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The Holocene Evolution of Romney Marsh: A Record of Sea-Level Change in a Back-Barrier Environment.

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with the requirements of the
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Abstract.

The reliability of sea-level trends obtained from sedimentary records in back-barrier environments is open to question. Indeed, the protective gravel barrier may control back-barrier sedimentation to such an extent that changes in marine and freshwater influence may be recorded independent of sea-level tendency (Duffy *et al.* 1989; Shaw and Forbes, 1987). In the Romney Marsh region, the Holocene sediments deposited behind the barrier complex of Dungeness, provide an ideal record with which to investigate the inter-relationship between barrier and back-barrier environments, as well as the interaction of sea-level rise, storms and sediment supply in the control of back-barrier sedimentation.

Stratigraphic, granulometric, magnetic and micropalaeontological analyses are utilised in a detailed reconstruction of the barrier / back-barrier interface at Scotney Marsh. At this interface the Holocene sedimentary succession records a marine regression at ca. 3000 cal. yrs. BP, from intertidal mudflats to terrestrial peat, and then marine transgression at ca. 2700-2500 cal. yrs. BP, with the return of intertidal mudflat sedimentation. A high rate of sediment supply (exceeding the rate of sea-level rise) facilitated marine regression despite the rising sea-level. The existence of the authigenic mineral greigite (Fe_3S_4) has, however, prevented the determination of the provenance of the marsh sediments using mineral magnetic techniques in this study.

Integration of this and other site-specific palaeoenvironmental reconstructions has overcome some of the problems of distinguishing between processes operating over differing spatial and temporal scales. Indeed, it has been demonstrated that in the Romney Marsh region, the rate of sea-level rise decelerated between ca. 6000 and 3000 cal. yrs. BP (from ca. 2.3mm/yr to ca. 0.8mm/yr), although a slow but perceptible rise is recorded throughout this period. At some time after ca. 3000 cal. yrs. BP, an acceleration of the rate of sea-level is recorded. Thus, an unequivocal record of Holocene sea-level changes is preserved in the back-barrier sediments. As a consequence of this investigation, a detailed model of the Holocene evolution of Romney Marsh is proposed.

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1 Introduction.

In this study, attempts will be made to resolve both methodological issues, relating to the study of past sea-levels as recorded in back-barrier environments and to utilise established techniques of palaeoenvironmental investigation to determine the Holocene evolution of Romney Marsh. Therefore, in this chapter, the aims and objectives of the study will be outlined, and an introduction to the study area will be completed.

1.1 Aims and Objectives of the Study.

It has been stated that sediments deposited behind gravel barriers may confuse sea-level reconstruction (Orford *et al.*, 1991). Equally, Duffy *et al.*, (1989) and Shaw and Forbes (1987) suggest that the gravel barriers themselves may control the back-barrier environment to such an extent that variations in marine and freshwater influence may be recorded independent of sea-level tendency. Despite this uncertainty regarding the suitability of utilising back-barrier sediments for sea-level reconstruction, the back-barrier sediments of the Romney Marsh area have been utilised in a number of recent studies considering past sea-levels (Tooley and Switsur, 1988; Waller *et al.*, 1988; Tooley, 1990; Long and Innes 1993, 1995a&b; Long *et al.*, 1996).

In contrast to the above view, some authors (Rampino and Sanders, 1981; Nichols, 1989) have suggested that the major controlling influence on back-barrier sedimentation is the rate of sea-level rise and not the barrier. It can be seen, therefore, that some debate exists as to whether back-barrier sediments preserve a reliable record of past sea-level changes. Recently, back-barrier sediments have been widely utilised in Holocene sea-level reconstructions, especially in south-east England where few (if any) other environments exist from which past sea-level records can be obtained. Indeed, studies have been carried out in Romney Marsh, Coombe Haven and Pevensy Levels (Jennings and Smyth, 1987; Smyth and Jennings, 1988), Stansore Point (Long and Tooley, 1995) and also in the East Kent Fens (Long, 1992). It is, therefore, of critical importance that the suitability of sediments deposited in back-barrier environments for sea-level reconstruction be established.

Aim 1 : To investigate whether a reliable proxy record of past sea-level is preserved in sediments that have accumulated behind a protective gravel barrier during the Holocene.

It has been stated that the most extensive gravel barrier in the United Kingdom is the Dungeness Foreland (Carter *et al.*, 1989), which forms the protective gravel barrier behind which the sediments of Romney Marsh have been deposited. Carter *et al.*, (*op.cit.*) also proposed that this barrier / back-barrier environment provided the best site in the United Kingdom to examine the inter-dependence of gravel barriers and back-barrier sedimentation. Indeed, therefore, this site also provides an excellent site at which to examine whether a reliable record of past sea-level changes can be reliably resolved *via* the study of back-barrier sediments.

In investigating the back-barrier sediments for past sea-level reconstruction, an attempt will be made to determine the trend of past sea-level changes that have occurred in the Romney Marsh area during the Holocene.

Aim 2 : To determine a record of past sea-level changes in the Romney Marsh area and south-east England during the Holocene.

In attempting to reconstruct the nature of and control on the palaeoenvironments recorded in the back-barrier, a combined approach will be utilised. First, large-scale stratigraphic investigations of the sediments of Romney Marsh were to be completed to provide stratigraphic integration of almost all of the site-specific studies completed on Romney Marsh. These stratigraphic transects were also to be utilised in order to establish the most effective location in which to complete the second phase of this combined approach; a high-resolution site-specific palaeoenvironmental investigation. It was determined that the most effective area in which to complete the detailed site-specific investigation would be at the barrier / back-barrier interface, as this environment might be the most

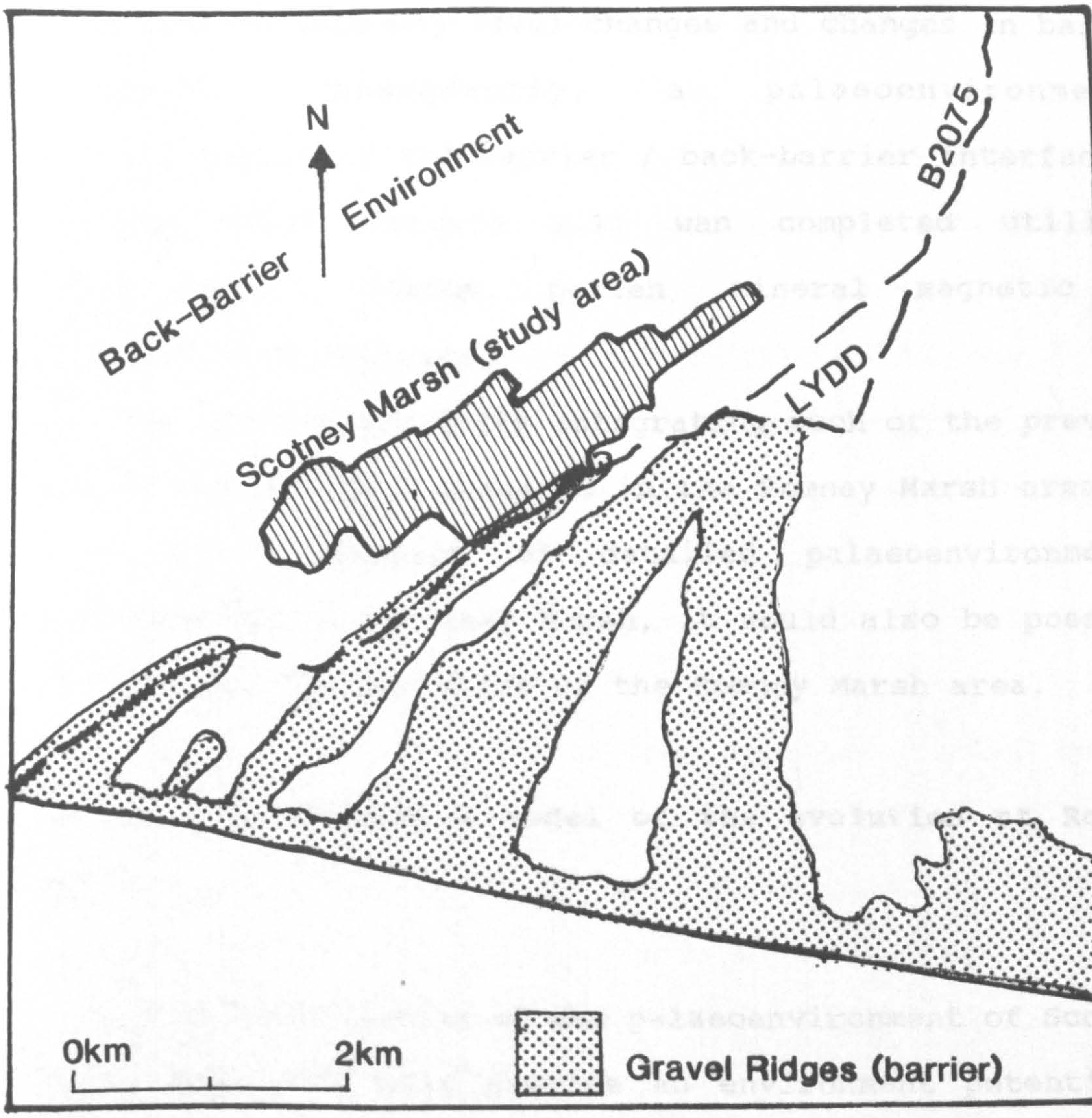


Figure 1.1: Location of Scotney Marsh (study area) at the barrier / back-barrier interface.

sensitive to both sea-level changes and changes in barrier dynamics. Consequently, a palaeoenvironmental reconstruction of the barrier / back-barrier interface at Scotney Marsh (figure 1.1) was completed utilising stratigraphic, diatom, pollen, mineral magnetic and particle size analyses.

By stratigraphically integrating much of the previous palaeoenvironmental research in the Romney Marsh area and also the completion of detailed palaeoenvironmental investigation at Scotney Marsh, it would also be possible to consider the evolution of the Romney Marsh area.

Aim 3 : To propose a model of the evolution of Romney Marsh.

The investigation of the palaeoenvironment of Scotney Marsh would not only provide an environment potentially sensitive to changes in sea-level and barrier dynamics, but would also be an investigation in an area of Romney Marsh hitherto barely studied. Only one previous palaeoenvironmental study has been completed in the fore-marsh areas of Romney Marsh, by Tooley and Switsur (1988) at Broomhill. This is in contrast to the mid-marsh (Field, 1983; Everett, 1985; Long and Innes, 1993, 1995a&b) and the back-marsh (Waller *et al.*, 1988; Tooley and Switsur, 1988; Waller, 1993, 1994; Long *et al.*, 1996) areas in which, relative to the fore-marsh, a considerable amount of study has been completed.

Aim 4 : To attempt to determine the provenance of the sediments that make up Romney Marsh.

In an attempt to further resolve the detail of the evolution of the Romney Marsh area, an attempt was also to be made in this study to determine the source of the sediments of the marsh. The technique utilised for the study marsh sediments provenance was to be environmental magnetism. Previously this technique has been used elsewhere on Romney Marsh at Denge Marsh in an attempt to determine the source of the sediments with only limited success (Spencer, 1992). It was considered that by the application of the technique to a new area and also by the utilisation of a different approach, that the problems experienced by Spencer (*op.cit.*) could be overcome and, thus, to be able to establish the source of the large quantities of sediment that make up the back-barrier stratigraphy of Romney Marsh.

1.2 Introduction to the Study Area.

The area referred to as Romney Marsh, or the Romney Marsh area, in this study is actually made up of Walland Marsh, the adjacent smaller levels and marshlands of the tributary river valleys (Rother, Brede and Tillingham), the gravel of the Dungeness Foreland, and also to Romney Marsh proper to the north-east of the Rhee Wall (figure 1.2, Eddison and Green, 1988). The Romney Marsh area is situated in the south-east of England on the border between East

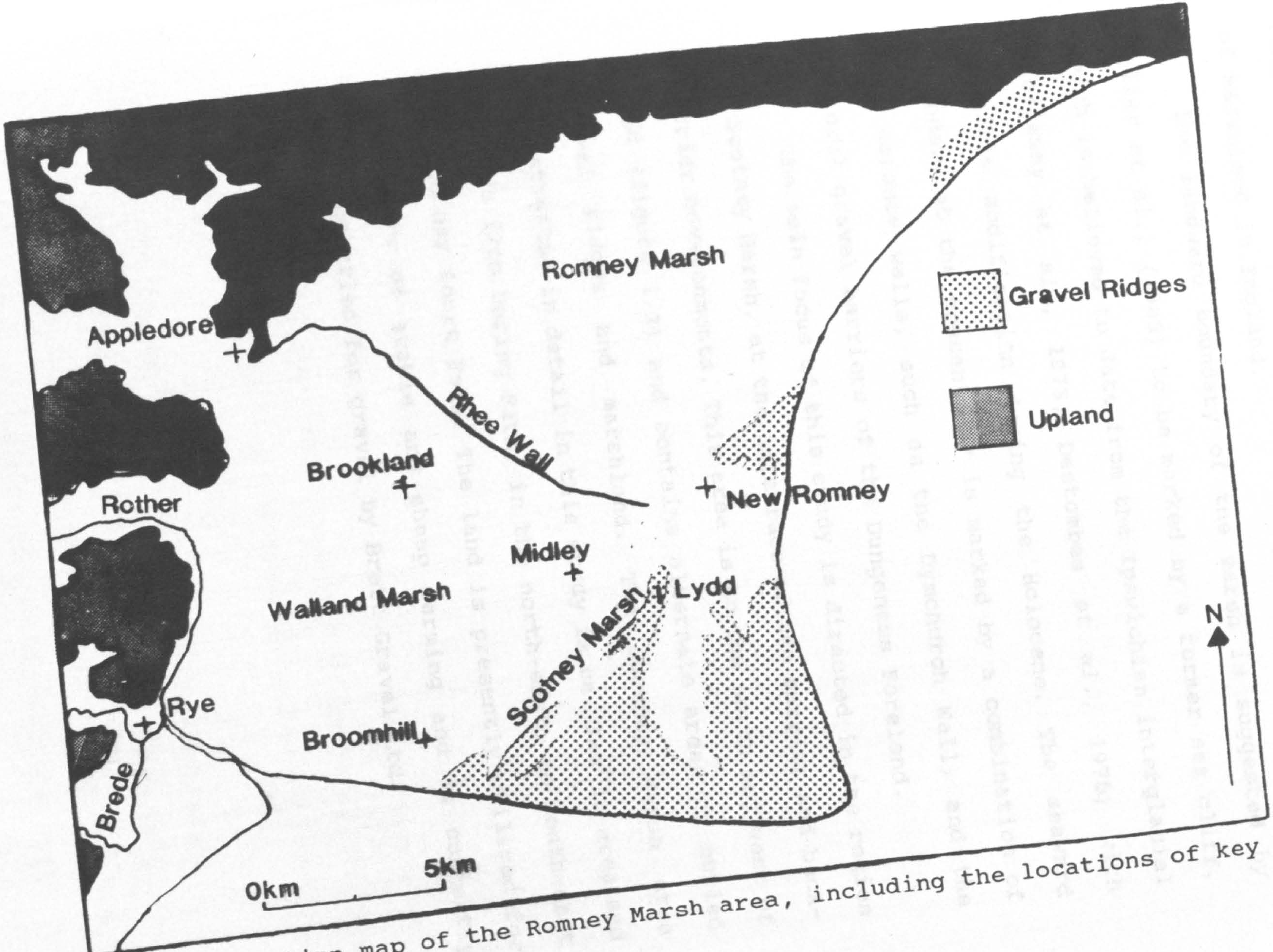


Figure 1.2: Location map of the Romney Marsh area, including the locations of key sites.

Sussex and Kent. Romney Marsh constitutes an area of 27,000 hectares and is, therefore, one of the three largest areas of marshland in England.

The landward boundary of the marsh is suggested by Waller *et al.*, (1988) to be marked by a former sea cliff, which is believed to date from the Ipswichian interglacial (Kellaway *et al.*, 1975; Destombes *et al.*, 1975) with possible modification during the Holocene. The seaward boundary at the present day is marked by a combination of sea defence walls, such as the Dymchurch Wall, and the natural gravel barriers of the Dungeness Foreland.

The main focus of this study is directed in the region of Scotney Marsh, at the interface of the barrier and back-barrier environments. This area is to the north and west of Lydd (figure 1.2) and contains alternate areas of buried gravel ridges and marshland. The Scotney Marsh area investigated in detail in this study is ca. $3km^2$ in area and extends from Dering Farm, in the north-east, to south-east of Scotney Court Farm. The land is presently utilised for a mixture of arable and sheep farming and is currently being quarried for gravel by Brett Gravel Ltd.

2 Literature Review.

2.1 Evolution of Romney Marsh : Background.

Romney Marsh is one of the three largest areas of marshland in England (Eddison and Green, 1988), the other two being the Fens and the Somerset Levels. In this study to determine the evolution of Romney Marsh, it is first of all necessary to thoroughly review the research that has been completed previously regarding Romney Marsh and, secondly, to review the theoretical and process-related studies of similar barrier and back-barrier environments. The evolution of Romney Marsh has been influenced by a number of inter-related processes; the rise in sea-level experienced in the south-east of England during the Holocene, the presence of a system of gravel barriers, and the complex drainage of the back-barrier environment are probably the most important of these processes. Therefore, an understanding of the published material on these sites and processes will assist both in incorporating the site-specific study at Scotney Marsh into a model of the evolution of Romney Marsh and in the consideration of the evolution of Romney Marsh in terms of regional sea-level change and barrier / back-barrier development.

2.1.1 The History of the Research of Romney Marsh.

The research into the evolution of Romney Marsh has taken place in a number of distinct stages. It has been stated by Eddison and Green (1988) that the study of the geographical evolution of Romney Marsh has lagged behind that of the other two large areas of coastal marshland in England. This statement will be examined here as the stages of palaeogeographical research are considered.

2.1.1.1 Early Research (pre-1968).

Early theories regarding the evolution of Romney Marsh came from writers such as Holloway in 1849. However, Cunliffe (1980) believes that the first serious attempt to understand the early history of the marsh was made by the local engineer James Elliot, who outlined his views on the natural changes of the marsh which took place during the Roman period (Elliot, 1847; 1852; 1862). The next major contribution to the understanding of marsh evolution was made by Gordon Ward. He utilised topographic evidence from Saxon charters, from which he attempted to build a picture of the marsh in pre-Norman times (Ward, 1931a&b; 1933a&b; 1936). It was also in the 1930s that W.V. Lewis considered the formation of Dungeness foreland (Lewis, 1932), whilst the broad history of the marsh was studied by Gilbert (1933) and Homan (1938). Lewis and Balchin (1940) further investigated the evolution of the Dungeness foreland and related this to changes in mean sea-level since the Roman period. Towards the end of the 1950s a further source of research was established, providing detailed information regarding the soils and upper sediments of Romney Marsh, in the early studies of the Soil Survey of England and Wales (Green and Askew, 1958a&b; 1959; 1960).

