whole cross section the magnitude of the relative density changes is subdued compared with those described from the other sites.

On the trace, features (i) and (ii) show the position of a crack [artefact] at 3.98m and a large dropstone at 4.40m depth. The area of low density (iiii) at the top of the section is produced by a hole (which could also be an artefact). Between 4.00m and 4.30m the density trace is very uniform (iv), a signature typical of a bioturbated area of sediment (see Bothkennar case study), at the base of which there is a rapid fall in density which corresponds to the change in lithofacies from sandy silt to sandy mud. At 4.30m depth the density increases noticeably over a few centimetres (v), although there is no major change of lithofacies, this change correlates with the sandy infilling of a major burrow shown in Figure 4(a). Below the dropstone (ii) the density increases steadily with depth, notably at (vi) corresponding to the bioturbation at 4.60m and at 4.83m (vii) corresponding to the coarse burrow at the base of the section.

In these cores the density signature varies with the seismic sequence from which the sediment was retrieved due to the particular fabric features in these sequences. The uniform, featureless, high density signature which appears characteristic of the Morrison Sequence reflects the accumulation of this sediment by uniform, mass flow processes, whereas the more complex, lower density signature obtained from all vibrocores throughout the Faeroe Shetland Channel Sequence appears to result from the bottom current re-working that has affected the area. Although, in both examples, every detail in the signature can be related to an individual fabric feature, it is considered advisable in the first instance to establish the nature of this relationship by the direct examination of some representative cores or core sections.

The Bothkennar estuarine clay deposits
By contrast with the study above, this case study illustrates the use of the densimeter to investigate very fine details of the soil fabric at scales down to a few millimetres. It is based on the suite of boreholes sunk at the EPSRC Bothkennar research site as part of a major study of the engineering geology and depositional history of the soft clay sequence at the site. The cores from this study have been scanned by the densimeter, thin slabs have been X-ray photographed and split cores have been described in detail after cleaning with an "osmotic" knife (a thin bladed knife connected to a low-voltage DC electric supply which prevents
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smearing by re-orientation of the clay particles in the electric field). These procedures have enabled a very detailed comparison to be made of the soil fabric and the corresponding density profile.

The EPSRC site is located on former estuarine tidal flats of Holocene age near the head of the Firth of Forth (Paul, Peacock & Wood, 1992). The uppermost 20m of the geological sequence consists of silty clays which are divisible principally into bedded, laminated and mottled facies: the first is composed of primary sedimentary units, separated by bedding surfaces, in the second the units are eroded and separated by numerous silt laminae; in the last the sediment has undergone post-depositional reworking by organisms which has partially or totally obliterated the primary structure. There is often repetitive interbedding of the bedded and mottled facies at a scale of 10-100mm. The three facies have differing engineering properties (Hight, Bond & Legge, 1992; Little, Muir Wood, Paul & Bouazza, 1992) that are attributed to differences in their fabric: these differences can be clearly seen on high resolution density traces from all the boreholes so far examined.

Figures 5 to 7 show density profiles from the different facies presented together with sediment logs and X-ray photographs. Figure 5 (4.87-5.17m depth in HW7) shows a typical section which includes both bedded and mottled facies. Radiocarbon dating (Paul, Peacock & Wood, 1992) in conjunction with the sea level curve for the Forth valley (Sissons & Brooks, 1992) suggests that the sediments were probably deposited in an inter-tidal to subtidal environment in water of only a few metres depth at most. The X-ray photograph and sediment log show that the lower part of this section (A in Fig. 5) is constructed from individual beds of silty clay each about 5-20mm in thickness, sometimes separated by silt partings and laminae (B). At the top of the section (C) the sediment is bioturbated and the primary sedimentary bedding has been lost. The bioturbation is marked by a visible pattern of motting, which shows a transition from a large, sparse style to a smaller, denser style over a distance of about 50mm. In the lowermost, bedded section. X-ray attenuation measurements reveal the presence of primary depositional bedding and can resolve density variations within individual beds. It should be noted that the radiographs were exposed and developed in conditions aimed at optimising the show of detail, so that on the photographs the apparent contrast does not necessarily represent equivalent density changes. There is, nevertheless, a broad correlation between the positions of the dark bands on the photograph and the higher density parts of the density profile. A detailed comparison of the density profile with the X-radiograph shows that individual bed contacts are visible (i), as are silt laminae (ii) and that the saw tooth pattern (iii, iv) arises from the upward gradation from higher density clayey-silt to lower density clay within each sedimentary unit. After the transition from bedded to mottled facies (v) the density profile is very uniform, due to mixing by bioturbation as indicated by the dense pattern of small motting on the cut surface. The onset of mixing occurs over a short distance (vi) which corresponds to the change in the pattern of motting noted above.

Figure 6 (3.65-3.95m depth in HW5) shows a section in sediments of the laminated facies, which were deposited in a subtidal channel in water of a few metres depth. The core shows a regular, repetitive sequence of fine silt laminae, separated by clay beds often of no more than 5mm thickness. For the most part these beds are horizontal (or nearly so) from one side of the core to the other, with undulating eroded tops and graded bases: there are occasional cross cutting horizons (e.g., at A in Fig. 6) and water escape structures (B). The X-radiograph reveals a wealth of such fabric information down to the limit of its resolution. The density profile is very different from that in Figure 5 and reflects the laminated structure. Again, there is broad correlation between the higher densities and the darker features apparent on the radiograph, although not all the thin, high contrast features correspond to an equally thin layer.
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on the density profile. This may suggest that the layers in the complete core are slightly inclined, while the section used for the radiograph happens to have been taken across the sense of the inclination. Two aspects are noticeable in particular: the magnitudes of the peaks are generally larger than in the bedded facies, and the majority of peaks are symmetrical. The former effect is clearly due to the density of packing in the silt laminae and the latter is due to their internal uniformity; only occasionally do gradations in density produce any discernable saw-tooth pattern.

Figure 7 (2.17-2.47m depth in HW7) shows a section of core from the sub-crust transition (Paul & Barras, 1993). This is composed of sediments belonging to the bedded facies which were deposited under inter-tidal conditions, and which often show evidence of fabric disruption by large burrowing organisms; they also show evidence such as silt clasts, mud balls and flakes which indicate energetic conditions with frequent episodes of minor erosion. Several such features are visible in Figure 7: e.g. a major burrow (A); disturbed and broken laminae (B); water escape (C). Overall, the effect of this disturbance is to introduce small scale lateral variation which reduces the average density contrasts within the section and so smooths the density trace in a manner analogous to, although less complete than, the bioturbation associated with the mottled facies. Comparison of the trace in Figure 7 with those of Figures 5 (lower part) and 6 illustrates this point.

DISCUSSION:

Comparison with conventional methods

Figure 8 shows a quantitative comparison of the value of the density measured by the densimeter with that obtained on this core by a conventional cutting ring at 50mm intervals (dotted line). The data is taken from an earlier study of sediments from the central North Sea (Paul & Jobson, 1991). Three points can be made: (1) the overall shape of the conventional profile matches that of the densimeter trace at the 50-100mm scale, although with considerable loss of detail at finer scales, (2) the trace largely lies within the [two standard deviation] error bound of the conventional method (calculated from previous work on this core (Paul & Jobson, 1987) to be ±0.08 Mg/m³) and thus the values are quantitatively comparable, although this is obviously dependent on the accuracy of the calibration, (3) there is a systematic difference between the curves over the lower two thirds of the core. This suggests either a small calibration discrepancy over the whole core, a significant change in sediment mineralogy one third of the way down, or a less than precisely vertical alignment of the core. For the purpose of fabric investigation, these last differences are not considered to be significant. The conventional measurements could in any event be used to recalibrate the trace (either overall or locally) if so required. The overall conclusion is that the densimeter gives results which are directly comparable to conventional methods, is capable of better resolution and, with accurate calibration, of equal or better accuracy.