2.1.1.2 The Soil Survey of Romney Marsh.

It has been suggested (Cunliffe, 1980) that no adequate overall consideration of marsh evolution was available until the systematic soil survey of the entire area of Romney Marsh by the Soil Survey of England and Wales. This was written by Green (1968), and spawned a new era of research in providing a robust spatial framework for investigation (Eddison and Green, 1988). Indeed, Eddison and Green proposed that all current research regarding the evolution of Romney Marsh must be set against the highly detailed record of the upper stratigraphic levels and the topographic features provided by Green (1968) in the Soil Survey.

This very important work does, however, have some drawbacks. Green divided the marsh into 'younger' and 'older' marshland on the basis of calcium content of the soils, and also provided two radiocarbon ages of only very local significance. Consequently, no timescale of marsh evolution was established. In addition, the study did not consider some of the bordering areas of Romney Marsh, *i.e.* Pett Level and the adjacent valleys (Eddison and Green, 1988).

2.1.1.3 The Romney Marsh Research Trust.

As has already been mentioned, Eddison and Green (1988) noted that the palaeogeographical research in Romney Marsh had lagged behind that of the Fens and the Somerset Levels. Consequently, the Romney Marsh Research Group was established in 1984 to provide the interdisciplinary approach that Eddison and Green (1988) considered necessary in an area of marshland where evolution has been influenced by both natural and anthropogenic forces. A similar approach had been successfully applied to the study of the Fens by the Fenland Research Committee, combining geomorphological, archaeological and historical research.

In 1987 the Romney Marsh Research Trust was established with two major goals; first, to promote, support and co-ordinate multi-disciplinary research, and secondly, to publish the results at both a local and national level (Eddison and Gardiner, 1995). To this end two monographs, the first being *Romney Marsh : Evolution, Occupation and Reclamation* (Eddison and Green, 1988), and the second *Romney Marsh : The Debatable Ground* (Eddison, 1995), have published the inter-disciplinary research of recent years.

In recent times, therefore, the research into the evolution of Romney Marsh has fast been reaching the stage that research into the other large marshland areas has attained. It can be seen that a wide-ranging literature base is becoming available, producing a wealth of research which is utilised here to present a review of the evolution of Romney Marsh.

2.1.2 The Initial Protective Barrier.

It is unlikely that the fine-grained inorganic and organic sediments that make up Romney Marsh could have been deposited or preserved were it not for the protection of the sand and gravel barriers that have existed along its boundaries throughout its history (Gulliver, 1897; Lewis and Balchin, 1940). This barrier would have created a quiet-water depositional environment, behind which the sediments of the marsh (such as the blue clay, peat and younger alluvium of Green (1968)) could accumulate. This model of evolution is supported by the works of Livett (1930), Ward (1934), Cunliffe (1980), Eddison (1983a), Lake and Shephard-Thorn (1987), and Greensmith and Gutmanis (1990). However, the nature of this barrier has been the subject of much debate, especially in terms of its initial formation, location and composition. Topley (1885) stated that the cause of the original formation of Romney Marsh is entirely unknown, whilst Mackinder (1907) suggested that 'the mind staggers at the complexity of the Fairlight to Hythe gravel bar formation'. It is vital, therefore, that this barrier be considered further in the understanding of the evolution of the marshland.

2.1.2.1 The Form of the Initial Barrier.

Green (1968) suggested that the initial barrier related to the 'Midley Sand' and proposed that the word 'midley' is derived from the Saxon 'middle island' Green also suggested that the Midley Sand played a vital role in the formation of the marsh stratigraphy. This sand formation acquired its name from the original sand bar outcrops at Midley, north of Lydd. Green has suggested that this Midley Sand shelves greatly under the younger deposits of peat and alluvium. It was proposed that the gravel barrier deposits were then deposited on to this sand bar (Furley, 1880; Cunliffe, 1980; Greensmith and Gutmanis, 1990). Similarly, Williamson (1959) envisaged a string of low islands running from Hastings to Hythe (see figure 2.1), with gravel being piled up along this line forming a shallow lagoon within.

The majority of theories on the evolution of Romney Marsh propose that marsh sediments where deposited behind a north-east to south-west trending barrier. Following the work of Green (1968), most writers, e.g. Cunliffe (1980) and Greensmith and Gutmanis (1990), have accepted that the Midley Sand was this barrier (Long and Innes, 1993). Greensmith and Gutmanis (1990) proposed that the Midley Sand was the initial barrier that developed between 5742-5718 and 2765-2741 cal. yrs. BP, and this formation is pivotal to their model of marsh evolution. Green (1968) stated that outcrops of the Midley Sand at Broomhill, Midley, and north-east of Dymchurch possibly indicated a discontinuous ridge, maybe a sand spit system or dunes. Consequently, Greensmith and Gutmanis (1990) believed that gravel was deposited on the seaward side of this series of banks. However, at Broomhill, Midley Sand occurs between gravel ridges (Green, 1968; Tooley, 1989; 1990b) and, thus,

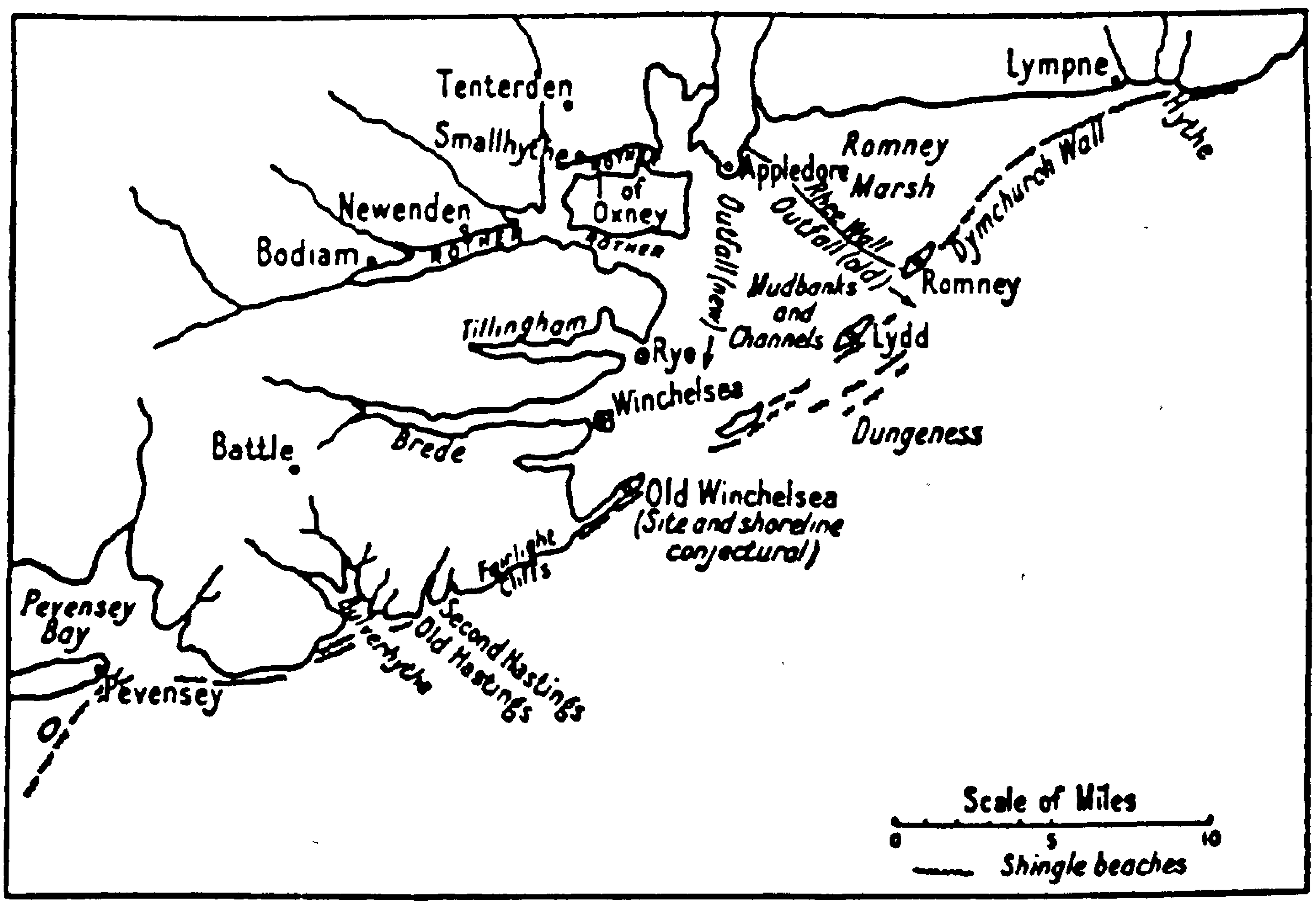


Figure 2.1: Initial low lying barriers across Rye Bay.

Source: Williamson (1959).

post-dates gravel deposition (Tooley, 1995).

Green (1968) proposed that the Midley Sand related to the basal sand; the sand found over much of the marsh, which has an undulating surface and outcrops as the higher Midley Sand deposits, e.g. at Midley Church Bank. This would mean, therefore, that where the sand outcrops no peat or younger alluvium would be expected below. Long and Innes (1993; 1995b) investigated this and demonstrated that, at Midley Church Bank, sand was continuous in only 4 of 26 boreholes and that, in fact, peat extended under much of this feature. This demonstrates that Green's (1968) initial interpretation of the Midley Sand at the typesite of Midley is in error. Furthermore, Long and Innes (1993; 1995b) determined that the lower sand was deposited before 3568-3487 cal. yrs. BP and the upper outcropping sand was deposited after 1813-1747 cal. yrs. BP. The 'Midley Sand' is, thus, a relatively young feature on Romney Marsh and cannot have been an initial protective barrier.

It appears, therefore, that the initial barrier was comprised predominantly of gravel and not sand (Long and Innes, 1993). Greensmith and Gutmanis (1990) suggest that the gravel deposits of the Dungeness foreland lie on a pre-Holocene erosion surface between altitudes of -32 and -35m OD, cut across the Lower Cretaceous Hastings Beds. In contrast, Eddison (1983a) believes that there is no indication that any bedrock structure has affected the formation or the position of the Dungeness foreland. Jennings and Smyth (1990) support this in proposing that

Dungeness has not accumulated in the shelter of a headland, but has developed on an open coastline. They suggest that this accumulation of gravel can only be explained by inputs exceeding outputs, and that any theory of the development of the Dungeness foreland must account for this massive accumulation on a currently high energy coast with a strong uni-directional longshore drift.

2.1.2.2 The Location of an Initial Barrier.

It is proposed that an initial gravel barrier stretched across Rye Bay from around Fairlight in the west to around Hythe in the east (Elliot, 1847; Lewin, 1862; Gulliver, 1897; Lewis, 1932; Lewis and Balchin, 1940; Gallois and Edmunds 1965; Green, 1968; Cunliffe, 1980; Eddison, 1983; Greensmith and Gutmanis, 1990; Long and Innes, 1993). Elliot (1847) suggested a natural barrier of gravel extended from Fairlight Head, through Lydd, New Romney and then towards Hythe. Lewin (1862), Drew (1864), and Burrows (1884) agreed with Elliot in placing the original barrier approximately along this line. In addition, Long and Innes (1993) noted that a marked difference in depositional environment existed on either side of the Lydd gravel and, thus, cautiously interpreted the gravel complex of Lydd to Broomhill to be part of an initial gravel beach barrier.

Lewis (1932), from examination of the gravel ridges at Camber Castle, proposed that deposition appeared to have started at a bend in the shoreline at Fairlight. The high supply of gravel and the longshore drift, caused by the predominant south-west winds, allowed a gravel barrier to grow across Rye bay (Drew, 1864; Mackinder, 1907; Livett, 1930; Ward, 1934; Steers, 1963; Eddison, 1983a).

The balance of opinion, therefore, appears to indicate that an initial gravel barrier extended across Rye Bay from around Fairlight in the south-west towards Hythe in the north-east. This barrier passed through Broomhill and Lydd, but, its exact location is, as yet, unresolved.

2.1.2.3 The Cause, Sediment Source, and Mode of Deposition of the Initial Barrier.

Drew (1864) suggested that the cause of deposition of the sediments making up an initial sand and gravel barrier was the slack water from the meeting of two tides; one from the English Channel and the other from the North Sea. This was refuted by Redman (1852), who cited the presence of similar deposits at Langley Point. However, Lewis (1932, 1937) suggested that tidal confluence occurred over a wide area from Dungeness to the Goodwins, off the coast of Deal, eastern Kent.

Elliot (1847) proposed that an island of Hastings' Sands formed a nucleus for further deposition of sand and gravel, although he later believed that this accumulation was aeolian in nature (Lewis, 1932). Cunliffe (1980) believed that deposition of the sand and gravel bar occurred because Rye Bay had become choked with debris as sediment accumulation outstripped erosion.

A possible mode of the transport of the gravel and the subsequent deposition and formation of the initial barrier, as discussed by Long *et al.* (1996), is landward barrier migration under a rising sea-level (Forbes *et al.*, 1991; Orford *et al.*, 1991). Indeed, Jennings and Smyth (1982) have proposed this as the cause of the formation of the gravel barrier at Langley Point.

Some debate exists concerning the route which the gravel has taken to be deposited at Dungeness, and, therefore, the source of the gravel deposits. Jennings and Smyth (1990) demonstrate that two possible sources of gravel exist; cliff retreat and the offshore zone. They suggest that there has not been sufficient Holocene weathering of the Sussex cliffs to provide the quantity of gravel to account for the deposits around Dungeness. It is, therefore, proposed that the offshore zone must have contributed a significant proportion of the gravel to the sediment processing system. During a glacial period, the sea-level would be lower and so the rivers and glaciers would have transported their sediment to the, then, exposed sea floor. Kerney (1963) and Williams (1980) believe that the cold stages of the Pleistocene would provide the glacial and periglacial conditions necessary to weather flint from Chalk and to deposit it on the floor of the English Channel.

Greensmith and Gutmanis (1990) note that the Dungeness gravel is very rounded, and may date from at least the

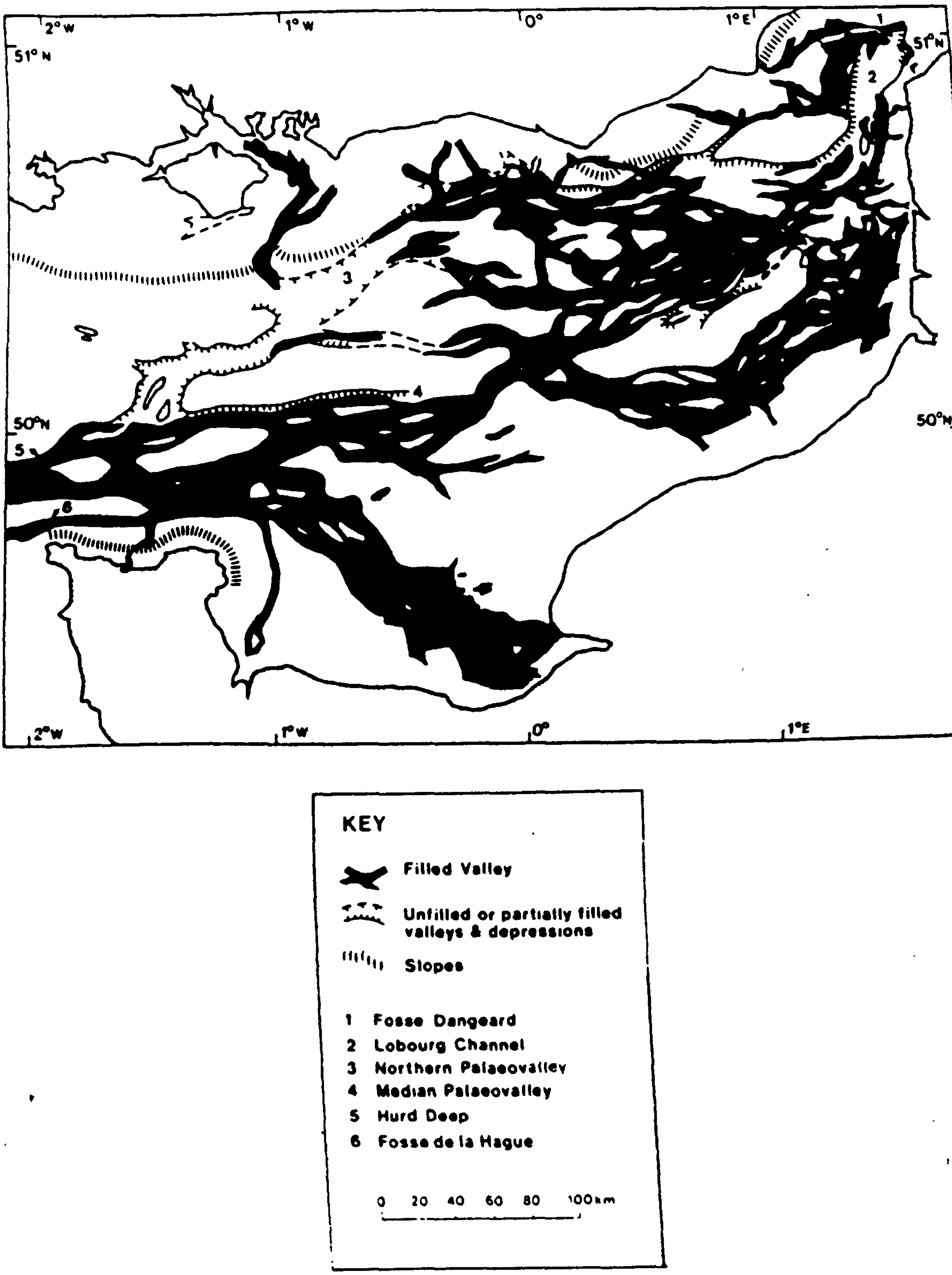


Figure 2.2: Palaeovalley system of the eastern English Channel.

Source: Smith (1985), after Auffert *et al.* (1982).

Quaternary. They also describe flint gravel deposition in the Pleistocene on the flanks of an offshore palaeovalley system. The location and morphology of these palaeovalleys are illustrated in figure 2.2. Smith (1985) outlines that the palaeovalley systems in the English Channel are filled with coarse gravel and, thus, are a potential source of sediment. As sea-level has risen during the present interglacial, the deposits have become available for reworking and have moved towards the coast in the advancing surf zone (Eddison, 1983a; Jennings and Smyth, 1990). During the Holocene, little fresh material has been contributed, although the redistribution of flint has increased.

Jennings and Smyth (1990) also propose that the Dungeness gravel may be predacious, feeding on the periodic destructions of up-drift gravel sources on the south coast, such as the Crumbles to the west. Nicholls (1991) also suggested that gravel may have become available from temporary sinks to the west to supply the growing Dungeness, and stated that the process by which gravel was transported was eastward longshore drift during the Holocene. In addition, this transport was considered to be variable, with increased storminess leading to an increase in gravel supply. Nicholls (*op. cit.*) also proposed that Dungeness was the easterly sink for the longshore drift cell extending from Selsy Bill to Dungeness.

Long *et al.* (1996), therefore, propose that the source of the gravel was probably a combination from sources to the west, in Sussex, *via* longshore drift, and from onshore movement of offshore sources. Eddison (1983a) outlined that a further source of gravel for the more recent deposits of the Dungeness foreland was *via* reworking of earlier beaches, eroded as the south coast is eroded back (figure 2.3). Hey (1967) noted that the contemporary supply of gravel is much lower than in former times, which may account for the 'cannibalisation' of earlier beaches (Lewis, 1932).

Clearly, longshore drift has been a very important process in transporting gravel to the Dungeness foreland. This was first described by Palmer (1834). He stated that when waves break obliquely on the beach, the swash carries gravel up the foreshore in the direction of motion, whilst the backwash then returns on the line of greatest slope, taking gravel with it. Therefore, the gravel zigzags up-drift along the beach. Eddison (1983a) agreed with this, and demonstrated that the gravel of the previous ridge is imprisoned and, therefore, not free to be carried in a north-easterly direction with the longshore drift.

It appears, therefore, that the gravel of the initial barrier accumulated *via* the dual processes of a swash-aligned barrier migrating onshore, with rapidly rising sea-level from the gravel sources on the floor of the English Channel, and also the longshore movement of gravel transporting course sediment form the west and leading to the barrier becoming more drift-aligned.

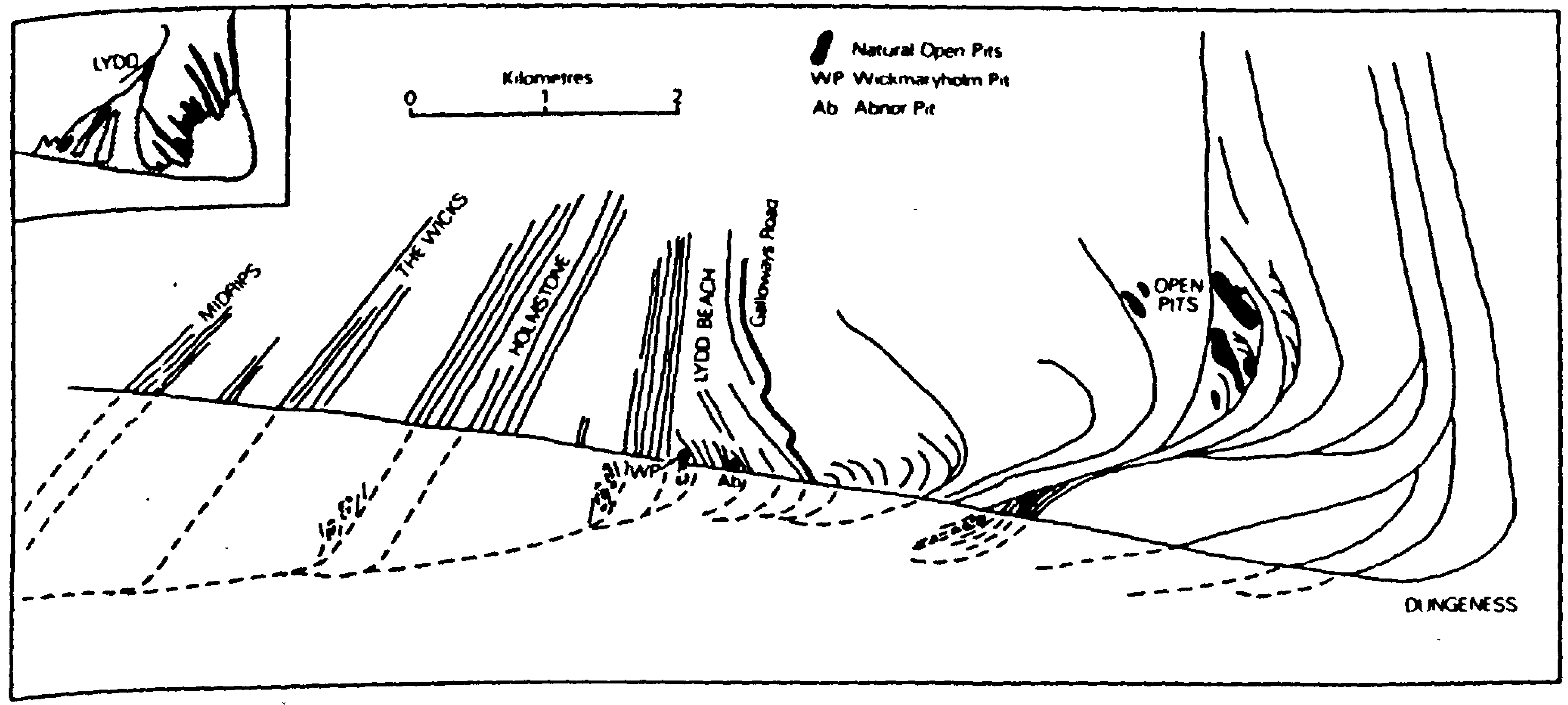


Figure 2.3: The gravel ridge populations of the Dungeness Foreland. Demonstrating changing orientation and reworking of gravel ridges.

Source: Eddison (1983a).

2.1.2.4 The Age of the Initial Barrier.

The age of the initial gravel barrier has been approached by a number of authors. Eddison (1983a) placed the arrival of gravel at about 5742-5718 cal. yrs. BP whilst Tooley (1995) stated that gravel probably began to arrive in the area no earlier than 6868-6792 cal. yrs. BP. Tooley and Switsur (1988) demonstrated that the arrival of gravel was not later than ca. 3305-3226 cal. yrs. BP, following the dating of the initiation of peat accumulation at this time in the swales between the ridges at Broomhill. Jones (1953) believed that Lydd was situated on the original gravel barrier and was inhabited by fisherfolk, and has found remains from Roman times. Similarly, Needham (1988) has demonstrated that Early Bronze Age axes have been discovered near Lydd, providing another minimum age for this barrier's emplacement in this location.

The chronology of evolution of this coastline is easy to determine at the eastern end as there is cartographic evidence of shoreline changes. However, this evidence diminishes westward (Plater and Long, 1995). Therefore, it is at the western extreme of the gravel foreland that debate exists as to when deposition of the initial gravel barrier began. Eddison (1983a) proposes that the deposition of the Midrips occurred after 3305-3226 cal. yrs. BP, and Greensmith and Gutmanis (1990) are in broad agreement with this. Also Lewis and Balchin (1940) have suggested that the gravel represents deposition in the period from 2750-4040 cal. yrs. BP.

Material for the dating of barrier formation is mostly difficult to obtain, or may bear only an equivocal age relationship to the barrier's formation (Greensmith and Gutmanis, 1990; Long *et al.*, 1996). Consequently, many studies have attempted to determine the age of the barrier from the proxy record preserved in the back-barrier sediments, e.g. Tooley and Switsur (1988), Waller *et al.* (1988) and Long and Innes (1993).

2.1.2.5 The Character of the Initial Barrier and the Dungeness Foreland.

As outlined previously, the protective gravel barrier of the Dungeness foreland has played an essential part in the evolution of Romney Marsh (Eddison, 1983a&c). The most comprehensive survey of the gravel beaches of Dungeness foreland was made by Lewis (1932) and Lewis and Balchin (1940). More recently, however, Eddison (1983b) has reviewed the evolution of the barrier beaches along this coastline. Further stratigraphic research on the Dungeness gravel has been completed by Tooley and Switsur (1988), Long and Fox (1988), Greensmith and Gutmanis (1990), Plater (1991), Plater and Long (1995) and Long and Hughes (1995).

The gravel of Dungeness is unique due to the number of gravel ridges; each ridge apparently representing a beach formed when storm conditions coincide with a high tide (Eddison, 1983b). Eddison has stated that 500 gravel ridges occur within a 10km strip of the coast. Such a rapid repetition of ridges could only be possible with a high supply of gravel (Eddison, 1983a).

The south coast of Dungeness is made up of groups of gravel ridges that represent past shorelines. These areas are illustrated in figure 2.3 and are: The Midrips, The Forelands, Holmstone, Lydd Beach, West Ripe and Denge Beaches (Lewis and Balchin, 1940; Long and Hughes, 1995). Drew (1864) noted that each ridge to the west is older than the one before.

It is clearly shown by Eddison (1983a) (figure 2.3) that the orientation of each of the groups of ridges (such as The Forelands), is 10° nearer to the north than the previous group, and suggests that this rotation began somewhere in Rye Bay south-west of The Midrips. Eddison (*op. cit.*) has proposed that this reflects the origin of the ness of Dungeness. The ness shape is first observed east of Galloways Lookout (Eddison, 1983a; Long and Fox, 1988; Plater and Long, 1995). Figure 2.3 demonstrates that the ness shape may well have occurred earlier (further westward) but, as the coast has been eroded back, any evidence has been removed.

Steers (1964) proposed that the erosion of the older beaches occurred as the River Rother began to flow out at Rye. Thus, gravel was dislocated from the Fairlight coast and so the Dungeness beach was able to swing around to the form the coastline it has today. Lewis (1932, 1937) suggested that gravel starvation (reduced gravel supply) resulted in erosion of the earlier (westward) ridges and their realignment more parallel to the tides. Gallois and

Edmunds (1965) suggested that erosion of the gravel and a high sea-level caused a bend in the Rye Bay and, with a river at Rye preventing gravel movement east, the coastline retreated to face the predominant south and south-westerly waves.

The origin of the shape of the ness has been a matter of some debate. Redman (1854) and Wheeler (1928) proposed that the bend of Dungeness was due to the mouth of the Rother, then at New Romney, holding up sediment. This was discounted by Lewis (1937), who demonstrated that although the Rother mouth moved to Rye, the ness still grew eastwards. Gilbert (1933) described Dungeness as 'one of the most remarkable gravel promontories in the world', and that British sailors have attached its name to similar formations at Puget Sound and the southern shore of Patagonia. It was proposed by Gilbert (*op. cit.*) that the ness grew as a shoreline at such an angle to the dominant waves as to arrest the gravel. Lewis (1932) believed that as the bend of the ness became sharper, longshore drift along the leeward side was lessened and so large supplies of gravel accumulated around the point, which was built up into the ridges and, therefore, the ness advanced seawards. This mode of evolution was also proposed by Gallois and Edmunds (1965).

Eddison (1983a) suggested that the origin of the ness lay in ridge formation during periods of storminess when ridges would be formed in quick succession. Therefore, more gravel accumulated in numerous shorter ridges. Eddison

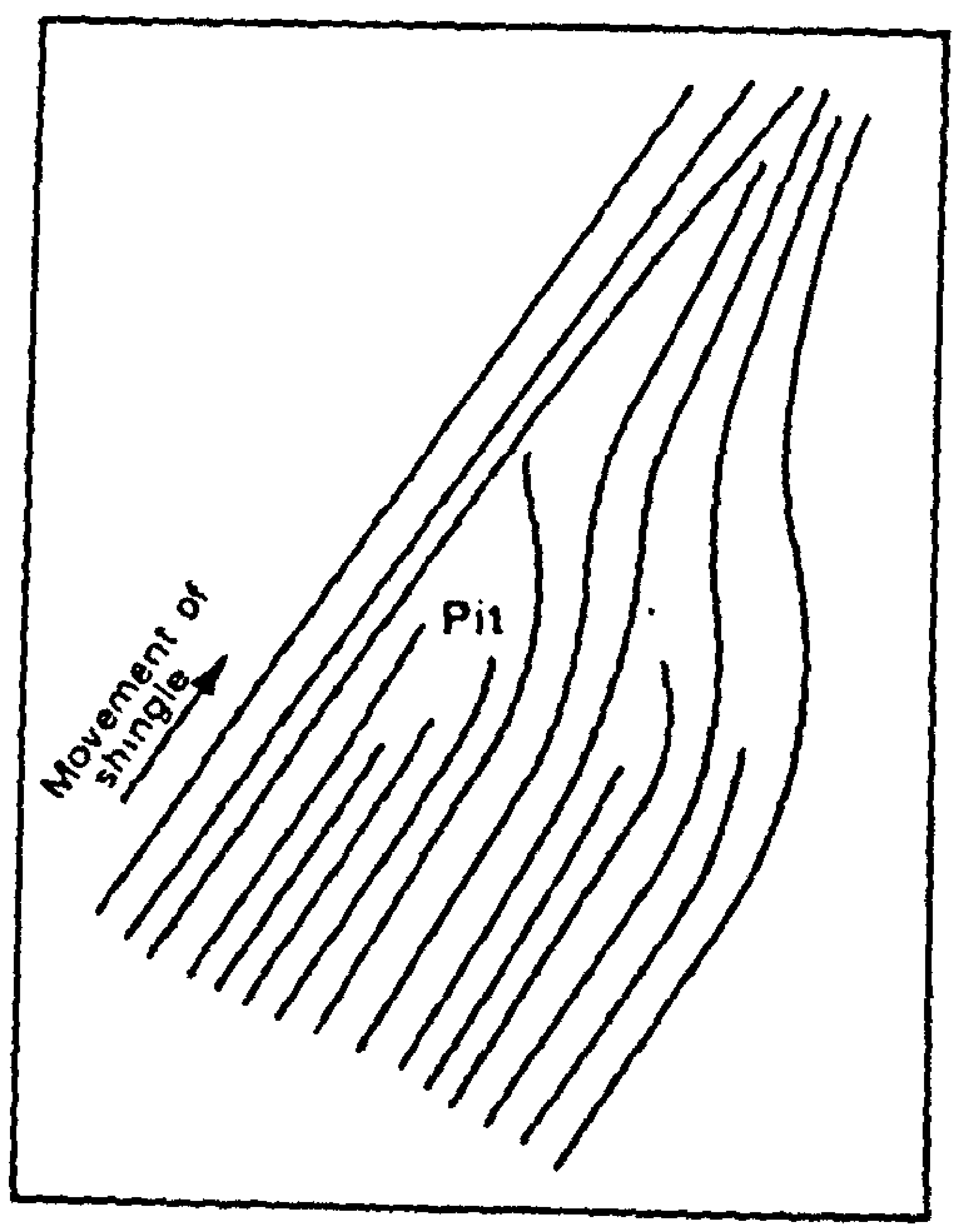


Figure 2.4: Postulated origin of the characteristic ness shape.

Source: Eddison (1983a).

suggests that the 'open pits' demonstrate that this has happened (figure 2.4). Later ridges would then have to curve around the shorter ones, and it is this that Eddison (*op. cit.*) believes is the origin of the bend of the ness. Thus, a curved shoreline occurs around open pits, such as Wickmaryholme Pit, sharpening the pre-existing ness.

It can be seen, therefore, that a wealth of research and consequent debate exists over the evolution of the Dungeness gravel complex. However, as Greensmith and Gutmanis (1990) illustrate, very little investigation of the sub-surface Holocene deposits has occurred, partly as they are nowhere exposed. Waller *et al*. (1988) have investigated the logs of boreholes reaching -12m O.D at the site of the 'B' power station. Hey (1967) observed excavations and boreholes to a depth of -14m O.D., demonstrating that the beach gravel rested on a shelly sand. Greensmith and Gutmanis (1990) have also used deep borehole data from the power station reaching depths of -40m O.D. in unlithified sediments. From these records, they have produced a depositional model of the Dungeness area and have reached some of the conclusions discussed above. In addition, Long *et al*. (1996) have investigated deep boreholes in the Rye area and have interpreted these sediments as indicating that a drift-aligned barrier became deposited across Rye Bay.

2.1.3 Sea-Level Changes

Considerable changes in sea-level have occurred during the period of formation of Romney Marsh and Dungeness foreland. Inevitably, such changes will have influenced the evolution of this area, especially the back-barrier and inter-ridge environments. Indeed, Tooley and Switsur (1988) stated that

> "...conclusions about the direction and rate of vertical movements along the south-east coast of England are crucial to understanding the mode of formation and the history of the coastal landforms and tidal flat, lagoonal, gravel and perimarine sedimentation in the Romney Marsh area, and the river valleys that debouch here".

2.1.3.1 Sea-level Changes in South-east England.

It was noted by Flemming (1982), Shennan (1983) and Plater (1992) that the quantity and quality of sea-level data for relative sea-level changes in south-east England are poor. However, some important and detailed studies have now been completed, e.g. Devoy (1982), Long (1991), Long and Tooley (1995) and Long and Innes (1993). Sea-level data were collated from the south-east of England by Devoy (1982), who took this to exterd from Ingoldmells in Lincolnshire to Poole in Dorset. From his dataset, 55 samples were believed to provide reliable ^{14}C dates and altitude evidence for the movement of mean high water spring tide (MHWST). Statistical analysis of these data suggested that a smoothly rising exponential rate and direction of sea-level rise did not provide the best solution to Holocene sea-level change, and that alterations in the rate and direction of rise had occurred. Devoy (1982) demonstrated that since ca. 9990-9894 cal. yrs. BP sea-level has risen from -30.00m OD to a maximum of +0.50m OD by ca. 1272-1241 cal. yrs. BP.