Resolution of fabric features

These case studies show that the method is able to resolve many small details of the sedimentary fabric. As already noted, this is best shown when the structures involved are sub-horizontal and laterally continuous across the core. Such features are the norm in a bedded, tidal mud such as the Bothkennar clay, which would therefore be expected to produce rich densimeter signatures. Features that are laterally impersistent, such as lenses, clasts, individual shell fragments etc, usually give less well defined features on the density profiles, although they may show up on the radiographs.
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The profiles show that under good conditions a vertical resolution down to 1-2mm appears possible and also that a density contrast between two adjacent layers of a little as 0.01 Mg m\(^{-3}\) detectable within the signatures. Calculation shows that at typical soft sediment water contents this corresponds to a water content difference between the layers of 1% or less. In these sediments, variations of this magnitude appear common over short (millimetre) distances and this may well be the case more generally, although the Authors are unaware of other detailed studies at this fine scale which might confirm this idea.

*Application to site investigations and the study of geological processes*

The densimeter method has proven very valuable for the examination of whole cores. As described earlier, it has enabled the facies succession to be constructed and the location of particular fabric features to be identified prior to the extrusion and sectioning of the sample. This has allowed the sampling pattern to be decided at an early stage and problematic or difficult fabrics to be identified. When combined with X-radiography a very detailed picture of the soil fabric has been obtained which aids in the explanation of the geotechnical behaviour of individual test specimens.

In addition to its applications to engineering investigations, the densimeter gives novel information about geological processes at the small scale, e.g. density variations within primary depositional bedding, the correlation between the degree of surface mottling and the extent of internal mixing by bioturbation; quantitative data on layer statistics and spatial spectra of laminae and other planar structures. Although it is beyond the scope of this paper to discuss these aspects in detail, the Authors note that these possibilities exist and hope to consider them in later publications.

**CONCLUSIONS**

These examples have shown that the detailed pattern of density variation within a sediment core can be established quantitatively at the millimetre scale by X-ray densimetry and that the results compare well with those from conventional methods which are commonly used for classification. In the vibrocores from the West Shetland slope, the distinction between gross seismic facies can be explained by the density profiles that arise from visible differences in their respective sedimentology. At Bothkennar, the fine details of the bedded, mottled and laminated facies can be identified in the density profiles and related closely to visible and radiographic features. Thus in each of these examples X-ray densimetry has provided additional information to assist the description and evaluation of geological samples and other site investigation data.

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FLANDRIAN STRATIGRAPHY AND SEDIMENTATION IN THE BOTHKENNAR-GRANGEMOUTH AREA, SCOTLAND

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Introduction

The low lying topography in the Bothkennar-Grangemouth area of central Scotland is developed on the carse clay, a local name for the Flandrian silty clays which occur widely in the middle and upper Forth valley. Although their geomorphology at outcrop has been extensively investigated previously (Sissons et al., 1966; Smith, 1968), only generalised descriptions have been published of the sediments themselves, to which a number of informal names have been applied (Francis et al., 1970; Laxton and Ross, 1983; Sissons, 1966; 1976; Browne et al., 1984). Recent engineering geological investigations in connection with the newly established Engineering and Physical Science Research Council (EPSRC) soft clay research site at Bothkennar [NGR NS 921 859], near Grangemouth, have enabled us to provide a detailed account of the sedimentology, age relationships and palaeoenvironment of the carse deposits in this area (Paul et al., 1992). Our results suggest that in this area the greatest thickness of the sediment was deposited during the post 6,500 BP Flandrian regression, in contrast to the situation farther west where deposition largely occurred during the preceding transgression (Sissons and Brooks, 1971). Thus the depocentre in the middle Forth estuary may have migrated first onshore and later offshore during the Flandrian transgression and regression, a model which may be of more general application to other east coast Scottish estuaries. In view of this migration we have also found it useful to define a formal lithostratigraphy for parts of the estuarine fill in order to avoid possible confusion arising from use of the above informal names in any chronostratigraphic sense (e.g. Paterson et al., 1981).

Deposits at the Bothkennar Soft Clay Research Site

The EPSRC Bothkennar research site is located on the west bank of the Forth estuary about 1 km south of the Kincardine bridge (Figure 1). This part of the foreshore was reclaimed around the year 1790 (Cadell, 1929) and the site lies within the reclaimed area. Figure 2 shows the lithological profile from borehole HW3, one of the 16 site investigation boreholes sunk during the engineering geological study of the site (Paul et al. 1992; Hawkins et al. 1989). This profile is typical of most of the site, other than the southeastern part (below). The bore terminates at an horizon of gravelly sand, which we correlate with the Bothkennar Gravel of Browne et al. (1984) on the basis of elevation.

Figure 1. Geological sketch map to show the Flandrian deposits in the middle Forth valley and the localities discussed in the text (based on Gustelow and Browne, 1981).
and lithology, and which is in turn believed to correlate with the widespread 'buried gravel' in the upper Forth valley (Sissons, 1969) of probable Late Devensian age. This gravel is succeeded by micaceous silty clays which extend to about 1.5-2 m below ground surface, which we correlate with the Claret Beds of Browne et al. (1984). In the southeastern part of the site the uppermost 7-8 m of this silty clay is replaced by a finely laminated clayey silt; evidence from boreholes and cone penetrometer tests (Hawkins et al. 1989) suggests that these deposits occupy a channel whose flank impinges on the EPSRC site and whose lithology suggests a tidal origin. We correlate these laminated deposits with the Grangemouth Beds of Browne et al. (1984) on the basis of their lithological similarity and cross-cutting relationship with the underlying silty clays.

It is possible to recognise two principal facies within these silty clay sediments at Bothkennar (Paul et al., 1992). The bedded facies is a silty clay in which the primary sedimentary layering remains visible and a variety of sedimentary structures may be seen, whereas in the mottled facies this bedding has been partially or totally destroyed by bioturbation. The two facies often occur in thin subunits, interbedded at the 10-100 mm scale, whose relative frequency varies with depth. At the base of the formation the bedded facies predominates, whereas the mottled facies increases steadily in relative proportion towards the top of the unit. In its uppermost few metres the sediment contains numerous large (1-10 cm) burrows which in general post-date any smaller scale motting.

The silty clay is terminated upwards by an unconformity which is often marked by a shell bed, containing principally Cerastoderma edule (paired valves and juveniles) together with Retusa obnusa, Hydrobia ulvae, Mytilus edulis and Scrobicularia plana. The sediments above the unconformity are composed of crudely stratified grey to brown clayey silts with lenticules of shells (fragments and disarticulated valves, chiefly Cerastoderma edule). They grade upwards into the artificially induced intertidal deposits known from the literature to have accumulated very rapidly following the late 18th century reclamation work. These deposits have weathered to form a surface crust which extends to a depth of approximately one metre.

Stratigraphic Proposals

We consider that there is sufficient consistency in the character of the Bothkennar Gravel, Claret Beds and Grangemouth Beds in the Bothkennar - Grangemouth area to justify their inclusion in a formal lithostratigraphic framework for the area. Table 1 shows this framework and describes the lithology and our environmental interpretation of the sediments. It is applied to the Bothkennar profile in Figure 2. The Flandrian sediments are assigned to two formations:
the Claret Formation, which comprises the partially bioturbated silty clays, and the Grangemouth Formation, which comprises the mixed suite of tidal channel, intertidal and modern reclamation deposits which lie unconformably on the Claret Formation. Below the Claret Formation are the sandy gravels of the Bothkennar Gravel Formation.

The Grangemouth Formation is subdivided into the formally named Grangemouth Docks, Skinflats and Saltgreen Members (Table 1) whose sediments can be recognised widely in the area (Brown et al., 1984; Robinson, 1993). The Grangemouth Docks Member is composed of finely laminated silts and clays of probable subtidal origin, which occupy a system of channels cut into the Claret Formation. The Skinflats Member is composed of crudely stratified silts and clays which contain the lenticles of shell debris described above and is marked at its base by the shell bed (cf Robinson 1993 for occurrences elsewhere in the area). The member is believed to represent a beach veneer developed, possibly as patches, on the intertidal surface of the Claret Formation. It may be confined to the lower geomorphic surfaces. The Saltgreen Member is composed of massive or crudely stratified silts and clays, usually weathered, which form the land surface in the areas of historical foreshore reclamation.

It is also possible, at Bothkennar, to recognise four informal lithological divisions of the Claret Formation (Hawkins et al., 1989; Hawkins et al., 1991; Paul et al., 1992) although it is not yet clear whether or not these divisions can be recognised beyond the EPSRC site itself. These divisions (Table 1) are believed to record a phase of marine transgression (Basal division) and regression (Lower to Upper divisions) in the Bothkennar area (Figure 2 and below).

The two principal sedimentary facies within the Claret Formation are central to this classification. As shown in Figure 2, the relative frequencies of the bedded and mottled facies vary considerably with depth; these relative frequencies appear to be laterally consistent across the site and so provide a basis for the division of the formation. In the Basal division the sediments of the Claret Formation are entirely bedded and commonly contain laminae of silty sand. They are not mottled and in this respect they differ from units of the bedded facies in the overlying sequence. By contrast, the overlying silty clays that comprise the bulk of the Claret sequence are characterised by both the bedded and mottled facies. The lower part of this succession, in which the bedded facies dominates the profile, is termed the Lower division. Although some sediments in this division may show mottling, it is usually insufficient to obliterate the primary bedding, which remains visible to visual inspection.
The higher part of the sequence comprises the Middle division in which the mottled facies is dominant. The motting can be intense, covering up to 100% of any vertical section, and shows a variety of styles and sizes. There is a common cyclic pattern in which large, low density motting is replaced upwards by successively finer, denser motting until an abrupt stop is reached at a bedding horizon. An entire cycle may occupy an interval of 500mm or more when fully developed, although there are also many examples of truncated or partial cycles. This situation has been observed frequently in all the boreholes at the EPSRC site (Figure 1: cf Hawkins et al., 1991; Paul et al., 1992). Also, in one borehole (HW3: Figure 2), a minor (1.5m thick) unit of finely laminated silt and clay has been recorded, which may be a local subtidal channel within the Middle division.

Above the Middle division there is a zone where the bedded facies again dominates the sequence and is associated with a coarsening of the grain size and the presence of silt laminae and other erosional contacts between the individual beds. We term this zone the Upper division. In this zone motting is relatively both infrequent and coarse; and detailed examination reveals the presence of numerous macroscopic burrows, often truncated by bedding surfaces which also show primary sedimentary structures such as mud clasts (up to 2-3mm), flakes and rip-ups. These latter features suggest an energetic environment perhaps associated with intermittent subaerial exposure; they are in accordance with our suggestion (below) that the upper division formed under inter-tidal conditions. The lower limit of this division is defined as the level above which macroscopic burrowing becomes evident; it is best developed in those boreholes from the centre and north of the EPSRC site.

Radiocarbon Ages and Palaeoenvironmental Changes in the Grangemouth Area

Several radiocarbon dates have been obtained from the Flandrian sediments around Grangemouth (Brown et al., 1984; Paul et al., 1992; Robinson 1993). Those of known stratigraphic position are summarised in Table 2. The dates suggest that at Bothkennar the sediments of the Basal Division of the Claret Formation were deposited prior to c.5,000 BP and the remaining sediments of the Claret Formation, which comprise the bulk of the succession, between about 5,000 and some time prior to 3,000 BP. This chronology shows that at Bothkennar deposition occurred mainly after the peak of the main Flandrian transgression (c.6500 BP) and thus implies a general reduction in water depth during the depositional period, corresponding to a palaeogeographic change from an offshore (sublittoral) to coastal (tidal flat) position as sea level fell. The local bases of the Skinflats and Grangemouth Docks Members of Grangemouth

<table>
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<tr>
<th>Sample Code</th>
<th>Lab. Ref.</th>
<th>Elevation (m OD)</th>
<th>Stratigraphic Location</th>
<th>Adjusted Age</th>
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<th>Lab. Ref.</th>
<th>Elevation (m OD)</th>
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<td>407±150</td>
<td>Kinnel Kdn</td>
</tr>
</tbody>
</table>

Notes
1. Adjusted ages based on uncorrected age at 40.0±600 years for further (Harrass 1985) and (Harrass 1990). There is other stratigraphic expression with this borehole. (Francis & Henderson, 1990).
Formation, where they lie unconformably on the Claret Formation, are both dated at or later than 3,000 BP and so the channelling appears roughly contemporaneous with the initial development of the Cerastoderma bed on the adjacent tidal to subtidal flats. The mid sequence date of approximately 3,800 BP suggests that the average rate of sedimentation decreased from around 13mm/yr to around 8mm/yr during the period of deposition of the Claret Formation, which is consistent with the predominance of the bioturbated mottled facies in the upper part of the sequence.

An approximate sea level curve for Bothkennar (Figure 15 in Paul et al., 1992) has been derived from that of Sissons & Brooks (1971) by adjusting their data from the Stirling area for the isobase fall to Bothkennar and then projecting the expected decay due to isostatic rebound down to present day sea level. The result agrees broadly with the curve determined independently by Robinson (1993) from other sites in the middle and upper Forth valley. The Stirling data is based on peat above the carse clay; if its level is taken to be approximately one metre below high water spring tide (cf. Sissons and Smith 1965), the tidal level can be reconstructed approximately by assuming the range to be similar to that of the present day.

The approximate water depth at any time during deposition can be determined from the elevation difference between the radiocarbon sample and the sea level curve at that date (Figure 2). There is some uncertainty due to post-depositional consolidation of the soft silty clay; a corrected water depth curve (Figure 2) has been calculated using geotechnical data from Nash et al. (1992). The results indicate that the water depth at Bothkennar decreased continuously during the deposition of the Lower to Upper divisions of the Claret formation, falling from around 20m below low water at the base of the formor to intertidal at the time of the Cerastoderma bed. The water depth at the time of the basal division is not known directly; however, the sedimentological data suggest that it may have been relatively shallow; faunal data (Paul et al., 1992) suggest a depth of at least 5-10m.

The sedimentology of the deposit reflects these changes. Figure 2 shows a comparison of the median grain size with the facies architecture and water depth model. In the lower part of the sequence the median size reduces from the Basal to the Lower divisions, probably as a result of a decrease in current activity. The Lower division has a relatively low proportion of the mottled facies, suggesting a period of sustained, possibly rapid, deposition. In the Middle division, as the water depth continued to reduce, there is an increased proportion of motting, indicating bioturbation during periods of reduced sedimentation. In general, in the Middle division the median grain size of the sequence increases upwards, suggesting a classic upward coarsening sequence indicative of falling relative sea level. The model suggests that the uppermost part of the sequence (above about -2m OD) was intertidal, which agrees with the sedimentary features described above from the Upper division. The sequence is completed by the Cerastoderma bed, which seems to have formed at around 2-3m above the low water spring tide level. This accords with the present day ecology of Cerastoderma edule (chiefly lower intertidal), and also with the observations of Sissons and Smith (1965) that the modern tidal flats accumulate to within about 1-2m of the spring tide high water level.

Patterns of Postglacial Sedimentation in the Forth Valley

Our interpretation of the local pattern around Grangemouth may be compared with wider evidence from the middle and upper Forth to suggest a model for Flandrian sedimentation in the valley as a whole. To the west, around Letham (Figure 1) and to the northwest, at the head of the estuary around and west of Stirling, carse clays up to about 10m thick overlie peats which have developed on a sequence of buried raised beaches (Sissons and Smith 1965; Sissons 1966, 1969; Francis et al., 1970; Sissons and Brooks, 1971; Laxton and Ross, 1983; Browne et al., 1984). These carse clays were deposited between the onset of the Flandrian transgression (c.8,500 at Stirling to c.7,500 west of Menteith) and the subsequent regression; it is known that deposition at the higher levels ended prior to about 6,500 BP with the establishment or re-establishment of peat growth (Sissons and Brooks, 1971; Browne, 1987). Thus at the margins of the basin deposition occurred chiefly during the transgressive phase.