Long and Innes (1993) considered regional sea-level changes in south-east England. They demonstrated that, in general, a curvilinear increase in mean sea-level (MSL) is apparent (figure 2.5). Also Long and Tooley (1995) added 6 new sea-level index points (SLIPs) from Hampshire to a database of SLIPs from Sussex (Jennings and Smyth, 1987; Smyth and Jennings, 1988), Romney Marsh (Tooley and Switsur, 1988; Waller *et al*., 1988; Tooley, 1988; Long and Innes, 1993), the East Kent Fens (Long, 1992) and the Thames Estuary (Devoy, 1977; 1979; 1982), and proposed a model for sea-level change during the Holocene for south-east England. In this record, Long and Tooley (1995) identified marked spatial variability in the distribution of sites, i.e. 14 dates from both the East Kent Fens and the Thames Estuary; 15 from Romney Marsh; 8 from Sussex and 6 from Hampshire, and also temporal variability of SLIPs' with the only dates older than 7000 cal. yrs. BP from the Thames Estuary and a distinct lack of data from the last 2000 cal. yrs. BP. Despite these problems, Long and Tooley (1995) identified that MSL appears to have risen from ca. -27.5m OD at 9000 cal. yrs. BP to ca. -8.0m OD at 6000-6500

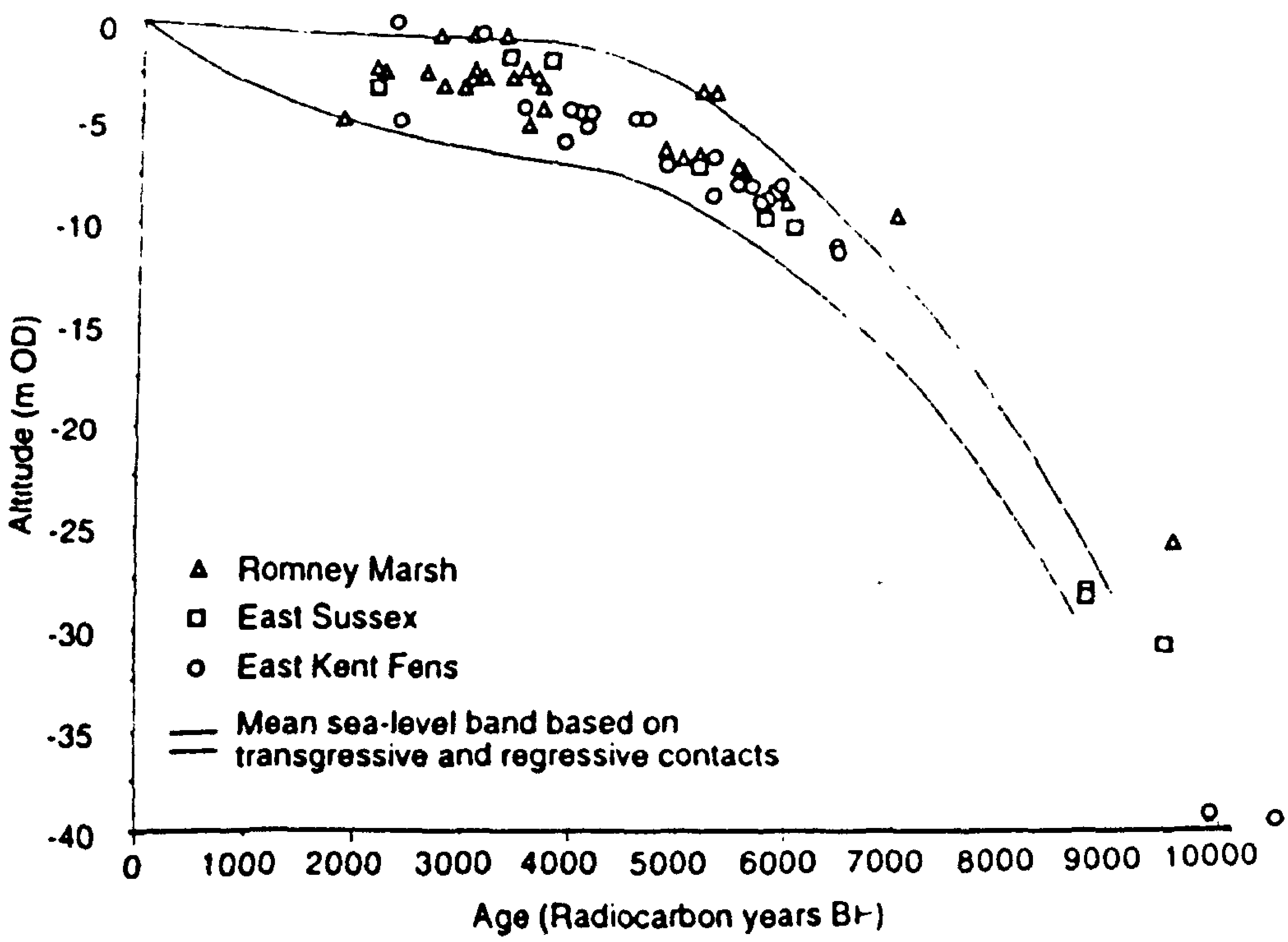


Figure 2.5: Time / altitude plot of transgressive and regressive contacts from Romney Marsh, East Sussex and the East Kent Fens.

Source: Long and Innes (1993).

cal. yrs. BP. From this time, the rate of sea-level rise slowed in the Thames Estuary and in the East Kent Fens. However, in Hampshire, Sussex and Romney Marsh Long and Tooley (1995) demonstrated that sea-level continued to rise a further 2.5m to ca. -2.5m OD by 4000 cal. yrs. BP. Between ca. 4000 and 2000 cal. yrs. BP, the rate of sea-level change was minimal in all areas of the south-east of England (Long and Tooley, 1995). However, since 2000 cal. yrs. BP mean sea-level (MSL) must have risen at a faster rate in the Thames Estuary and the East Kent Fens (where a sea-level of up to 5.0m has been recorded (Devoy, 1982 and Long, 1992)) than in Romney Marsh, Sussex and Hampshire (where a rise of between 1.5 and 3.0m has occurred). This disparity in late Holocene sea-level rise has been identified by both Long and Innes (1993) and Long and Tooley (1995).

Consequently, the East Kent Fens and the Thames Estuary have experienced a broadly similar pattern of sea-level movement, although with some slight differences (Long and Shennan, 1993). Equally, Romney Marsh, Sussex and Hampshire appear to have a broadly similar pattern of Holocene sea-level movement in the last 5000 cal. yrs. BP. The sea-level records of the two groups differ significantly during two critical time periods (Long, 1992; Long and Innes, 1993; Long and Tooley, 1995), between 5000-4000 cal. yrs. BP and again between 2000-0 cal. yrs. BP.

Long (1992) and Long and Innes (1993) proposed that the observed differences in the sea-level record may be due to differential crustal movement between the two geographical areas. Long and Tooley (1995) state that the line delimiting the two areas of apparently different crustal movement appears to occur in a relatively narrow area between the East Kent Fens and Romney Marsh. The Variscan Thrust Front (figure 2.6) was identified by Wallace (1968; 1981) and Hamblin *et al.* (1992) to intercept the coastline between the East Kent Fens and Romney Marsh. This Variscan Front has been demonstrated (Meissner *et al.* 1981) to still be tectonically active. Consequently, Long and Tooley (1995) suggest that the Variscan Front may be one possible source for the proposed differential crustal movement.

2.1.3.2 Early Considerations of Sea-level Change in Romney Marsh.

The fact that relative sea-level has changed through time in the Romney Marsh area was noted by a number of early workers in the area. For example, Dowker (1897) noted that trees in the peat horizons were evidence of a change in the height of the land relative to the sea. Homan (1938) also identified that ancient trees were found at a lower level than when they were growing. Also Gilbert (1933) stated that trees could not have grown on sand unless 'the bay' had risen above the level of the tides, and suggested that an uplift of 7.50-9.00m would have been needed. Edmunds (1954) also proposed that an uplift of 7.50m would be necessary to enable forest growth. J.K. Dubey (in

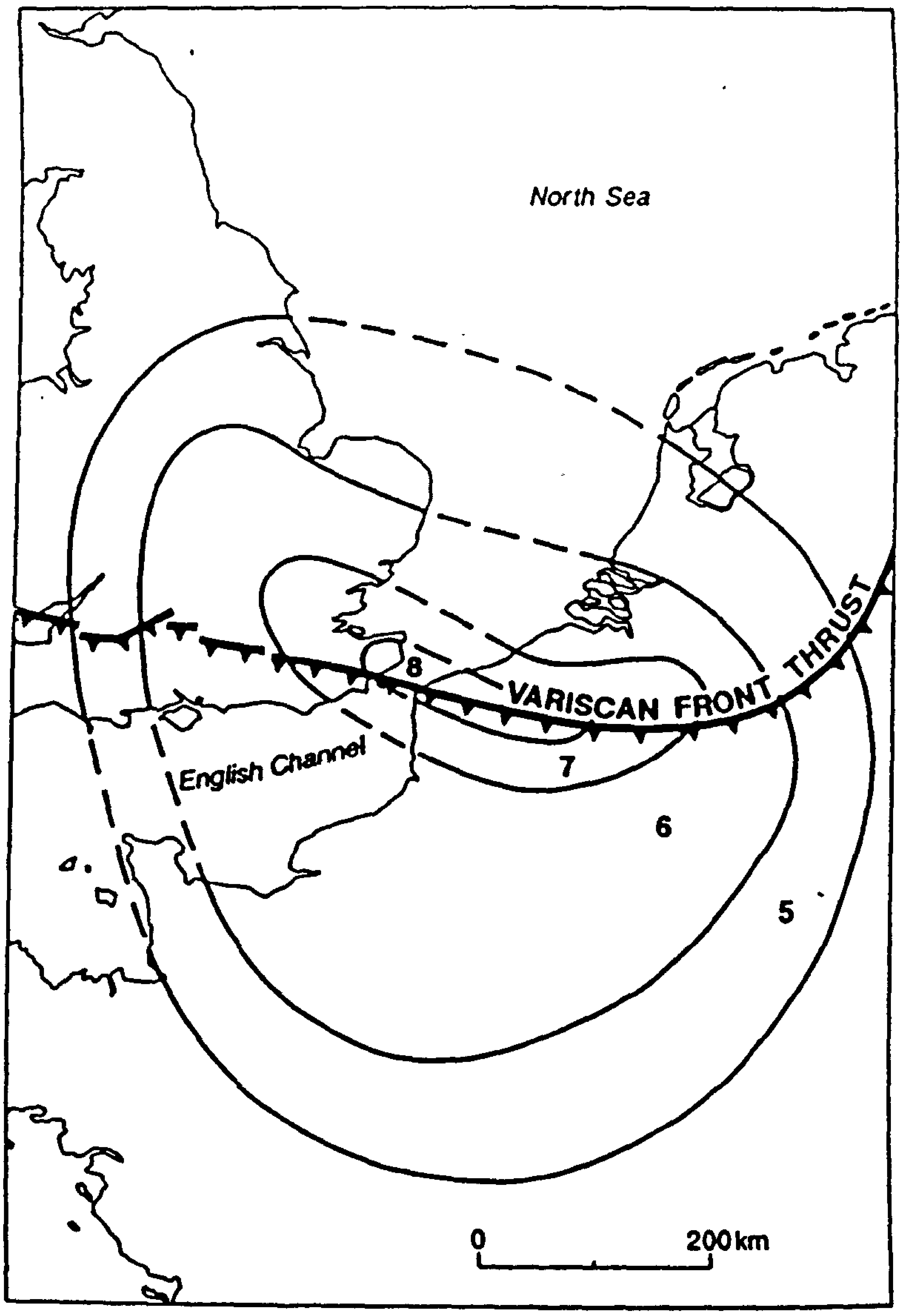
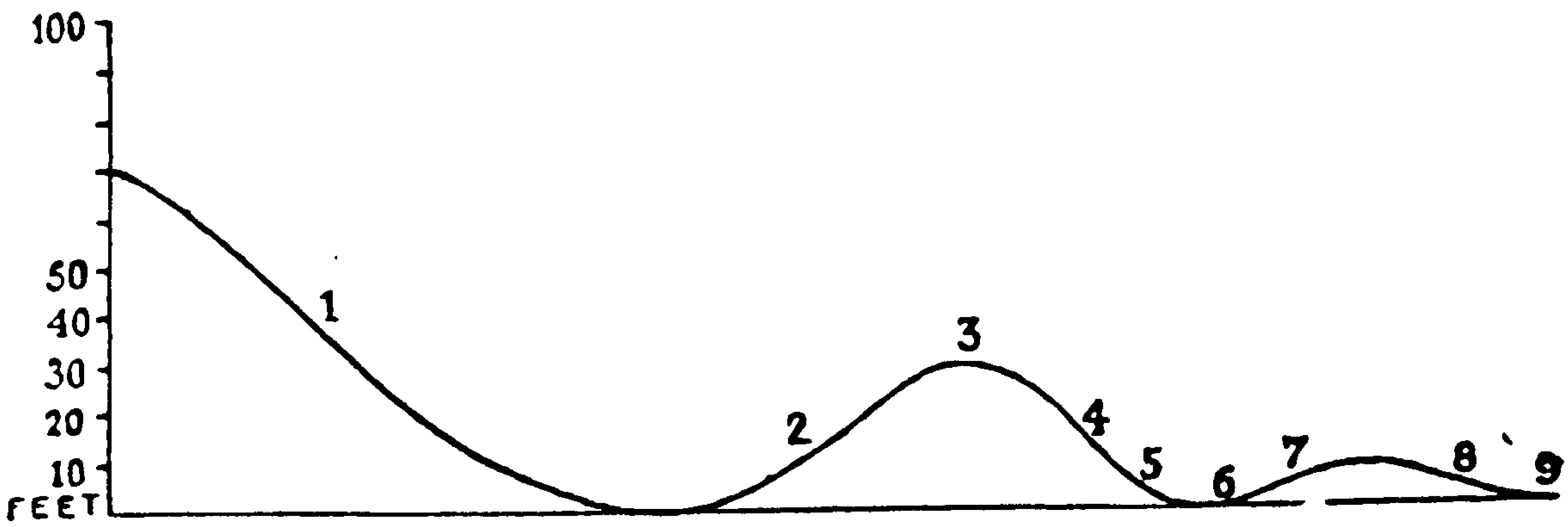


Figure 2.6: The position of the Variscan Front Thrust. Source: Long and Tooley (1995), after Hamblin et *al.* (1993).

Burrows *et al.*, 1949) stated that Romney Marsh has undergone uplift and subsidence and has been alternately sea, grassland and woodland. Indeed, the moor-logs in the Appledore Dowels were cited as evidence of this alternating sequence. Dubey (*op. cit.*) identified that, due to the existence of sand underneath the peat, there was initially sea covering the area. This was followed by uplift, leading to peat growth, and then subsidence during which marine conditions again dominated.

Spatial variability in relative sea-level change was also identified by Homan (1938) and Jolly (1939) from geomorphological evidence, such as raised shorelines in Scotland and buried forests in southern England, and also archaeological evidence. Both posed the question 'is it Scotland that is rising or south-east England that is sinking?'. The temporal variability of sea-level change has also been considered. Dowker (1897), for example, stated that it was "commonly supposed" that the land level was lower than at present during the Roman times. Jolly (1939), also identified that land in the Thames Valley had sunk ca. 5.00m compared to tidal levels in Roman times.

Gilbert (1933) provided the earliest account of changes in relative sea-level during the evolution of Romney Marsh, and produced a diagram illustrating the proposed changes (figure 2.7). He suggested that the initial land depression was the last in a series of earth movements, and stated that "this latest depression, rather than coming to an abrupt close, should die out in a



1: Neolithic depression.

2: Forest uplift.

3: Forest growth.

4: Forest depression.

5: Peat bed.

5 and 6: Deposition of alluvium, marsh beaches and Hythe promontory.

7: Post-forest uplift.

8: Post-forest depression.

9: Last stages of Dungeness.

Figure 2.7: The first suggested model of sea-level change in Romney Marsh.

Source: Gilbert (1993).

diminuendo of minor oscillations". Gilbert suggested that, following a period of land depression, a period uplift was experienced; "the uprise had brought about a great transformation; the erstwhile bay had become dry land". He termed this uplift the "forest uplift", during which time the forest beds were growing. Land depression eventually resumed, which brought about tidal invasion through the river mouths and the development of a lagoon, during which silt and clay kept pace with the depression and salt-loving plants grew on the marsh. Again Gilbert (1933) proposed that slight uplift of the land followed, which aided the initiation of Dungeness foreland. This was followed by land depression which caused the destruction of Old Winchelsea. This led to relative sea-levels similar to those existing at the present, and the last stages of the evolution of Dungeness. Gilbert (1933), stated that the Rhee Wall could not have done its job, regardless of whether its purpose was to drain or to protect, if the sea-level had not been up to 2.40m lower than at present. Gilbert (1933) proposed that it is these gradual oscillations, which allowed the marsh sedimentation to keep pace with land depression, that account for Romney Marsh's "wondrous fertility".

Clearly some uncertainties exist in Gilbert's' hypothesis. For example, there is the absence of a clear chronology, but it is at least an important attempt to reconstruct the relative sea-level changes that occurred in the Romney Marsh area during its formation.

2.1.3.3 A Sea-Level Record From Altitudes of the Gravel Ridges?

Research has been completed by a number of workers as to whether any proxy record of Holocene sea-level change is preserved in the gravel ridges of the Dungeness complex. Drew (1864) noted that the level of older ridges was approximately ca. 1.80m below modern shore lines, and proposed that this was due to a change in tidal range. Dowker (1897), however, believed that the change of level was due to a change in the relative land and sea level.

The most comprehensive investigation of this sea-level control was carried out by Lewis and Balchin (1940). They suggested that sustained changes in the altitude of the ridges are a reflection of changes in relative sea-level. As a result of this work, they proposed a chronology for ridge construction and changes in relative sea-level since the Neolithic period. To examine the relationship between sea-level and the altitude of the gravel ridges, Lewis and Balchin (1940) realised that the height of a single storm ridge indicated very little. However, if 10 or more ridges in a group were used, the evidence might indicate a real change of relative sea-level. They also believed that certain ridges could be attributed to particular events. For example, the great storm of AD 1287 was believed to be responsible for the highest of the gravel fulls at +6.2m OD. Using consistent changes in altitude to reflect changes in sea-level, and with a chronology for beach development based on cartographic and historical evidence, Lewis and

Balchin inferred the following changes in sea-level relative to the present day, as summarised by Plater and Long (1995).

Present day (1940)	=	0m
15^{th} century	=	-0.30m
13^{th} century	=	0m
8^{th} century	=	-0.30m
Roman Period	=	-1.50 to -1.80m

Table 2.1 : Lewis and Balchin's (1940) inferred sea-level changes, as summarised by Plater and Long (1995).

Lewis and Balchin (1940) found that a ca. 1.80m rise in sea-level from the fifteenth century to today had occurred. Plater and Long (1995) appear to have misinterpreted this work inferring a rise of 0.30m (Table 2.1).