To the west of Grangemouth the sediments of the Letham beds (Browne et al., 1984) lie in a apparently similar stratigraphic position to the basal Claret Formation. However, they differ in lithology (grey silt and sand) and are locally separated from the carse deposits (which include the Claret Formation) by peat (Browne et al., 1984). Upstream, they also differ in age from the basal Claret Formation, being associated with the buried beaches of the Forth valley (>8,500 BP) and with a cool temperate fauna (Browne, 1987). Thus the deposits which immediately overlie the Bothkennar Gravel Formation across this wider area are apparently diachronous. At the local scale, the results from the Grangemouth area show the deposition of the Claret Formation also to have been diachronous, since deposition of the Grangemouth Formation began about or later than 4000 BP at Grangemouth itself (Browne et al., 1984; Robinson, 1993) and about or later than 3000 BP at Bothkennar and its environs (Table 2 and Robinson, 1993). Thus a range of published data from sites to both the west and northwest of the Grangemouth area show that, as in the boreholes considered here, both the onset and cessation of sedimentation
sedimentation occurred primarily during the main Flandrian regression, whereas, in the Stirling and Letham areas (i.e. close to the basin margins), sedimentation occurred primarily during the preceding transgression.

Such a pattern can be explained by the proposal that during the transgression the depositional locus (depocentre) migrated shorewards and retreated seawards during the following regression. Based on this proposal, Figure 3 shows a conjectural litho- and chronostratigraphy for the Flandrian sediments along a west to east line in the middle Forth valley which would result from our model; an analogous pattern of younging would apply from northwest to southeast. The Figure emphasises our argument that, away from the margins of the basin, the succession records the progressive infilling of the Forth valley as the sea withdrew after the Flandrian maximum and hence that the Claret Formation is a composite diachronous unit in which differing palaeoenvironments of differing ages are recorded at varying positions. We believe that this model provides an explanation of the available sedimentological and chronological data over this wider area, although we readily accept that our model is provisional and may be subject to revision in the light of further research.

Conclusions

This work has shown that the Flandrian sediments in the Grangemouth area are a mixed suite of bedded, bioturbated and laminated silty clays of nearshore to tidal flat origin, which date from c.5000 to c.3000 yrs BP. They record deposition during a period of falling sea level during which the locus of sedimentation migrated across the area from landwards to seawards. We suggest that there is sufficient consistency in this infill to establish a lithostratigraphic sequence, but do not consider that this provides a chronostratigraphic framework due to the diachronous nature of the deposits.

We conjecture that the Flandrian transgression in the middle and upper Forth valley led to sedimentation chiefly on the flanks of the estuary, whereas the regression from the highest Flandrian sea level was accompanied by migration of the depocentre offshore. On a broader scale, similar migration of depocentres is likely to have occurred during the deposition of Flandrian estuarine sediments elsewhere in Scotland, for instance in the Tay valley and in the Moray Firth. Moreover, it can be further speculated that depocentre migration consequent on rising or falling sea level was a major factor controlling the deposition of Scottish Late-glacial as well as Flandrian sediments, particularly those laid down away from proximal glaciomarine environments.
Acknowledgements

Field and laboratory assistance at all stages of the work were provided by Mr H. Barras to whom we extend our very sincere thanks. We thank Mr J.J.M. Powell of BRE for advice on sampling procedures, the staff of BRE for the collection of cores from the HW series of boreholes and Mr D.F.T. Nash for access to the Delf cores held at Bristol University. Also we thank the NERC and the Edinburgh office of the British Geological Survey for access to unpublished data from the Grangemouth area and thank Mr M.A.E. Browne of BGS for helpful discussions. The financial support of the SERC (grant nos: GR/E7/47448 and GR/H/14151) is very gratefully acknowledged.

References


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The engineering geology of the Carse clay at the National Soft Clay Research Site, Bothkennar

M. A. Paul,* J. D. Peacock* and B. F. Wood*

It is possible to subdivide the Carse clay deposits above the Bothkennar gravel into bedded, laminated, mottled and weathered facies. The first is composed of silty clays which retain original sedimentary structures from quiet water tidal deposition. The second records locally more energetic conditions which gave rise to the deposition of interbedded silt laminae. Reduced rates of sedimentation at times allowed burrowing organisms to rework the sediment and hence to produce the mottled facies. The last facies is a near-surface zone which has been affected by dessication, oxidation and modern pedogenesis. Most of the succession above the Bothkennar gravel can be correlated with the Claret beds which occur widely in the area. The lowest part of the sequence has some similarity to the Letham beds, but the exact correlation is uncertain. The Claret beds are succeeded by modern tidal flat deposits, which rest on them with marked disconformity. Radiocarbon dating indicates their age to 5000–3000 BP. The borehole depths and the sea level curve together suggest that the Claret beds accumulated offshore in water whose depth reduced from about 20 m to 5–10 m, which agrees broadly with faunal data. The grain size distribution shows little variation with depth. The principal clay mineral is illite, but the plasticity appears enhanced by organic residues. The principal ions in the soil water are sodium, chloride and sulphate: their concentration appears similar to that of the adjacent estuary water. The soil has an electrical resistivity of 1–4 Ωm, controlled by the water content and pore water solutes. The P wave velocity is isotropic and equivalent to that in water.

Il est possible de sub-diviser les couches d’argile du Carse au dessus de Gravier de Bothkennar en des faciés stratifiés, feuilletés, laménés, marbrés et altérés. La première se compose d’argiles limoneuses qui ont gardé leur structure sedimentaire originale provenant d’une déposition calme par marées. La deuxième montre localement des conditions plus énergétiques résultant en des dépôts lardes de lamelles de limon. A certains moments des taux réduits de sédimentation ont permis à des organismes fouisseurs de re travailler le sédiment et par conséquent de produire le faciès marbré. Le dernier faciès est une zone près de la surface qui a été affectée par la desiccation, l’oxydation et une pédogenèse moderne. La majorité des couches au dessus du Gravier de Bothkennar peuvent être corréllées avec les bancs de Claret qui sont rencontrés partout dans la région. La partie la plus basse de la séquence a quelques similarités avec les bancs de Letham mais la corrélation exacte est douteuse. Les bancs de Claret sont suivis par des dépôts marécageux modernes formés dans la zone des marées qui reposent sur eux d’une façon non-conforme marquée. Une mesure de l’âge au radiocarbone indique leur âge entre 5000 et 3000 BP. Les profondeurs des sondages et la courbe du niveau de la mer suggèrent que les bancs de Claret se sont accumulés offshore dans de l’eau dont la profondeur s’est réduite de 20 m à 5–10 m ce qui est en gros en accord avec les données de la faune. La distribution de la taille des grains montre peu de variation avec la profondeur. Le minéral argileux principal est l’illite, mais la plasticité apparaît augmentée par les résidus organiques. Les ions principaux dans l’eau du sol sont les ions sodium, chlorure et sulfate: leur concentration apparaît semblable à celle de l’estuaire voisin. Le sol a une résistivité électrique de 1–4 Ωm, contrôlée par la teneur en eau et les matériaux dissous dans l’eau. La vitesse des ondes-P est isotrope et équivalente à celle de l’eau.

KEYWORDS: clays; fabric/structure of soils; geology; microscopy.

INTRODUCTION

The National Soft Clay Research Site is located at Bothkennar, on the west bank of the Forth

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Estuary, and was chosen after an exhaustive search throughout the UK (Hawkins, Larnach, Lloyd & Nash, 1989). The site lies within the outcrop of the Carse Clay which forms the Forth lowlands (Fig. 1) and which comprises the main soft clay sequence at the site. This sediment is of
Postglacial age, and was formed at a time of higher relative sea level in an extension of the estuary which inundated the Forth valley as far as Stirling. At the site itself the Carse Clay is overlain by modern deposits of the reclaimed tidal flats.

The general Late and Postglacial Quaternary stratigraphy of the area is well known (Sissons, 1969; Browne, Graham & Gregory, 1984), and in the Bothkennar area consists of a buried gravel (the Bothkennar gravel), above which lie the argillaceous Letham and Claret Beds which form the soft clay sequence itself. This sequence is in part overlain by the clayey silts of the Grangemouth Beds, and at the margins of the estuary is completed by modern intertidal deposits. The deposits below the Bothkennar Gravel, which are not addressed in this study, consist of further silty clays which overlie till, which in turn lies on bedrock.

This Paper gives an account of the engineering geology of the deposits from the level of the Bothkennar gravel up to the ground surface. The strategy has been to study in detail the geological character of the deposits in order to construct a facies classification of the sedimentary units, and to correlate them with the Late Quaternary succession elsewhere in the Forth Valley. This facies classification provides a framework within which to understand the geotechnical properties of the sediments and their variation over the site.

This classification is based principally on the visible fabric of the sediment, together with mineralogical and microfabric data, and is linked to an outline investigation of certain simple geotechnical properties. It is anticipated that this approach will enable any variation with depth of both these and more complex geotechnical properties to be better understood.

PROGRAMME OF INVESTIGATIONS

In April 1989 a borehole (HW3) was sunk in the southwestern corner of the site (grid reference NS 9206 8585, local co-ordinates 1030 E, 4993 N). This location was chosen because the buried Bothkennar gravel was known to be deepest in this area (Nash & Lloyd, 1988) and hence the soft clay sequence was expected to be best developed. This borehole penetrated the clay to a depth of 199 m and terminated at the top of the Bothkennar gravel. It thus included the full local depth of the clay above the buried gravel. Continuous 900 mm flights of 100 mm dia. piston samples were collected using thin-walled alloy tube. The sample quality was generally good, although some disturbance was noted in the uppermost 100–150 mm of some flights. The sample tubes were sealed with wax on collection, and then stored at constant temperature until required.

Before opening, non-destructive measurements were made of bulk density by continuous X-ray attenuation. On opening, measurements were made to determine P wave velocity. Each 300 mm section was then split and described in detail, based on visual and photographic logging.
after cleaning with an osmotic knife. Measurements were then made of electrical resistivity, vane shear strength and water content. Undisturbed sections of the core were preserved for electron and optical microscopy. The remaining material was subsampled for analysis of particle size, Atterberg limits, mineral composition, organic content and pore fluid chemistry. The particle size was determined after dispersion by the addition of sodium hexametaphosphate and ultrasonic agitation. Atterberg limits were determined on air-dried samples, on samples brought from the natural water content without prior drying, and on samples treated with hydrogen peroxide.

Specialized techniques were used to investigate the subsamples. Mineral identification was made by X-ray diffraction (XRD) and energy dispersive X-ray spectroscopy (EDX), particle size was measured by pipette and with a Micromeritics 5100 SediGraph, and microfabric was recorded by electron microscopy. Palaeontological analyses were also made of some samples. The anion chemistry of the leachable pore fluid residues was analysed by high performance liquid chromatography (HPLC) and the cations by atomic absorption spectrophotometry (AAS).

Scanning electron microscope (SEM) samples were prepared by freeze-drying using the technique of Smart & Tovey (1982). A small core of approximately 15 mm diameter was cut from the waxed subsample, and immediately frozen in liquid nitrogen buffered by liquid propane. It was then placed under vacuum in a temperature-controlled chamber, held at −50°C for about 24 h, and then raised in a controlled manner to −20°C while the vacuum was monitored. When no further deterioration of the vacuum was noted, the sample was returned to room temperature. In this way the production of interstitial ice was hopefully avoided, so preserving delicate structural features.

SEDIMENT DESCRIPTION

Figure 2 shows the geological succession in borehole HW3. Five broad lithological units can be identified (L1–L5) on the basis of sedimentology and macrofabric. These units are believed to have local stratigraphic significance, and can be related to the stratigraphy proposed by Browne et al. (1984) for the Grangemouth area. Within these units, three principal facies can be identified: a bedded facies, in which the primary sedimentary layering remains visible, a mottled facies, in which the bedding has been partially or totally destroyed by bioturbation, and a laminated facies in which numerous silt laminae are present at a spacing of a few centimetres. In addition, a separate weathered facies is recognized at the surface.

Bedded facies

The bedded facies is a silty clay to clayey silt (Fig. 3) in which original sedimentary bedding is visible. Individual beds range in thickness from a few millimetres to about ten centimetres, and are separated by surfaces on which primary sedimentary structures are visible. These structures include ripples, load casts, rip-up mudflakes and mudlump features. This facies may also contain local silt laminae and pockets, the bases of which show local erosional structures and which pass transitionally back into the overlying muds. Such features are relatively infrequent.

At the base of the succession, immediately above the Bothkennar gravel, there is a minor subfacies in which the sediment is locally more poorly sorted. There may be minor interbedding.
Fig. 3. Section of borehole HW3: bedded facies (17.5-17.8 m depth)

Fig. 4. Section of borehole HW3: mottled facies (9.7-10.0 m depth)

Fig. 5. Section of borehole HW3: laminated facies (7.9-8.2 m depth)

Fig. 6. Section of borehole HW3: shell bed in weathered facies (1.6-1.9 m depth)
of the bedded facies within the mottled facies: this is omitted from Fig. 2 for clarity.

**Mottled facies**

The mottled facies is dominantly a silty clay (Fig. 4) in which primary bedding and laminations are poorly defined or absent, although there are local fining upwards structures. The principal feature of the facies is the presence of mottles at scales from a few millimetres to 1–2 cm, which occupy 20–60% of the surface in vertical section. In the unoxidized sediment such mottles are slightly paler than their surroundings, although this distinction disappears on exposure to the atmosphere.

Three types of mottling have been recognized: relatively sparse mottles which are circular to elliptical (up to 10 mm) in vertical section, smaller (1–3 mm) mottles which contain curved and hooked elements, and which cover about 50% of the surface, and fine, threadlike horizontal mottles which are densely packed. These types frequently succeed one another in a regular, cyclic pattern in the sequence already described. Individual mottles can terminate abruptly against local bedding surfaces, and at any point in the facies there may be minor horizons of unmottled sediment, which are usually slightly coarser in texture.

The effect of mottling on various engineering properties has been considered in detail by Hawkins, Lloyd & Nash (1991). They also report that mottles are often associated with small (0.5–2 mm) cavities, which they interpret as rootlet holes. Such cavities were also seen in this study; however, an alternative explanation for the mottling based on bioturbation as given later is preferred.

**Laminated facies**

The laminated facies (Fig. 5) is developed locally at 7.75–9.40 m depth in HW3. It consists of individual beds of silty clay, separated on their bedding surfaces by silt laminae typically 1–2 mm thick, although some may be thicker. Local erosional features are common at their lower boundaries. The laminae are about 30–50 mm apart. Most of the silty clay beds show little bioturbation, although within the facies there are occasional subsidiary beds of finer sediment up to about 50 mm thick which are mottled.

In general, the laminated facies can be viewed as a development of the bedded facies, in which a high proportion of the silty clay beds are separat-

![Fig. 7. X-ray attenuation (bulk density) signatures of sedimentary facies: (a) bedded facies; (b) mottled facies; (c) laminated facies](image-url)
ed by laminae, rather than being in direct contact. At 8.86 m and 9.37 m depth there are thick beds (20 mm and 10 mm) of medium to coarse sand, which may have been deposited following erosion of a channel bank during a storm.

Weathered facies

Within about 4 m of the surface the sediment undergoes a gradual change of colour from the characteristic very dark grey to black through dark grey to greyish brown (Munsell colours). This change is associated with the gradual elimination of primary sedimentary structures, which are replaced by fissuring and other pedogenic features. This altered material is referred to the weathered facies (broadly equivalent to the surface crust, but also including a sub-crust transition zone).

The facies transgresses a major unconformity between the main clay sequence (the Claret Beds) and the overlying clayey silts of the reclaimed tidal flats. In borehole HW3, and elsewhere on the site, this unconformity is marked by a bed of Ceratoderma edule (Fig. 6), many of which are paired and articulated, and apparently in positions of life.

X-ray attenuation signatures

The measurement of bulk density by X-ray attenuation has been described by Been (1981) and has been used by Edge & Sills (1989) to examine density variations in sediment cores. A continuous profile of X-ray attenuation, calibrated for sediment density, was obtained for HW3. Each of the above facies has a distinctive X-ray attenuation signature; all of these have proved valuable in the recognition and delimitation of the facies (Fig. 7). In each case the variation closely follows the visible sedimentary structure.