Since the work of Lewis and Balchin (1940), a number of authors have questioned their conclusions. Long and Fox (1988), Plater and Long (1995) and Long and Hughes (1995) suggest that a sea-level record based on gravel altitudes is dubious. Long and Fox (1988) demonstrated how a number of factors other than sea-level can affect the altitude of gravel ridges. Changes in the shape of the coast, both up- and down-drift (for example, the changes of the location of the mouth of the River Rother), would influence factors such as the supply of sediment. Also tidal amplitude would play an important role in the development of gravel ridges. With a greater tidal amplitude, higher ridges and lower troughs would be possible (Long and Fox, 1988). In addition, the sediment supply would also have a strong bearing on the size of ridge that could accumulate.

Long and Fox (1988) also suggested that over the period of evolution of the gravel complex, it is possible that the orientation of the storms may have altered. In addition, humans were present throughout the evolution of this coastline and so may have had an impact on its morphology. Plater and Long (1995) also illustrate this complexity, stating that the altitude of a gravel ridge is a function of meteorology, tidal dynamics, sediment supply, and mean sea-level, and propose that changes in the altitude of gravel deposits are a product of interaction between long term changes in sea-level and short term changes in the magnitude of storm events.

As an alternative to gravel ridge altitude, Lewis and Balchin (1940) also proposed that the alternating areas of marsh and gravel between Jury's Gut and Galloways Lookout illustrated a record of sea-level change, with gravel being deposited when sea-level was high and marsh deposited when sea-level was low. However, Long and Innes (1993) and Long and Hughes (1995) believe that it is more likely that the alternating areas of gravel and marsh were caused by changes in storm incidence and gravel supply.

Long and Hughes (1995) suggest that the altitude of any gravel ridge varies too much to be truly indicative of sea-level. In addition, through comparison with time / altitude graphs, they show that sea-levels between 4507-4416 and cal. yrs. BP were low. In addition, little or no evidence was found of a change in the altitude of the water table at sites inland of the foreland, *i.e.* at Wickmaryholme Pit, which would be a direct consequence of change in the local sea-level (Long and Innes 1993; Tooley and Switsur 1988). Long and Hughes (1995) suggest that changes in gravel altitude could be due to increased or decreased extension of gravel barriers with changes in wave competence, which may be linked to changes in storm magnitude. The successive areas of gravel and marsh across the southern area of Romney Marsh are suggested by Long and Hughes (1995) to be more likely to be linked to changes in the frequency and magnitude of storm incidence than to sea-level change.

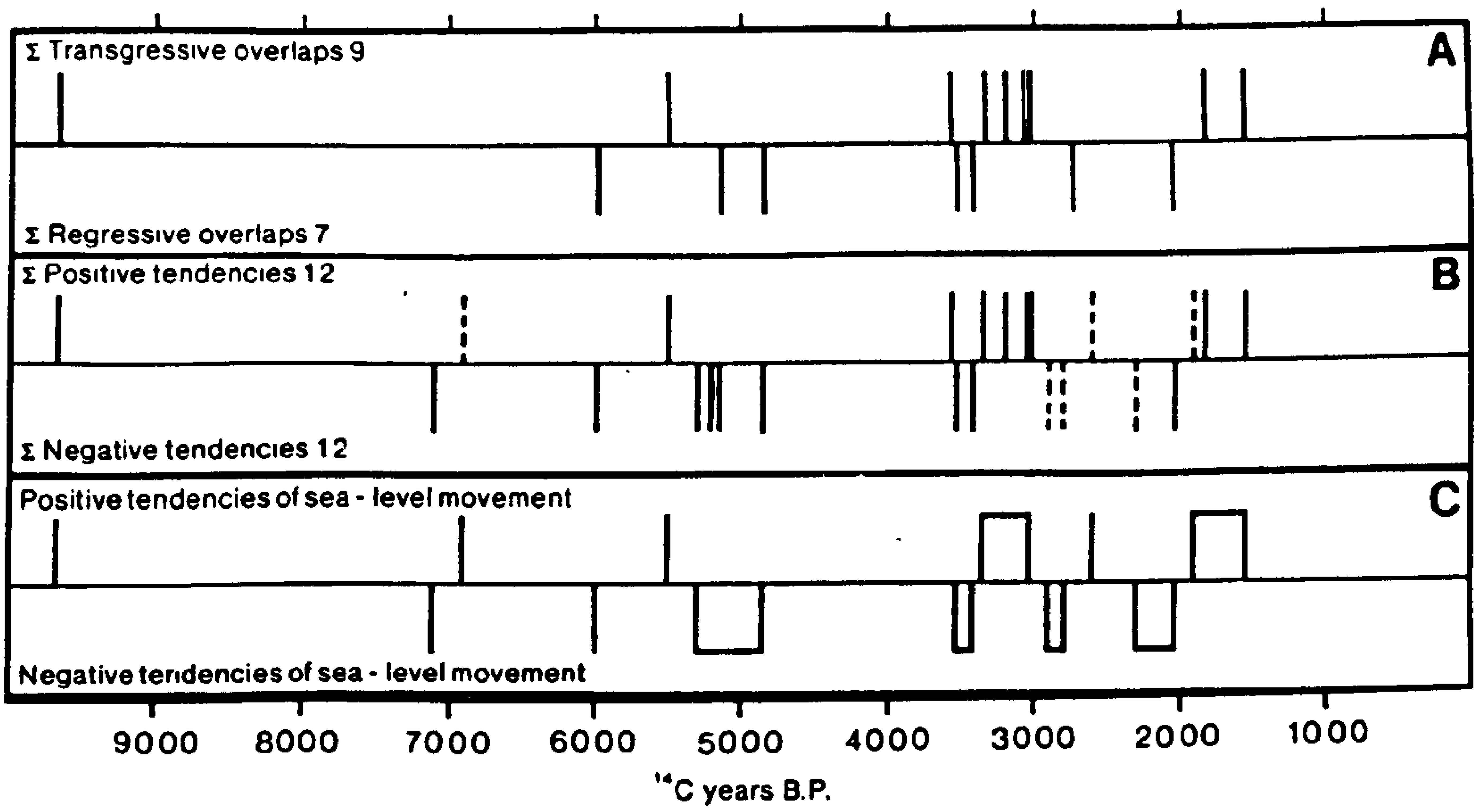
It appears that the altitudes of the gravel ridges may be only partially influenced by sea-level, along with a number of inter-related processes, and, thus, no unequivocal record of sea-level change can be resolved from the altitudes of the gravel ridges of the Dungeness foreland.

2.1.3.4 Investigations of Sea-Level Changes from Back-Barrier Sediments in Romney Marsh.

With no conclusive record of sea-level change being resolved from the altitudes of the gravel ridges many studies have utilised the marsh sediments to determine changes in sea-level during the Holocene. Tooley (1995) noted a maximum age for changes in sea-level in the Romney Marsh region from a rare peat at Tilling Green, Rye at 10559-9958 cal. yrs. BP (Welin *et al.*, 1974). However, no supporting biostratigraphic data are available and, therefore, no altitudinal relationship to former sea-level can be made (Long *et al.*, 1996). Tooley (1995) notes that similar deposits have been recorded on other European / southern North Sea coasts at the commencement of the Holocene.

Changes in relative sea-level were investigated by Tooley and Switsur (1988) at two locations on Romney Marsh; one at Horsemarsh Sewer in the north-west of the marsh, and the other on the southern edge of the marsh at Broomhill. Data from these two locations yielded six SLIPs; three transgressive and three regressive overlaps. To consider these data in a regional framework, the site's chronologies and palaeogeographies were compared to establish whether any correlation existed. Utilising these data and those of other authors, *i.e.* Callow *et al.* (1964; 1966), Welin *et al.* (1971; 1972) and Waller *et al.* (1988), figure 2.8 was produced as a partial chronology of sea-level tendencies.

The data from Tooley and Switsur (1988), along with those of the other authors outlined above, indicate a positive tendency (an increase in the marine influence) of sea-level over an extensive area of Romney Marsh and up into the major adjoining valleys at around 3000-3500 cal.



A: A distribution of 16 radiocarbon dates on transgressive or regressive overlaps in Dungeness, Romney Marsh, Walland Marsh and the adjacent river valleys.

B: A distribution of 24 SLIPs showing positive and negative tendencies.

C: Provisional partial chronology of tendencies of sea-level movement.

Figure 2.8: An exploratory model of sea-level investigation.

Source Tooley and Switsur (1988).

yrs. BP. The index points are from a number of contrasting palaeoenvironments as described by Hageman (1969) and Tooley (1978; 1985b).

Tooley and Switsur (1988) suggested that the temporal concurrence of their index points from contrasting palaeoenvironments, demonstrated regional processes. Tooley and Switsur proposed that their results necessitated a re-examination of the conclusions of Jennings and Smyth (1982; 1985) who believed that the deposition of similar environments to the west, in East Sussex, were best explained by local processes, particularly the emplacement and removal of barriers. A second cluster of dates were also identified around 5742-5718 cal. yrs. BP (figure 2.8) where index points at Horsemarsh Sewer, Wittersham Level and Pett Level *record regressive overlaps, i.e. reduction* of the marine influence. It was acknowledged that the data set was small and the periods of positive and negative tendency sea-level movement were provisional and, in part, speculative, but the between-site correlation suggested the existence of an 'area'. An 'area' was defined by Shennan (1982, 1983a&b) to be when between-site correlation exists, a 'site' being a tract of ground over which a single sea-level index point is believed to be representative.

A comprehensive review of the recorded changes in sea-level is presented by Long and Innes (1993). They suggest that some of the dates utilised by Tooley and Switsur (1988) have no litho- or bio-stratigraphic context, and, therefore, provide an equivocal sea-level indicator.

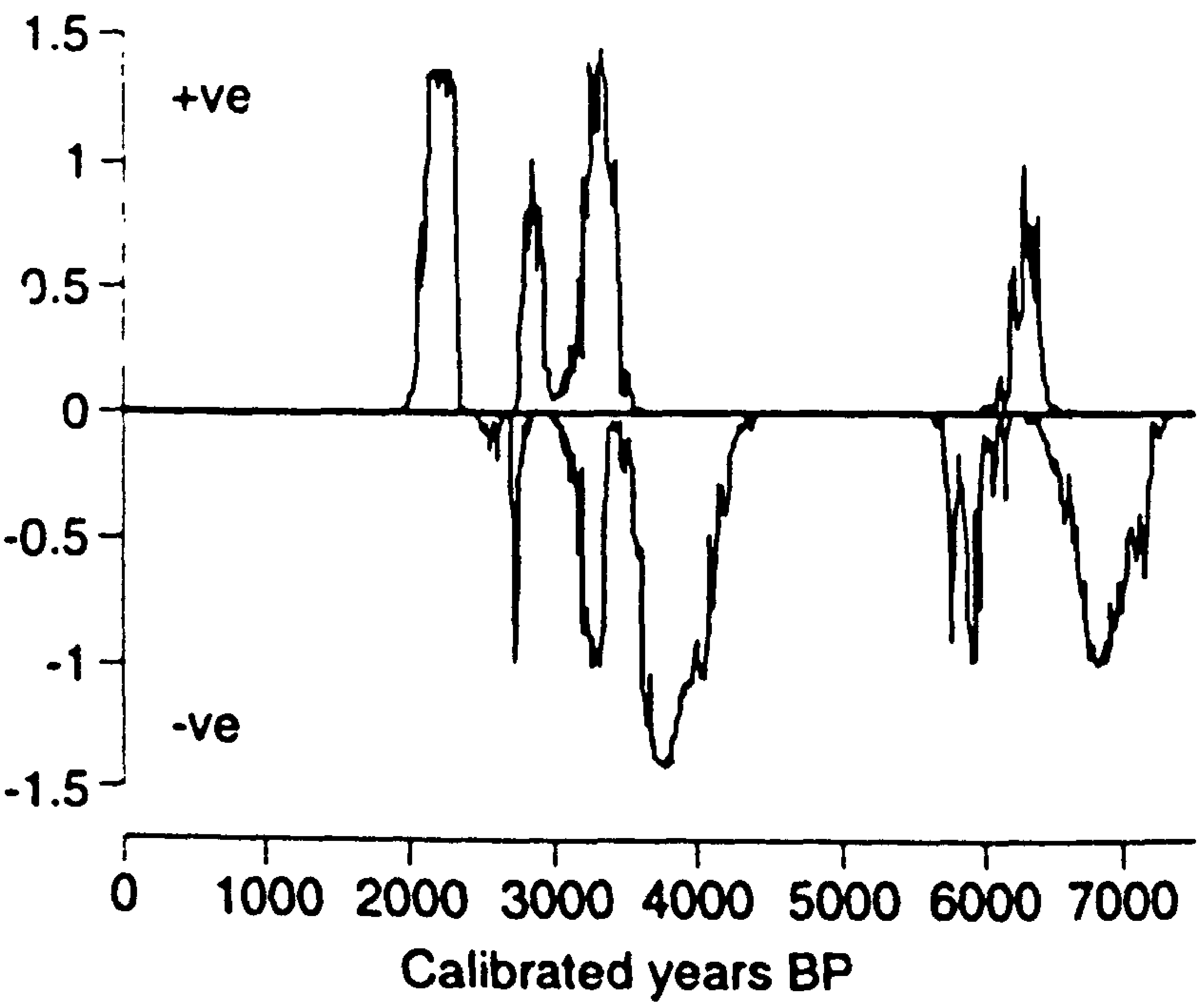


Figure 2.9: Tendencies of sea-level movements from Romney Marsh.

Source: Long and Innes (1993).

Conversely, Long and Innes (1993) only utilised sites for which a clear sea-level tendency had been determined, *i.e.* those of Tooley and Switsur (1988), Waller *et al.* (1988), Tooley (1990b) and also data from their own study at Midley Church Bank. From these data, they were able to provide a revised chronology for sea-level changes in the Romney Marsh region (figure 2.9). Figure 2.9 (Long and Innes, 1993) illustrates that two sites (Brede Bridge and Horsemarsh Sewer) provide a record of negative tendency (recording the onset of peat accumulation) with an age of older than 4500 cal. yrs. BP. Conversely, only one age of older than 4500 cal. yrs. BP is recorded for a positive tendency, marking the return of marine conditions to Horsemarsh Sewer, prior to the main period of peat accumulation.

A group of ages was identified (figure 2.9) between 4500-3700 cal. yrs. BP; these are from Broomhill (Tooley and Switsur, 1988) and Midley Church Bank (Long and Innes, 1993). Despite the proximity of these two sites, significant disparities exist between their tendency chronologies, and, as Long and Innes suggest, are at odds with the stratigraphic record. In the period between the initiation of peat at Midley Church Bank, *i.e.* core 10B' 4143-3896 cal. yrs. BP and core 6B 3362-3211 cal. yrs. BP, two positive tendencies are recorded from Broomhill which are not represented in the biostratigraphy of the Midley Church Bank. This may relate to low rates of sedimentation, combined with a low rate of sea-level rise causing varied sedimentary responses to subtle changes in the rate of sea-level change, as illustrated from records of the rates of peat accumulation in the back-marsh areas of Romney Marsh (Waller *et al.*, 1988).

In relating the chronology of sea-level tendencies to the deposition of marsh sediments, Long and Innes (1993) illustrated that between 6850 and 4050 cal. yrs. BP, a net rise in sea-level of ca. 6.0m OD was experienced in the Romney Marsh region. They illustrate that this is equivalent to a rise of 0.21cm cal. a^{-1} in the valleys, where peat flourished at this time, *i.e.* at Brede Bridge, Old Place and Horsemarsh Sewer. At these sites Waller *et al.* (1988) have proved a rate of peat accumulation of 0.20cm cal. a^{-1}, which is comparable to the rate of sea-level rise. During this time marine conditions dominated on the marsh. However, Long and Innes suggest that following the regional slowing of the rate of sea-level rise, *i.e.* after ca. 4500 cal. yrs. BP, peat-forming communities spread across the marsh. The slowing in the rates of sea-level rise are also illustrated by a slowing in the rate of peat accumulation. Long and Innes (1993) demonstrate that after ca. 1950 cal. yrs. BP, a rise in sea-level of between ca. 1.0 and 2.5m was recorded up to the present day, and they state that it is somewhat surprising that the peat-forming communities that had previously been able to cope with rising sea-level were now inundated by marine conditions across Romney Marsh. Indeed, only from Pannel Bridge (Waller, 1987) was peat accumulation able to keep pace with sea-level rise.

Additional studies of time / altitude changes in the Romney Marsh region have been completed by Long and Innes (1995a), Long and Hughes (1995) and Long *et al*. (1996). The results of these studies essentially re-affirm the trends proposed by Long and Innes (1993).

Plater and Long (1995) suggested that a good deal is known about changes in relative sea-level between 10,000 and 4,000 yrs. BP, yet little is known about the period from 4,000 yrs. BP to the present day. The areas of Romney Marsh that could provide a vital insight to this period of time are Denge Marsh and Denge Beach. However, Plater and Long (*op. cit.*) state that the stratigraphic and micropalaeontological techniques applied to sea-level study, *e.g.* Jelgersma (1961), Tooley (1978), Shennan (1982), are of little use at Denge Marsh and Denge Beach due to the scarcity of intercalated marine and terrestrial sediments. Therefore, Plater and Long (1995) applied a developmental approach in which the morphological characteristics of successive recurve populations were investigated in the context of sea-level change and storm incidence during the evolution of Dungeness.

It was proposed that the 1.00m increase in ridge altitude eastward at Denge Beach may be interpreted as long term sea-level change coupled with storm event magnitude between the Roman period and AD 750. They also believed that along-profile morphology can account for much of the altitudinal variability in the five observed ridge populations (Lewis and Balchin, 1940). The investigations carried out by Plater and Long (1995) also revealed that sea-level change in the late Holocene was of insufficient magnitude to cause any large scale change in marsh stratigraphy, environmental salinity or sediment provenance.

It can be seen, therefore, that detailed analysis of changes in sea-level of the back-barrier sediments of Romney Marsh has provided a detailed proxy record of changes in sea-level and coastal evolution during the Holocene. The dated SLIPs obtained from the back-barrier sediments of Romney Marsh thus far are predominantly from the back- to mid-marsh areas, *i.e.* the Brede Valley, Horsemarsh Sewer, Rye, Brookland and Midley Church Bank. As yet fore-marsh sites have received relatively little attention with the exception of Broomhill (Tooley and Switsur, 1988).

2.1.4 Marsh Sediments : Introduction.

The most comprehensive study of the marsh sediments is that completed by Green (1968), who proposed a stratigraphic sequence for Romney Marsh with sands overlain by blue-grey clay, the main marsh peat and finally the younger alluvium. Since this broad review of the sediments of Romney Marsh, a number of more detailed and site-specific studies have been completed, some of which have encountered the typical sedimentary sequence described by Green (1968), *i.e.* Field (1983), Everett (1985) and Waller et *al.* (1988). However, stratigraphic investigations completed by Tooley and Switsur (1988) at Broomhill and Horsemarsh Sewer, and also Innes and Long (1992) and Long and Innes (1993; 1995b) at Midley Church bank have demonstrated that stratigraphic variability exists across areas of Romney Marsh.

Long and Innes (1995a) completed a transect that stratigraphically linked a number of site-specific studies. In this study, it was demonstrated that distinct stratigraphic sequences are recorded in the back-, mid- and fore-marshes. For example, Long and Innes (*op. cit.*) identified a single peat bed in the fore-marsh sites between altitudes of 0.00m OD and +1.50m OD, whereas in the mid-marsh, this peat unit tended to be thicker and occurred between altitudes of -2.00m OD and +0.50m OD. In the back-marsh, Long and Innes (1995a) found that most sites contained one main peat unit and at least one other thin peat unit. In addition, it was illustrated that variations in both the altitude and ages of the stratigraphic units occurred. For example, the altitude of the contact between the blue clay and the main marsh peat unit tends to rise from the edge of the back-barrier embayment towards the centre of the marsh, and also becomes younger.

2.1.4.1 The Marsh Sediments : The Lower Inorganic Sediments.

The palaeoenvironmental conditions existing during the deposition of the lower inorganic sediments have been studied in a number of sites, *i.e.* Tooley and Switsur (1988), Waller *et al.* (1988), Long and Innes (1995a) and Long *et al.* (1996). Problems have been encountered in investigating the basal sands due to the paucity of diatoms (Long and Innes, 1995a). However, those diatoms that were recorded tended to indicate that open tidal deposition prevailed.

The deposition of the overlying blue clay is better understood and has also been examined at a number of sites, *i.e.* Horsemarsh Sewer and Broomhill (Tooley and Switsur, 1988) and at Brookland (Long and Innes, 1995a). These studies have demonstrated that the deposition of this sedimentary unit occurred on a mudflat, initially in the subtidal to intertidal zone and eventually becoming more brackish and supratidal as the organic content of the sediments increases at the upper contact with the peat unit.

2.1.4.2 The Marsh Sediments : The Organic Peat Unit.

The palaeoenvironmental conditions that existed throughout the period of peat accumulation have been investigated thoroughly at many sites across Romney Marsh. Consequently, a detailed summary of the palaeovegetational conditions can be presented. Through a significant period of time during the evolution of Romney Marsh, peat communities have dominated, both as a unit gradually spreading across the marsh and also in the more sheltered environments in the adjoining valleys to the west and north of Romney Marsh.

The main peat unit has been investigated in the valleys to the west of Romney Marsh by Waller *et al.* (1988) and Waller (1993; 1994a). The palaeovegetation in two of these valleys has been reconstructed (Pannel Bridge and Brede Bridge), and have been demonstrated to be relatively similar (Waller, 1993; 1994a). Waller has recorded that open wetland conditions with saltmarsh were superseded by stable alder carr conditions that prevailed for ca. 2000 years. Eventually, the alder carr was replaced with wetland conditions, with *Cyperaceae* and *Myrica* dominating before alder carr conditions returned. At the upper contact of the peat, Waller (1993; 1994a) demonstrated that marine conditions returned, with wetland and saltmarsh communities present.

In another study of a back-marsh site, Tooley and Switsur (1988) recorded the removal and the return of marine conditions in the vegetation succession at Horsemarsh Sewer. Here, saltmarsh was replaced by freshwater reedswamp and eventually fen to oak woodland with some reedswamp conditions. However, an increase in marine conditions upward was traced by, first, the increase in the dominance of freshwater reedswamp and aquatic taxa and then saltmarsh at the upper, transgressive contact. Field (1983), in the Appledore Dowels, identified a *Phragmites* reedswamp that was replaced, in succession, by carr conditions, sedges, ferns, and alder. Eventually, the environment dried out further and a mature fen carr prevailed.

A further study of the organic sediments of the back-marsh has been completed by Long *et al*. (1996) in the Rye area. Here, too, the pollen records the removal of marine conditions with saltmarsh becoming reedswamp, and an increase in alder being indicative of an alder carr at, or near to, the site. In addition, some acidophilous vegetation became established during the accumulation of the peat after ca. 4500 cal. yrs. BP. Unfortunately, the upper contact of the peat unit had been truncated, and, thus, no indication of a gradual return of marine conditions was proven. In one core (core 33a), however, some increase in saltmarsh plants *i.e.* Chenopodiaceae, was demonstrated, suggesting that a gradual increase in the marine influence did occur and that the upper contact of some of the peat was eroded by some later event.

Further out into the marsh, in the mid-marsh at Brookland, Long and Innes (1995a) have investigated the palaeovegetational conditions from the main marsh peat unit. Here, Long and Innes, have recorded the removal (5040-4842 cal. yrs. BP) and subsequent return of marine conditions (1500-1278 cal. yrs. BP). Initially saltmarsh conditions prevailed, becoming a freshwater reedswamp, with grasses and aquatic plants. Upward, the vegetation became typical of a fen-carr community with alder and some willow and birch. A subsequent fall in the frequency of *Alnus*, and a rise in the frequency of Cyperaceae, may indicate an opening up of the alder community, due to a change in the trophic conditions, and, thus, drier more acid conditions. Alternatively, this change may reflect a rise in the water table, as an increase in the frequency of aquatic taxa was also recorded. Eventually, a clear rise in the water table is suggested, by the re-establishment of freshwater reedswamp conditions as the marine influence approached (ca. 3025 cal. yrs. BP). Indeed, a short-lived period of saltmarsh is recorded at the upper transgressive contact (Long and Innes, 1995a).

One characteristic that is present in all but one study of the palaeovegetation of Romney Marsh, is that the further away from the Wealden margin of the marsh, the lesser the amount of Wealden arboreal taxa recorded. However, in the study of Everett (1985) from Walland Marsh a relatively high presence of arboreal taxa was recorded. Equally, the high frequencies of *Gramineae*, which are characteristic of pollen studies in Romney Marsh (Tooley and Switsur, 1988 and Long and Innes, 1993), were not noted by Everett. The reason for this is not clear but the results are certainly anomalous.

An extensive reconstruction of the palaeovegetation at Midley Church Bank has been completed by Innes and Long (1992) and Long and Innes (1993; 1995b). Peat growth began at this site at 2341-2185 cal. yrs. BP (in core 10b) with saltmarsh conditions dominating and indicating a regressive contact. Following this, the marine influence was further removed allowing freshwater reedswamp and, eventually, oak and alder fen conditions to prevail whilst aquatic plants diminished. In a second core studied by Long and Innes (core 6B), peat initiation was recorded ca. 1000 years later (at 3362-3211 cal. yrs. BP) than at core 10B. The peat occurs directly onto sand, with saltmarsh being replaced by alder carr conditions and a drying peat surface. The return of marine conditions was recorded by an increase in *Gramineae* and saltmarsh taxa (in core 2B, at 2937-2793 cal. yr. BP). Elsewhere at Midley Church Bank (in core 10B), the increased marine influence is indicated by an increase in aquatic and wetland plants and the establishment of saltmarsh at the upper, transgressive contact of the peat unit, at 2341-2185 cal. yrs. BP.

In the fore-marsh area, Tooley and Switsur (1988) and Tooley (1990b) studied the palaeovegetation by sampling peat units from between the gravel swales, possibly of an initial gravel barrier. From two cores (Broomhill 1 and A) Tooley and Switsur (*op. cit.*) demonstrated that saltmarsh and freshwater reedswamp, with aquatic plants, had dominated throughout the accumulation of the peat units. However, Long *et al.* (1996) suggest that the upper part of the peat contact in the area of Broomhill is commonly abrupt and varies in age, with a clear truncation recorded in core Broomhill A (Tooley and Switsur, 1988; Tooley, 1990b).

All of the palaeovegetational studies reviewed thus far have come from peat units believed to have been deposited behind an initial gravel barrier. One study, however, has been completed at a site outside this area, at

Wickmaryholme Pit. This is one of a series of natural gravel pits present on the Dungeness foreland (Eddison, 1983; Ferry and Waters, 1988). Currently this pit is close to the coastline, but old maps of the area, *i.e.* Powker's map of AD 1617 (reproduced in Green, 1968), demonstrate that the site was formerly at least 350m from the coast. The organic sediments in this peat trace the removal of marine conditions allowing, first, saltmarsh and then a freshwater pond to exist. This pond gradually shallowed as *Myriophyllum alterniflorum* was replaced by *Potamogeton* and *Typha angustifolia*. Following the shallowing of the pool, saltmarsh conditions increased as the marine presence increased at the site. This period of peat accumulation occurred at a date much later than those discussed from the main back-barrier environment, *i.e.* between 2308-1737 cal. yrs. BP and ca. 700-1000 cal. yrs. BP.

2.1.4.3 The Upper Inorganic Sediments (The Younger Alluvium of Green (1968)).

Relatively few detailed palaeoenvironmental studies have been completed on the inorganic sediments overlying the peat. One reason for this may be due to the paucity of environmental indicators contained within them. For example, Long and Innes (1995a) noted that very few diatoms were present in the sediment overlying the peat at Brookland, and those which were present were indicative of a mixed marine, marine-brackish and brackish environments.

In the inorganic sediments overlying the peat in the valleys to the west of Romney Marsh, Waller *et al.* (1988) have identified, from foramanifera, a range of environments, from freshwater to high energy nearshore sediments, indicating the complex and varied depositional characteristics that existed during deposition.

At Broomhill, Tooley and Switsur (1988) identified that the inorganic sediments overlying the main marsh peat were rich in foraminifera. The foraminiferal assemblage was dominated by *Protelphidium anglicum* and *Jadammina macrescens*, both indicating brackish water conditions. In addition, Tooley (1990b) presented the results of diatom analysis from Broomhill which were indicative of a brackish to marine environment with open access to the sea.

2.1.4.4 Denge Marsh Sediments.

It has been suggested by Plater and Long (1996) that the marsh sediments of the Denge Marsh area (Long and Fox, 1988; Plater, 1992) bear a broad similarity to the post-peat deposits of the main marsh (Green, 1968; Waller *et al.*, 1988). Plater (1992) has demonstrated that the sediments of Brickwall Farm, Denge Marsh were deposited under progressively lower energy marine to brackish tidal conditions. Plater and Long (1995) identified a tripartite sequence of inorganic sediments that were deposited under decreasing energy conditions, which, they suggest, may be associated with the progradation of the Dungeness foreland. The lower part of these sediments contain diatoms deposited on an intertidal mudflat to lower marsh (Plater and Long, 1995). An additional study of the marsh sediments outside the main back-barrier was completed by Long and Hughes (1995) in the South Brooks. They demonstrated that these marsh sediments were deposited in a marine inlet or channel.

2.1.5 The Drainage of the Back-barrier Environment.

It has been demonstrated by Long and Innes (1995a) that an understanding of the stratigraphy of the back-barrier sediments can provide evidence regarding the drainage of the back-barrier. Long and Innes have suggested that in the back-barrier environment areas of the channel sediments exist, and that by delimiting the location of these sediments the pattern of back-barrier drainage can be resolved.

Eddison (1983b) has stated that it is important that the location of the river Rother is considered in any review of the evolution of Romney Marsh due to the important role that back-barrier drainage will have played. The River Rother has a catchment of 47,000 ha on the Eastern High Weald in Kent and Sussex, and is the principal river passing to the sea across the reclaimed marshland between Fairlight and Hythe (Eddison, 1988). Indeed, White (1928) noted that the rich alluvial land of Romney Marsh is due largely to the gift of the Rother which has ranged across the area.

Livett (1930) recorded that Nennius, in the eighth century, described Lommon marsh (Romney Marsh) as having many rivers flowing into it, and only one river, the Lemn (Limen), going out to the sea. The River Rother was called the Limen until the sixteenth century (Ward, 1952), the name Rother being a modern invention from the Sussex village of Rotherfield near its source (Brooks, 1988). Eddison (1983a) recognised that as the barriers are higher than the marsh, any major break is likely to form a river estuary, which (in Romney Marsh) is generally an extensive one and likely to catch the River Rother sooner or later.

Three major rivers flowed across the marsh throughout its evolution: the Brede, the Tillingham, and the Rother. These tend to be referred to collectively as the Rother once they flow across the marsh. It is their course across the marsh, and the location of their estuaries that have been the focus of debate. Many authors believe that the three major gaps in the gravel barrier have provided estuarine openings at Hythe (in the north-east of the marsh), at Romney[1], and south of Rye (the current exit of the marsh rivers) (Drew, 1864; Robertson, 1880; Dowker, 1897; Gulliver, 1897; Lewis, 1932; Homan, 1938; Piper, 1950; Ward, 1952; Williamson, 1959; Green, 1968; Cunliffe, 1980; Eddison, 1983a; Tatton-Brown, 1984; Tooley, 1995). This view is not held by all authors, and debate continues over the exact location, nature and timing of the river exits from the marsh. In discussing this it will be necessary to consider other issues that have had an

[1]'Romney' is used to refer to Old Romney at this time in the evolution of the marsh, as New Romney was yet to be created.

influence on the history of the rivers, such as the Rhee Wall. Also, the immediate consequences of the changes of river channel must be considered, such as the catastrophic destruction of Old Winchelsea and the ruin of Romney as an important port.

2.1.5.1 The Location of the River Rother on the Marshland.

Much of the debate regarding the drainage of the back-barrier environment has revolved around whether the major channel draining the marshland flowed across the northern margin of the marsh or through the southern areas of Walland Marsh. Therefore, before considering the potential gaps in the barrier, through which the back-barrier water exited the marsh, a consideration of the route that the rivers took across the marshland will be made.

2.1.5.1i A Former Channel of the Rother in the North of Romney Marsh?

Ward (1933b) suggested that any theory of the evolution of Romney Marsh must consider the Limen at West Hythe. The tidal inlet of the Limen has been described by Gilbert (1933) as a broad channel at the foot of the hills that led to the sea. Some authors have proposed that the river Limen debouched into a large estuary at Appledore, from which a tidal inlet flowed to the sea (Elliot, 1852; Robertson, 1880; Furley, 1880; Steers, 1963). Ward (1933b) noted that some authors believed that any waterway in the northern area of the marsh was an inlet of the sea and not connected to a river. It was also proposed (Ward, 1933a) that any channel may have been very small, very estuarine and very 'un-riverly', but he believes that some watercourse called the Limen existed here. It is suggested, therefore, that the Limen in the north-east of Romney Marsh was actually a major tidal creek draining the back-barrier saltmarsh (Eddison, 1983a).

Elliot (1852) and Topley (1875) stated that no evidence of a watercourse across the north of the Romney Marsh had been found at the time of their writing. In addition, Eddison (1983a) also states that no feature showing a silted Rother exists in the Appledore Dowels. Lewin (1862) recognised that the physical evidence on Romney Marsh is against the existence of a river in the northern area of the marsh, and he believed that the marsh had always drained southwards from Appledore, as there was no evidence of channels in the region of the hills of the northern border of Romney Marsh. The drainage in northern Romney Marsh is believed by Smart *et al*. (1966) to be north and eastwards into a broad depression towards Hythe which has been confused with the River Rother.

A comprehensive study by Wass (1995) has greatly assisted in resolving the nature of this northern course. Wass illustrates that, thus far, theories have been based on historical evidence presented by Holmes (1907) and Ward (1931, 1933a&b), who refer to the Limen south of Kits Bridge and Warehorne Bridge. Also Lambarde (1576) referred to Danes at the mouth of the Limen near Newchurch. But

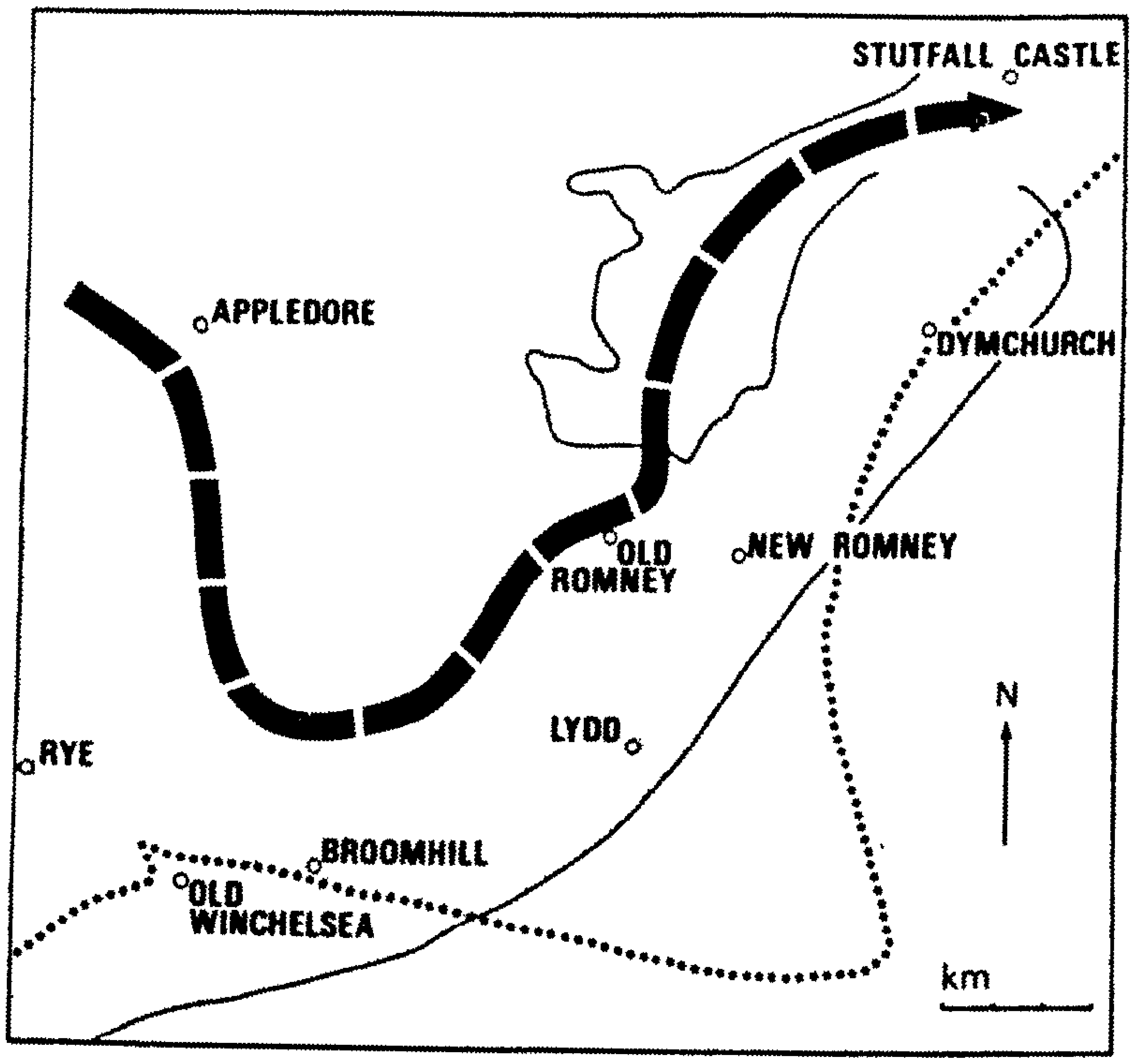


Figure 2.10: A proposed course of the Rother into a Hythe estuary.