Figure 7 (a) shows the normal vertical density profile of the bedded facies; the individual peaks correspond to sedimentary layers in the core, and have a saw-tooth structure similar to that described by Edge & Sills (1989). Fig. 7 (b) shows the lack of variation characteristic of the mottled facies, which may be presumed to be the result of bioturbation and which corresponds to the lack of primary sedimentary structures in this facies. Fig. 7 (c) shows the effect of narrow silt laminae in the laminated facies, which produce sharp peaks of increased attenuation, corresponding to locally raised density.

These signatures are useful in two ways: they reveal features of the sedimentary fabric which might otherwise be overlooked, and they indicate the vertical transition of facies, which is not always easily seen from visual inspection. The signature of the weathered facies cannot be generalised in the same way: it shows a broadly uniform signature, disrupted in places by significant fabric features such as fissures and laminae.

Particle size and composition

The particle size distribution has been determined at about 130 positions down the core, at a spacing of about 150 mm. Fig. 8 shows the profile of the major size fractions. Overall, the mean grain size increases from bottom to top in the profile. The clay sized fraction usually comprises 35–50% of the material: the proportion of sub 1 μm particles is greater in the mottled than in the bedded facies, but otherwise the two facies are similar in grading. Sediment from the bedded facies has an average size of 4–5 μm, and a clay-
sized content of 35–45%. Sediment from the mottled facies is slightly finer, having an average size of about 3 μm and a clay-sized content of 40–50%.

In the laminated facies, individual laminae are composed of well-sorted medium to coarse silt, interbedded with mottled clayey silts. At the base of the core, up to about 2 m above the Bothkennar gravel, the material is less well sorted, but has a similar average size and clay content. The sediment above the Cerastoderma bed is a rather different material, formed of silt with some clay (average size 8 μm and clay fraction 20–30%).

**Mineral composition**

The bulk mineral composition has been determined at 20 positions by X-ray powder diffraction, using fluorite as an internal quantitative standard. Additional tests have been carried out separately on the silt fraction and clay fraction, the latter by means of glycolated and furnace-dried samples in order to investigate the clay mineralogy in more detail.

Figure 9 shows the XRD traces obtained near the base (19-52 m) and top (3-15 m) of the core (below the shell bed). The traces are very similar, as are the other analyses from intermediate positions.

The results show that the mineral assemblage throughout the sediment is of almost constant composition. The relative strengths of the peaks vary, which suggests (with qualifications) that some variation occurs in the proportions of the minerals present, due to changes in the silt:clay ratio. In the coarse silt fraction, the dominant
mineral is quartz, with subsidiary feldspar and various ferromagnesian minerals. In the fine silt/clay fraction, the principal additional minerals are illite and chlorite, with probable biotite and possible minor smectite. Kaolinite was not recognized, although it might be expected as a felspar degradation product.

**Soil structure**

The bedded facies has an open structure based on domains of clay mineral particles arranged in a general honeycomb pattern (Fig. 10). Both edge-edge and edge-face contacts appear common, although clay particles have also been seen in silt-sized aggregates. Many silt particles float in the clay matrix, although occasional aggregates of silt grains are also seen. There is a wide range of pore sizes from 10–20 μm downwards. Occasional local bonding between silt particles also occurs, involving aluminosilicates, iron compounds or silica (Fig. 11).

By contrast, the mottled facies shows widespread evidence of presumed biogenic disturbance. For example, individual burrows have been observed with a cemented coating (Fig. 12), and an apparent internal mucus lining. Elsewhere, the clay platelets are organized in an open boxwork (Fig. 13) whose domain structure is relatively poorly developed. Silt particles appear less frequent and are not usually bonded.

Within both facies there is clear evidence of post-depositional alteration, although this appears to be erratic and selective. Individual silt
grains may appear sharp and angular, with clean, well-formed facets characteristic of mechanical (probably glacial) fracture. Alternatively others from the same area can appear heavily corroded, with evidence of secondary mineral growth (Fig. 14).

GEOLOGICAL SUCCESSION AND QUATERNARY HISTORY
Depositional processes and environments

Taking as a whole, the sedimentological evidence from HW3 and the adjacent bores suggests deposition under quiet water conditions not normally characteristic of intertidal deposits, and which therefore support the view that the Claret Beds were the product of an offshore, shallow water environment, mainly as a result of tidal transport. The principal control on the bedded-mottled facies transition appears to have been sedimentation rate and grain size. The laminated facies, which is of restricted development, is believed to have formed under localized conditions of erosion and deposition by temporarily stronger tidal currents.

The mottled facies is believed to have formed from originally bedded sediments due to post-depositional reworking by marine organisms, notably worms and other burrowers. This conclusion is based on a detailed comparison of the facies with published trace fossil assemblages, electron microscopy and discussions with colleagues in marine biology (Kingston, 1990). An alternative hypothesis—that the mottled facies has resulted from invasion of the sediment by plant roots (Hawkins, Lloyd & Nash, 1991)—is considered less likely due to the lack of in situ fibrous plant material and the implications of the sea level model (described later) for water depths, which preclude the growth of terrestrial or semi-aquatic plants.

In HW3 the laminated facies is considered to be a local replacement of the bedded facies as a result of temporary periods of more energetic wave and current conditions. This facies has been found previously in adjacent areas of the site.

Facies architecture and local stratigraphy

It is convenient to consider the detailed architecture of the sequence in terms of the lithological units L1–L5 (Fig. 2).

The sediments of L1—the lowest unit—are shelly, clayey silts which belong entirely to the bedded facies. The sedimentological and faunal evidence suggests that they are a condensed succession laid down in predominantly shallow marine conditions. Considerable current activity is suggested by the grading of the sand and by the laminations of sand and silt. The latter suggest analogies with sub-tidal channels. These sediments are stratigraphically equivalent to sediments found immediately above the Bothkennar gravel at many localities in the Bothkennar and Grangemouth areas, and have been correlated (Browne et al., 1984) with the silts of early Holocene (10000–9000 BP) buried beaches of the Forth (Letham Beds). However, radiocarbon dating suggests that this unit, although lithologically similar to the Letham Beds, is later in age. In this Paper, therefore, unit L1 is noted as Letham Beds, *sensu lato*.

The overlying silty clays (units L2–L4), which occur between -16.2 m and +1.4 m OD and thus comprise the bulk of the sequence, can be confidently correlated with the Claret Beds of Browne et al. (1984). In HW3 they are characterized by all the three facies described. Unit L3 is composed of sediment from the laminated facies, and units L2 and L4 of sediments from the bedded and mottled facies; the former dominate L2, whereas the latter are the more abundant in L4.

The bedding in both L2 and L4 is poorly defined, and in both cases bioturbated and non-bioturbated sediments alternate. There is little or no macrofauna, and an impoverished microfauna indicates marginal estuarine conditions (Robinson, 1990). The bioturbation usually amounts to less than 60% of the bed in which it occurs. The individual elements of the bioturbation are comparable in size to those of *Chondrites*, but they are not otherwise identical to this trace fossil, differing from it chiefly in the presence of a higher proportion of horizontal burrows with occasional vertical branches. These contrast with the many branches that are angled to the horizontal in *Chondrites* (cf. Bromley & Ekdale, 1984).

Unit L3 can be subdivided into laminated and bioturbated subfacies. The former recalls to some extent the laminated silts and clays which unconformably overlie the Claret Beds in the south-east of the site (Nash & Lloyd, 1988) and which are known elsewhere in the area as the Grangemouth Beds. On the basis of these laminations and the local occurrence of steep cross-beds, they have been attributed by Browne et al. (1984) to deposition under a tidal regime.

As already discussed, the complete sequence can be correlated with the Quaternary stratigraphy established for the Grangemouth area. The buried gravel encountered at -16.9 m OD is considered to be the Bothkennar gravel, which forms a marker horizon in this area. The succeeding beds can therefore be correlated on the basis of lithology with the Letham and Claret Beds, which form the bulk of the sequence. At
about 1.5 m OD there is a strong unconformity, above which occur sediments of the modern tidal flats and those sediments which accumulated following late 18th century reclamation work (Cadell, 1929), which extend to the ground surface.

Comparison with adjacent boreholes

The conditions of deposition of the deposits in HW3 can be further elucidated by considering other boreholes in the Bothkennar and Grangemouth areas, and also by examining wider evidence relating to the fauna and sea level changes in this part of the Forth Estuary.

The beds between the Bothkennar gravel and the Claret Beds corresponding to the lowest unit (L1) in HW3 have been intersected in several earlier boreholes both within and to the south and east of the Soft Clay Research Site, some of which have yielded a diverse fauna (Browne, 1987; Hawkins, Larnach, Lloyd & Nash, 1989; Graham, 1976). This fauna is indicative of water depths of at least 5–10 m, rather than the more shallow depths commonly supposed. For example, it is clear that the mollusca from these strata in BGS borehole B1 (no. 1 from the Bothkennar series (Browne et al., 1984)), although dominantly those which currently extend into very shallow water, also include species such as Nucula nucleus and Nucula minuta that are found offshore at depths greater than 6–10 m. Likewise the Claret Beds, where they are fossiliferous, have yielded diverse molluscan faunas consistent with similar offshore conditions.

The closest modern analogue to these faunas is the Abra assemblage which is currently widespread in the Firth of Forth south and east of the Forth bridges. (Elliot & Kingston, 1987). Although on an exposed coast molluscs can be cast up by wave action from depths of many metres, such conditions are unlikely to have occurred in this sheltered part of the estuary, and so the fauna is considered a reliable, if crude, depth indicator. In contrast, most faunal assemblages from the Grangemouth Beds at both Bothkennar and Grangemouth lack offshore elements, and are generally in keeping with deposition not far below low water.

To summarise, both the Letham Beds and the Claret Beds in the Bothkennar area were laid down offshore, probably at depths greater than 5–10 m, whereas most of the Grangemouth Beds were laid down in very shallow water.

Radiocarbon dating and comparison with the sea level curve

Three radiocarbon dates obtained at the Oxford accelerator dating laboratory are given in Table 1. One of these (OxA-3388) was carried out on insufficient carbon and must be regarded as notional, but the other two dates are considered reliable. They suggest that the Letham Beds were deposited before 5000 BP (i.e. radiocarbon years before 1950), and the Claret Beds between 5000 and 3000 BP. OxA-3387 is a date for the local base of the Grangemouth Beds where they lie unconformably on the Claret Beds, and so is not a maximum age for the former unit. Indeed, comparison with the Bothkennar series no. 22 borehole (Browne, et al., 1984) 2 km to the south suggests that deposition of the Grangemouth Beds began there at about 4000 BP, allowing a maximum time span for the deposition of the Claret Beds of no more than 1000–1500 years in the Bothkennar area generally. Other dates from the Grangemouth area quoted by Browne et al. (1984) present difficulties of interpretation, but suggest generally that deposition of the Claret Beds started earlier there than at Bothkennar. If so, the Claret/Letham bed stratigraphy is diachronous (Fig. 15), and deposition of the Letham beds at Bothkennar may well have overlapped with that of the Claret Beds at Grangemouth.

This evidence may be compared with the sea

<table>
<thead>
<tr>
<th>Borehole</th>
<th>Depth: m</th>
<th>Stratigraphic position</th>
<th>Sample details</th>
<th>Laboratory reference</th>
<th>Adjusted age$^{1}$ 14C years BP ± 1σ</th>
</tr>
</thead>
<tbody>
<tr>
<td>D2</td>
<td>−4.2</td>
<td>Base of Grangemouth Beds</td>
<td>Cerastoderma edule</td>
<td>OxA 3387$^2$</td>
<td>2945 ± 80</td>
</tr>
<tr>
<td>D1</td>
<td>−4.2</td>
<td>Mid Claret Beds</td>
<td>Saxicavella jeffreysi</td>
<td>OxA 3388$^3$</td>
<td>3825 ± 130</td>
</tr>
<tr>
<td>HW3</td>
<td>−16.5</td>
<td>Letham Beds</td>
<td>Spisula substranata</td>
<td>OxA 3389$^4$</td>
<td>5075 ± 90</td>
</tr>
</tbody>
</table>

$^1$ Adjusted ages based on an apparent age of 405 ± 40 for sea water.

$^2$ δ$^{13}$C PDB = +0.4 ppt.

$^3$ δ$^{13}$C PDB not measured, pretreatment incomplete.

$^4$ δ$^{13}$C PDB = +1.3 ppt.
level curve for this part of the Forth Estuary (Fig. 15). Specific details of the comparison are summarized in Table 2. This suggests that the water depth at the Soft Clay Research Site was decreasing throughout the Letham Beds/Claret Beds sequence, and fell from around 22 m below low water at the base of the former to about 6 m or less during the deposition of the Grangemouth Beds. These suggested depths may be compared with those derived from the faunal data presented earlier. For the Letham Beds the faunal evidence merely points to deposition at Bothkennar in water deeper than 5–10 m, and thus agrees with the 22 m deduced from the sea level curve. For the Claret and Grangemouth Beds the faunal evidence is again in keeping with the water depths suggested by a comparison of the dated levels with the sea level curve, i.e. more than 5–10 m depth (Claret Beds) and immediately sub-tidal (bulk of the Grangemouth Beds).

Table 2. Approximate water depths and levels in boreholes at Soft Clay Research Site, BGS Bothkennar and Grangemouth (compiled from Hawkins et al., 1989 (D2); Browne et al., 1984 (BGS Bothkennar no. 1, (B1)); this study (HW3))

<table>
<thead>
<tr>
<th>Borehole</th>
<th>Elevation: m OD</th>
<th>Water depth¹</th>
<th>5000 BP</th>
<th>3000 BP</th>
<th>Present</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Borehole HW3</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Top of L4</td>
<td>+2</td>
<td>NP</td>
<td>NP</td>
<td>NP</td>
<td>lwm</td>
</tr>
<tr>
<td>Base of L4</td>
<td>−5</td>
<td>NP</td>
<td>8?</td>
<td>BSS</td>
<td></td>
</tr>
<tr>
<td>Base of L3 and top of L2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Base of Claret</td>
<td>−17</td>
<td>22</td>
<td>BSS</td>
<td>BSS</td>
<td></td>
</tr>
<tr>
<td>Base of Letham Beds</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Borehole D2</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Base of Grangemouth Beds</td>
<td></td>
<td>−4</td>
<td>NP</td>
<td>6</td>
<td>BSS</td>
</tr>
<tr>
<td><strong>BGS Bothkennar B1</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Top of Claret Beds</td>
<td>2</td>
<td>NP</td>
<td>12/NP</td>
<td>BSS</td>
<td>Intertidal</td>
</tr>
<tr>
<td>Base of Claret Beds</td>
<td>−7</td>
<td>16/BSS</td>
<td>BSS</td>
<td>BSS</td>
<td></td>
</tr>
<tr>
<td>Base of Letham Beds</td>
<td>−11</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

¹ Depths are shown relative to low water, assuming the modern tidal range. NP bed not present at this time. BSS below contemporary sediment surface at the time stated.
GEOTECHNICAL CHARACTERIZATION

The results of characterization tests are now reported and the extent to which they may be influenced by variations in sedimentary facies is considered. These tests include not only standard geotechnical characterization tests, but also geochemical investigations of the pore water and various geophysical measurements.

Natural water content and plasticity

Water content and vane shear strength were both routinely determined at intervals of 50 mm and 150 mm respectively throughout the core. The profiles of water content (Fig. 16) and strength (Fig. 17) show the expected division into crust, transition and the main sequence reported by many investigators. The general trend in the main sequence is one of decreasing water content and increasing strength with depth, although in detail, small variations can be matched to sedimentological changes, and hence also follow facies changes.

The Atterberg limits have been determined at 1 m intervals. There are systematic differences between air-dried and undried samples (Hawkins et al., 1989), and between samples treated with peroxyde and the other samples. These differences are linked to the activity of the clay-sized fraction. It would be expected that the illite/chlorite/rock flour mixture found at Bothkennar would have a relatively low activity, perhaps around 0.5; however, as shown in Fig. 18, the activity of clay from soil tested from its natural water content can be up to 1.25. Such activity is normally due to smectite (swelling) minerals. However, from the XRD analysis these do not appear to be quantitatively significant at Bothkennar, and so another explanation must be sought.