Source: Green (1988).

Elliot (1862, 1874) and Topley (1875) found no evidence of this river and argued that the meandering was old relic sea creeks. A river course running south from Appledore and then north-east towards New Romney appears to be the river course favoured by Green (1988) (see figure 2.10). Wass (1987, 1995), however, suggests that the balance of evidence favours a tidal inlet over fluvial conditions for the channel from Appledore to Hamstreet.

The coarse nature of the sediments and the microfossil data presented by Wass (1995) suggest a marine origin, as does the fact that the creek tapers from east to west. Wass also identified laminations towards the east, indicating strong tidal conditions depositing a sandy layer.

Ostracod evidence presented by Wass (1995) also points to the higher energy environment being in the east, and foraminiferal evidence illustrates that saltmarsh conditions prevailed in the west, whereas extensive intertidal mudflat conditions existed in the east. Wass (1995) concludes, therefore, that this channel represents the upper end of a sheltered arm of a tidal inlet. This conclusion agrees with the proposal of Green (1968) and Green (1988) that the Limen may have flowed into a large estuary in Romney Marsh proper, as defined by the New Marshland of Green (1968) (see figure 2.11).

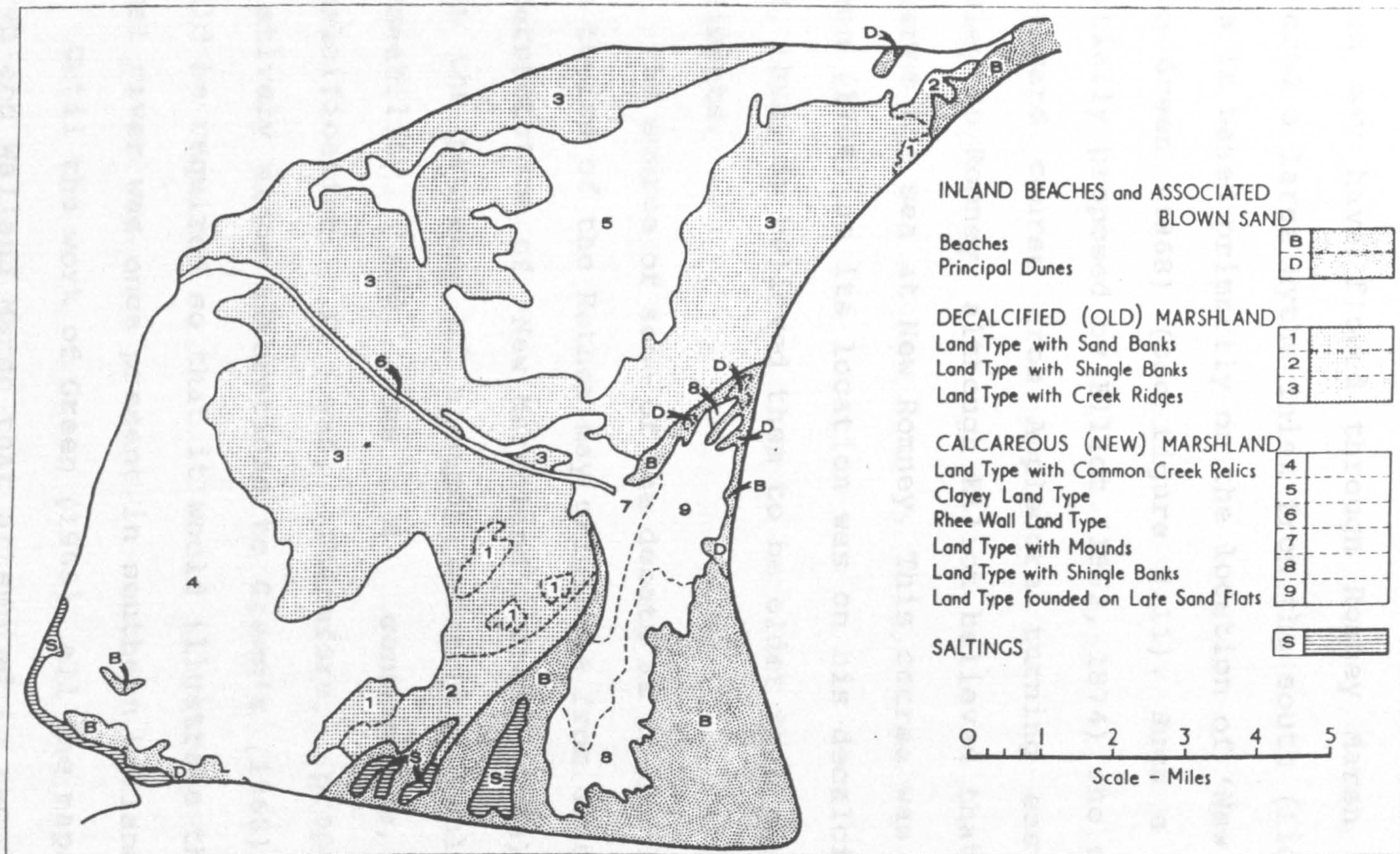


Figure 2.11: Proposed location of a large Hythe estuary in Romney Marsh proper, marked by the sediments labelled 5.

Source: Green (1968).

2.1.5.1ii A Southerly Arcing Rother Channel in Walland Marsh?

Green (1988) illustrated that he believed that the Limen may have flowed through Romney Marsh proper and entered a large Hythe inlet from the south (figure 2.10). This is based primarily on the location of 'New marshland' from Green (1968) (see figure 2.11). Such a course was initially proposed by Elliot (1862, 1874) who suggested a southward course from Appledore turning eastward past Midley to Romney, although Elliot believed that the river entered the sea at New Romney. This course was opposed by Green (1968) as its location was on his decalcified soils and, thus, he believed them to be older than any estuarine sediments.

The source of some of the debate as to the location of the course of the Rother may originate from Green's (1968) interpretation of 'New Marshland'. Green (1988) suggested that the presence of a sandy substratum would increase permeability and, as a consequence, promote decalcification. It was, therefore, proposed that relatively minor alterations to Green's (1968) soils map would be required so that it would illustrate that a large tidal river was once present in southern Walland Marsh.

Until the work of Green (1968), all the maps of Romney Marsh and Walland Marsh that attempted to reconstruct the topography of the area before the thirteenth century, showed a large channel in a broad southerly arc through

Walland Marsh (Tatton-Brown 1988). However, it is now assumed by many writers, e.g. Cunliffe (1980), that the main course of the Rother ran from Appledore to New Romney down a small meandering stream. Tatton-Brown (1988) suggests that recent study reveals that this stream was very small, and that perhaps we should return to the earlier ideas, with a river in southern Walland Marsh. This is in agreement with Green (1988) and Eddison (1983b) who stated that this channel was too small to be an estuarine Rother. This would also explain the concern of Brooks (1988) over *Rumen ea* (broad river) as to whether the name referred to a large channel in Walland Marsh rather than the small stream from Snargate.

A number of workers proposed a Rother channel making a broad southerly arc into Walland Marsh (Elliot, 1862, 1874; Livett, 1930; Ward, 1952; Parkin, 1973; Green 1988). Elliot (1862) proposed that a former river flowed from Appledore to Fairfield, to Midley, and then to Romney. This, he believed, was in existence in AD 1066. Elliot (*op. cit.*) believed that by AD 1287, the Rother was flowing to New Romney from East Guldeford to join the AD 1066 course north of Lower Agney. Green (1968) considered these two channels unlikely, although he did suggest that the Rother may have followed this course at some earlier time.

Ward (1952) and Parkin (1973) are in broad agreement with Elliot (1862), in that their reconstructed channels (figures 2.12 and 2.13) follow a similar course. However, Tatton-Brown (1984) suggested that these authors did not

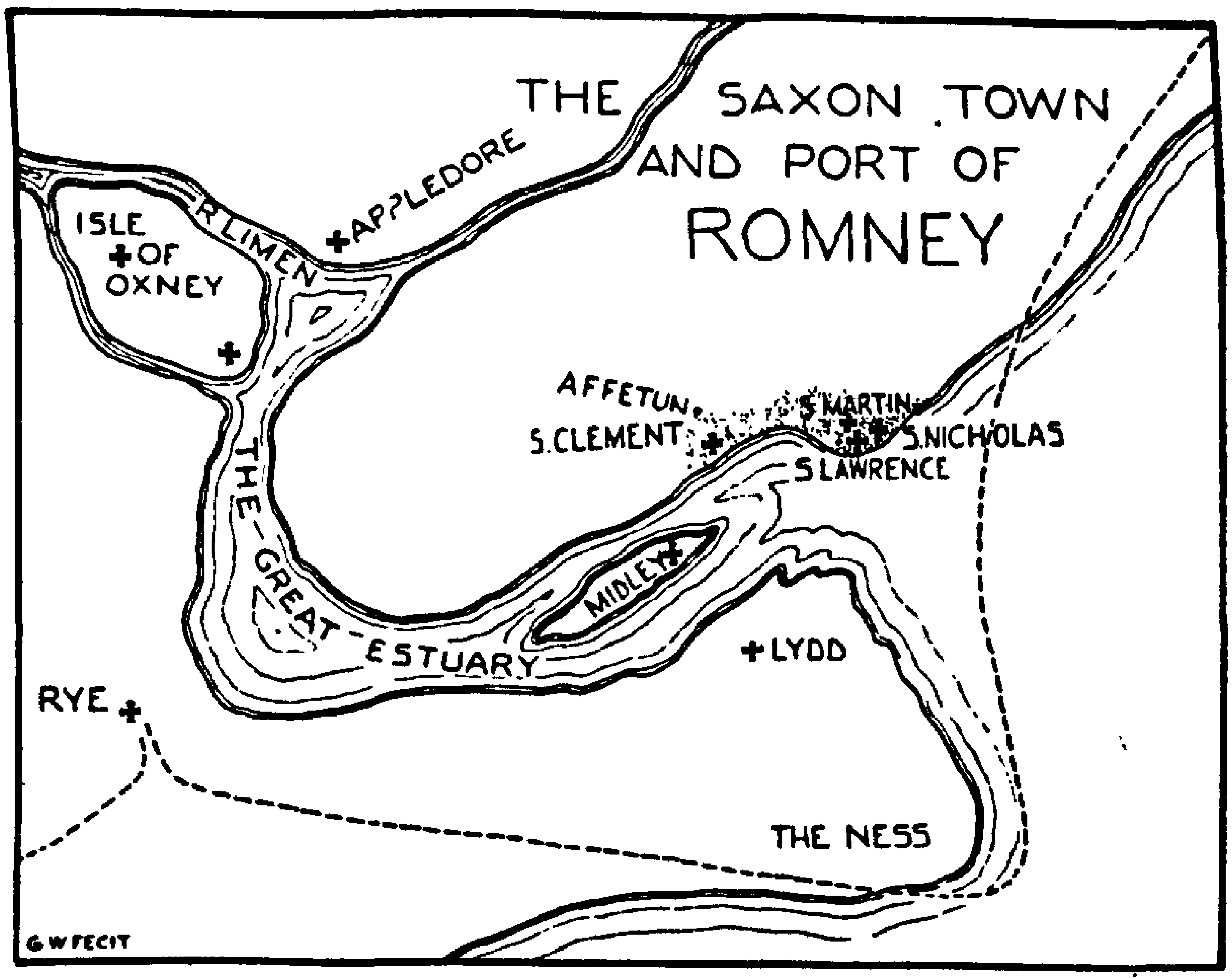


Figure 2.12: A map of Saxon Romney Marsh with a southerly arcing river in Walland Marsh.

Source: Ward (1952).

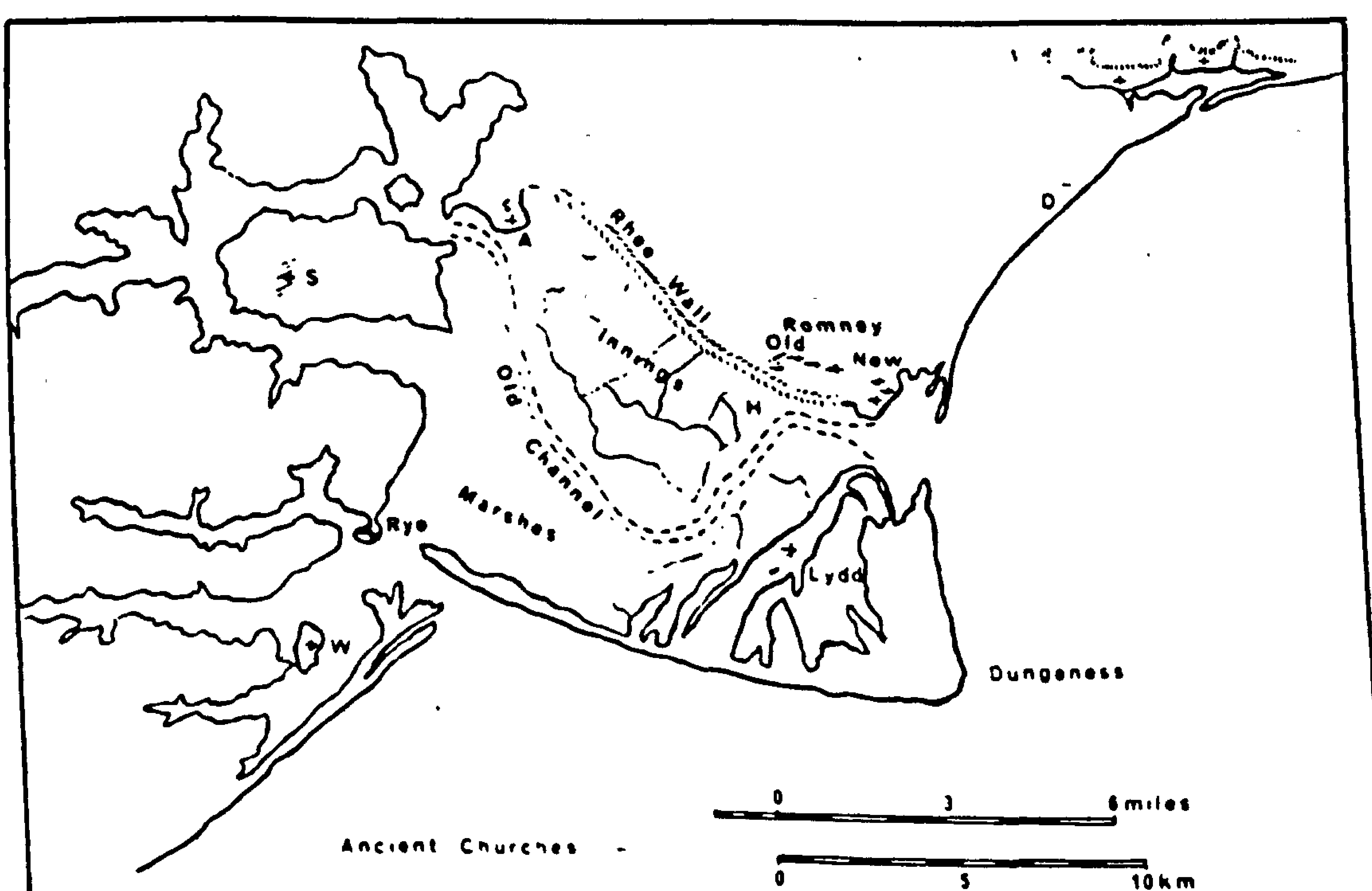


Figure 2.13: A map of Romney Marsh in ca. AD 1200 demonstrating a southerly arcing former river channel.

Source: Parkin (1973).

understand the significance of the Rhee Wall, and so had proposed a 'non-existent' channel in Walland Marsh. Tatton-Brown (1988) subsequently agreed that the principal Rother channel was probably on line with the later Wainway Channel, in south Walland Marsh, then running north-east to Old Romney and, thus, following a course similar to that proposed by Ward (1952) and Parkin (1973). Early maps from the late 16th and 17th centuries show a Wainway Channel as a broad inlet in south Walland Marsh (Symondson 1594; Poker 1617). Other maps showing the same southerly arc of a former river channel in this area have been presented by Lewin (1862), Dowker (1897), Lewis (1932), Gilbert (1933) (figures 2.14, 2.15, 2.16 & 2.17).

Importantly, a stratigraphic transect completed by Long and Innes (1995a) has provided the first detailed geomorphological evidence regarding the existence and location of a major tidal river in Walland Marsh. Long and Innes mapped extensive channel sands extending southwards from Little Cheyne Court towards Broomhill and have suggested that the sediments indicate that a major tidal channel persisted in this southern area of Walland Marsh.

It appears, therefore, that the Rother flowed across Walland Marsh in a southerly arc towards Romney and into a large tidal estuary within Romney Marsh proper. Cunliffe (1980) has proposed that water has exited the back-barrier at three different places throughout its evolution, at Hythe, Romney and at Rye. These inlets will now be considered.