After treatment with hydrogen peroxyde, the activity falls to about 0.4. This agrees more
closely with that expected from the mineralogy, and suggests that the higher values obtained from the untreated soil result from the presence of organic material. The highest values are usually associated with sediment from the bioturbated mottled facies. Only a small amount of detrital vegetable matter has been recovered and only a little organic material noted on micrographs, and so it is considered most likely that the organic material is present in the form of molecular residues from soft-bodied organisms responsible for the mottling, and attached directly to the clay minerals. Their exact nature is unknown, but it is speculated that those of importance in the present respect may be either large molecules such as polysaccharides, which can bridge adjacent clay platelets, or smaller residues which contain hydrophilic components such as hydroxyl and amino groups. These could increase the weight of water bound to a given weight of clay, so increasing its measured plasticity. Oxidation of these compounds by hydrogen peroxide reduces the plasticity to a value consistent with the inorganic mineralogy.

Air drying reduces the activity to a value between the values of the natural and treated soils (Fig. 18). This effect has been noted (in terms of plasticity index) by Hawkins et al. (1989). A possible mechanism might be the effective decrease in specific surface due to clay aggregation, as recently proposed by Pandian, Nagaraj & Sivakumar Babu (1991). A second (not mutually exclusive) mechanism, which follows from the present work, is partial oxidation of the organic residues, which itself introduces a partial change in the activity.

**Pore fluid chemistry**

Sodium, potassium, magnesium and calcium ions have been extracted using a lithium acetate exchange technique and analysed using atomic absorption spectrophotometry. Chloride was
extracted using distilled water and analysed by high performance liquid chromatography. The results are shown in Fig. 19. Due to retention on the clay minerals, it is not possible to determine the in situ cation concentrations by an extraction method, and so these results have been determined as total extracted weight (in milliequivalents)/soil solid weight.

That the divalent cations (Mg, Ca) are relatively constant with depth is attributed to binding on clay minerals. Sodium is relatively mobile and thus susceptible to leaching: the decrease in exchangeable sodium at the top and base of the core is attributed to freshwater input from the surface and the Bothkennar gravel, respectively. The pattern for the chloride anion is similar to that for sodium, as might be expected; both correlate with the undried activity profile (Fig. 18), which in turn is believed to reflect the profile of organic content. The peak values obtained are 15–20 parts per thousand (ppt) at a depth of 9–10 m. These are similar to the adjacent estuary water, and it seems reasonable to suppose that to a first approximation the pore fluid composition derives from that of the adjacent estuary. The reduction in values above and below this depth is again attributed to dilution by fresh water from the ground surface and from the Bothkennar gravel.

**Engineering geophysics**

Measurements have been made of P wave velocity at 150 kHz (Fig. 20), and electrical impedance at 1 kHz and 10 kHz (Fig. 21). The results show that, below the crust and associated transition, the velocity is consistently 1500–1600 m/s, and that this velocity is substantially isotropic. The resistivity immediately below the crust is 1–2 Ωm, rising to 4 Ωm as the Bothkennar gravel is approached; this pattern follows inversely the fall in water content. Together with
the low absolute values, this suggests that the resistivity is principally controlled by the pore water and its electrolyte content. In the crust and transition zone, both the acoustic and impedance results are more complex, due to the influence of partial saturation and sedimentological changes. Velocities of 1000–1500 m/s have been recorded and resistivity values reach 40 Ωm.

DISCUSSION OF RESULTS
This study has enabled the deposits to be subdivided on the basis of sedimentary facies. It is of interest to discuss to what extent the engineering properties of the deposit will be related to this subdivision. It is convenient to consider the properties under two headings: those which are controlled by some aspect of the soil skeleton, and those which are controlled by the pore fluid. The former may be expected to be influenced by the sedimentary facies, whereas the latter are not facies-dependent.

The electrical resistivity and acoustic velocity are facies-independent, as shown by the results presented. The first is unsurprising, since it relates to the invasive input of groundwater the properties of which are apparently acquired elsewhere. The last requires some explanation, since P wave velocity would be expected to be controlled by soil structure. The explanation may well be that in soft soils the first arrival P wave is transmitted via the soil water. This accounts for the lack of anisotropy and the velocity close to 1500 m/s. A similar feature is seen in shallow marine clays (Paul & Jobson, 1987). At greater depths than those investigated in this study, soil structure may be expected to have an effect. A corollary of this explanation is that S wave velocities, although not measured in this investigation, may be expected to be facies-linked.

It is anticipated that solid plasticity, catiopermeability, stiffness and yield will be facies-linked to some extent, although the magnitude of the differences may well be small. The principal facies differences below the weathered zone are the presence of primary sedimentary bedding, together with limited particle bonding (bedded facies); the presence of closely spaced silt laminations (laminated facies); and the homogenization of the sediment by burrowing organisms, combined with enhanced organic content and cation exchange capacity (mottled facies).

On the basis of a simple view of solid properties it is anticipated that the presence of interparticle bonding will increase the stiffness of the soil, and will also increase the stress at which the elastic-plastic transition occurs. Thus sediment from the bedded facies will be (slightly) stronger and stiffer than that from the mottled facies.

The presence of silt laminations within the bedded and the laminated facies will clearly increase their permeability, reduce consolidation times and introduce anisotropy.

This study suggests that the organic content provides a control on soil plasticity, and hence that some increase in plasticity would be expected in the mottled facies. Conversely it would be expected that, compared with the other two facies, material from the mottled facies should possess greater isotropy, be less stiff as a result of a lack of particle bonding, and be less permeable due to the destruction of horizontal fabric.

CONCLUSIONS
It is possible to divide the sediment into four facies: bedded, laminated, mottled and weathered. These facies have genetic significance and are believed to correlate with the variation in some soil properties. The sediment originated by tidal transport in an offshore situation, the sedimentary structures from which are recorded by the bedded facies. Locally more energetic conditions gave rise to the laminated facies. Reduced rates of sedimentation at times allowed burrowing organisms to rework the sediment and so to produce the mottled facies. The weathered facies is a near surface zone which has been affected by desiccation, oxidation and modern pedogenesis.

The grain size distribution shows facies-based variations with depth, superimposed on a general upward coarsening trend. The principal clay mineral is illite. The silt fraction contains a mixed assemblage of quartz, felspar and ferromagnesian minerals.

SEM has revealed that the clay microfabric is a generally open boxwork, dominated by edge contacts. The mottled facies has a microfabric which shows evidence of disturbance by presumed biogenic activity. The bedded facies shows a more closely packed structure with a range of domain sizes and voids, with some bonding between silt particles. There is evidence in both facies of chemical corrosion and growth of secondary minerals.

As reported by other workers, soil plasticity is much affected by drying. This study has also suggested that organic residues are the cause of enhanced plasticity compared with that expected from the clay mineral suite. The water content and laboratory vane strength are similar to those reported by other workers.

The principal ions present in the soil water are sodium, and chloride. Potassium, magnesium, calcium and sulphate are present to a lesser extent. The ionic concentration appears to be similar to that of the adjacent estuary water, with
some dilution by freshwater near the surface and from the Bothkennar gravel.

The soil has an electrical resistivity of usually less than 4 Ohm, which is controlled by the water content and pore-water solutes. The acoustic velocity is isotropic and equivalent to that in water. The bulk magnetic susceptibility is generally low, except in the laminated facies, where it is raised by the presumed presence of magnetite.

Most of the succession above the Bothkennar gravel can be correlated with the Clarot Beds which occur widely in the area. The lowest part of the sequence has some similarity to the Letham Beds, but the exact correlation is suspect. The Clarot Beds are succeeded by modern tidal flat deposits, which rest on them with marked disconformity. Radiocarbon dating indicates their age to be 5000–3000 BP. The borehole depths and the sea level curve together suggest that the Clarot Beds accumulated offshore in water whose depth reduced from about 20 m to 5–10 m, which agrees broadly with the available faunal data.

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