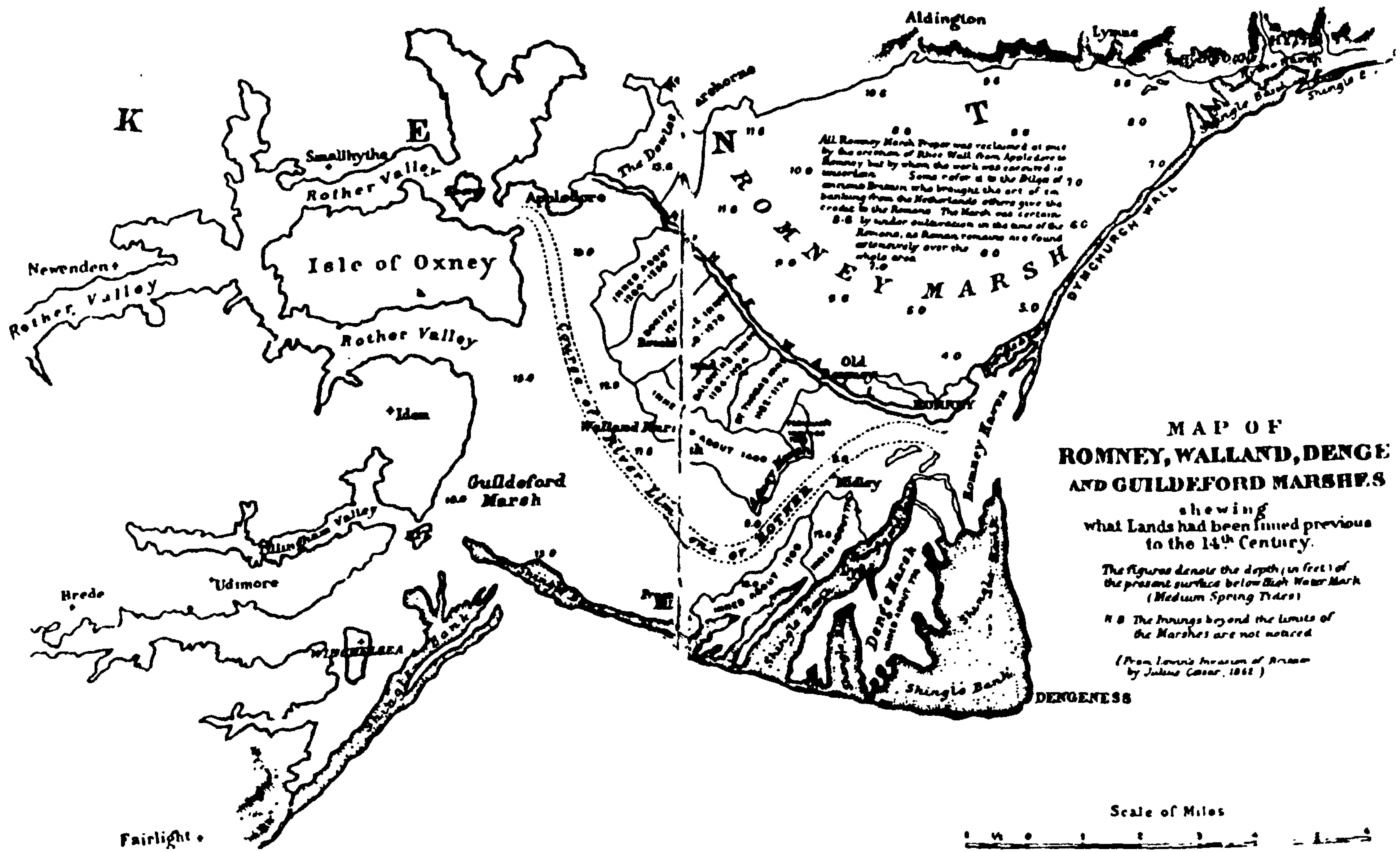


Figure 2.14: A map of medieval Romney Marsh with a southerly arcing river Rother.

Source: Brentall (1972), after Lewin (1862).

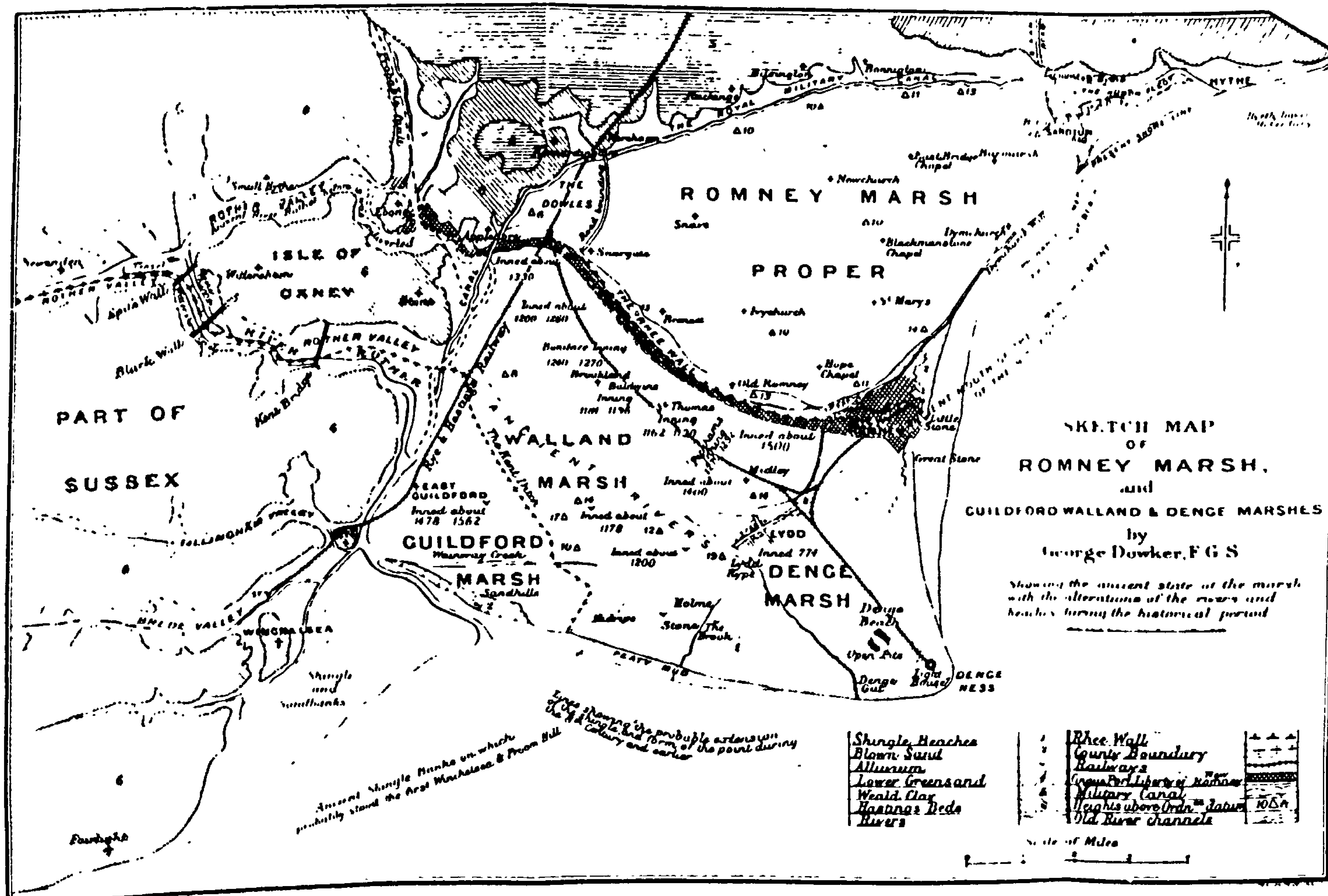


Figure 2.15: Map demonstrating a south-westerly trending river across Walland Marsh.

Source: Dowker (1897).

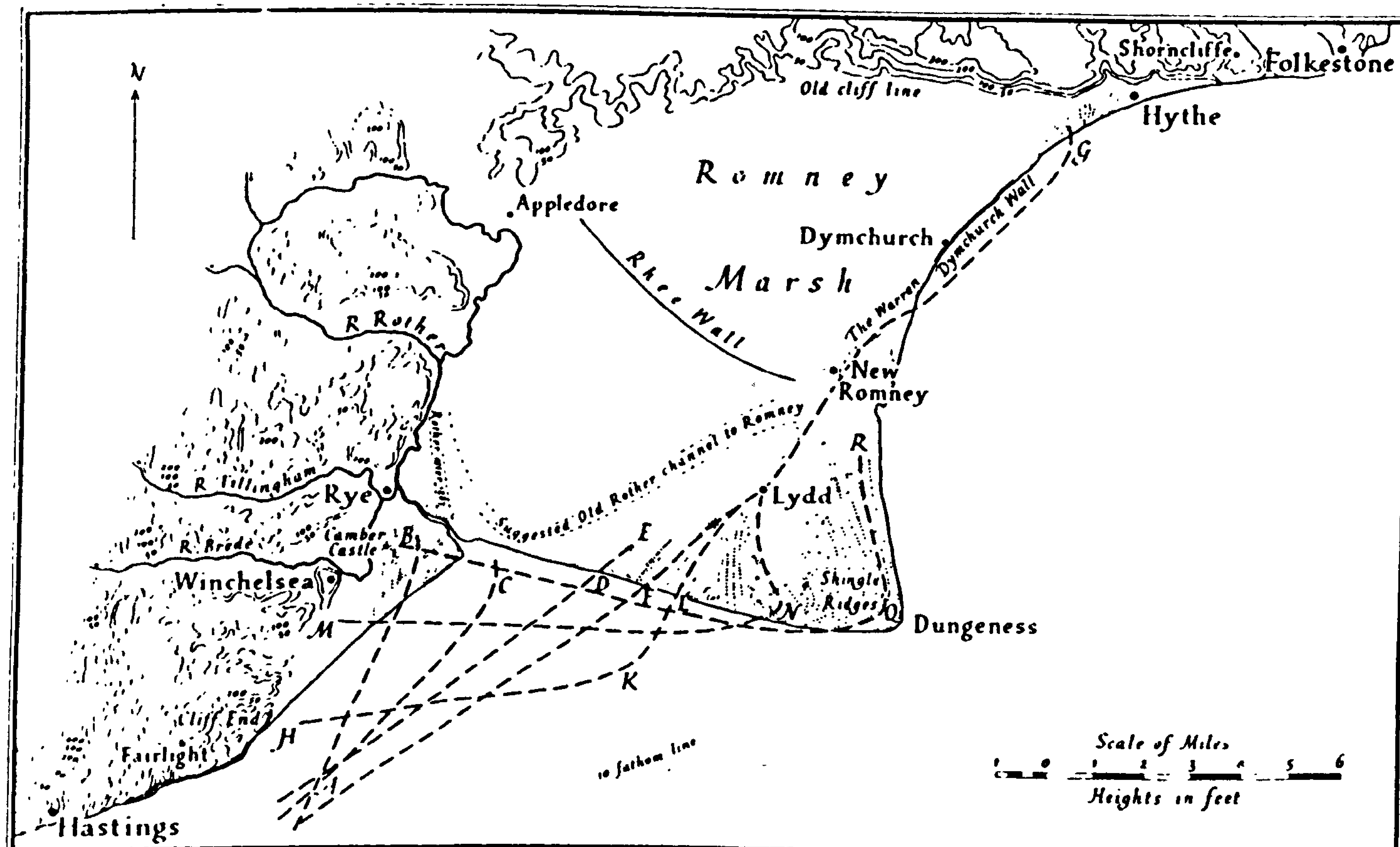


Figure 2.16: A map of Romney Marsh demonstrating a southerly arcing former river channel.

Source: Lewis (1932).

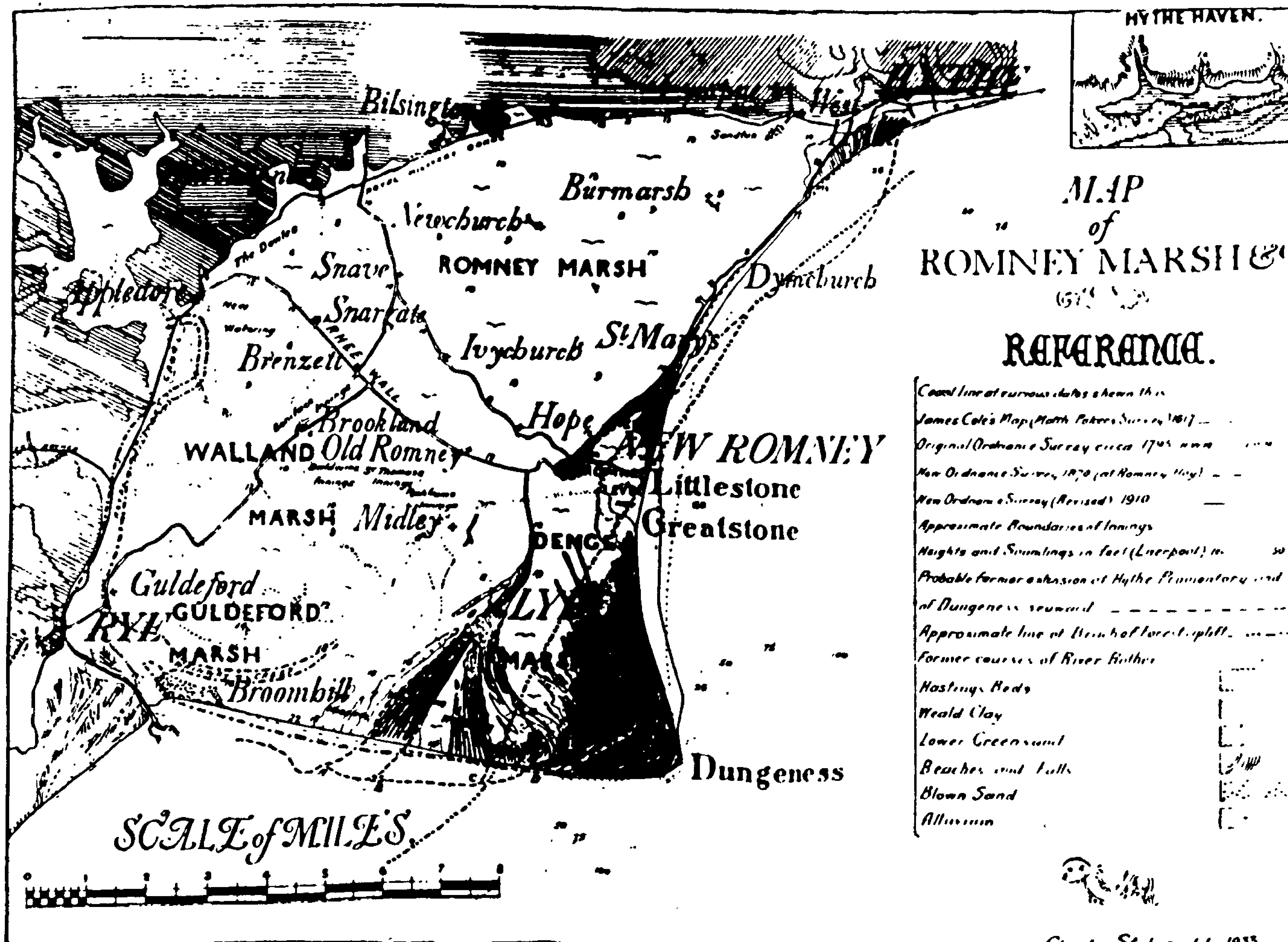


Figure 2.17: Map illustrating a 'Wainway Creek' with a distinct southerly arc.

Source: Gilbert (1933).

2.1.5.2i The Rother Estuary at Hythe.

Many authors have proposed a river Limen exiting the marsh at Hythe (Meryon, 1845; Drew, 1864; Hussey, 1867; Bishop of Dover, 1880; Furley, 1880; Robertson, 1880; Gulliver, 1897; Stokes, 1930; Ward, 1931 1933a&b; Gilbert, 1933; Homan, 1938; Piper, 1950; Edmunds 1954; Green and Avery, 1955; Steers, 1963; Green, 1968; Cunliffe, 1980; Eddison, 1983a,b; Tatton-Brown, 1984; Tooley 1995). Green (1968) proposed that Hythe was the oldest and the longest-lived of the outlets on the marsh.

The timing of this outlet is also a matter of debate, in terms of both when it was in existence and when it may have closed. A number of authors propose that this outlet was in existence in Roman times (Elliot, 1852; Bishop of Dover, 1880; Hussey, 1867; Lewis, 1932; Steers, 1963; Tatton-Brown, 1984; Tooley, 1995). However, Edmunds (1954) stated that the Limen must have ceased to flow to Hythe by at least the Roman times, as he believed that the Rhee Wall was built at this time and, thus, would prevent a river reaching Hythe. Other authors have suggested dates for when the river outlet was at Hythe, *i.e.* until AD 700 (Drew, 1864), the end of the eighth century (Homan, 1938), the end of the ninth century (Green, 1968), and in the eleventh century (Piper, 1950). Ward (1931b, 1933a&b) has proposed a number of dates in the eighth century for the existence of the Limen in this area from the historical record. It is illustrated by Green (1968) that the Limen must have been present in AD 300 due to the Saxon shore fort at Stuttfall, and that the inlet was still open in AD 732.

2.1.5.2ii The Rother Estuary at Romney.

There is a consensus of opinion that, at some time, the Rother came to Romney and created a prosperous port. It is proposed that an inlet was created south of Romney, following a slight sea-level rise, but that no river, as such, was in the inlet which gradually extended inland until it captured the Rother (Steers, 1964; Cunliffe, 1980). The nature and location of this change of channel by the river Rother, and the history of the river whilst at this outlet, remain the focus of continuing research.

Some confusion exists as to the channel that the Rother took to Romney, as has been outlined in section 2.1.5.1. A number of authors do not address this issue and, instead, discuss the canalisation of the Rother in the Rhee Wall as being the only course of the Rother to Romney (Hussey, 1867; Gulliver, 1897; Piper, 1950).

According to Green (1988), the date for the breaching of the gravel near New Romney is completely unknown, although a minimum date is given by the charter of AD 741. The timing of the arrival of the Rother at Romney has been discussed by a number of authors. Drew (1864) proposed that the Rother had reached Romney by AD 700, whilst Brooks (1981) and Green (1988) suggested AD 741. Indeed, Brooks (1988) noted that the first reference to the Limen at Romney was a charter of AD 741 in which AEthelberht II granted the church of Lyminge a fishery at the mouth of the

River Limen. From the Saxon charter, Dowker (1897) proposed a date for the Rother at this location in the seventh or eighth century, and Cunliffe (1980) proposed an early medieval date for the arrival of the Rother at Romney.

At some point in time, however, the Rhee Wall was built. The reason or date for the construction of the wall are open to question. One obvious impact of the Rhee Wall was that it was no longer possible for a Rother/Limen to reach Hythe, as it would have had to flow through the wall (Livett, 1930; Gilbert, 1933; Ward, 1940; Edmunds, 1954; Steers, 1963; Tatton-Brown, 1984). The Rhee Wall was actually two walls between 50 and 100 metres apart (Eddison, 1983b). Indeed, Dugdale (1662) believed the Rhee Wall to be 'the most ancient embankment in the realm'. The main issue of debate surrounding the Rhee Wall is whether it was built to drain the area of Romney Marsh proper, or to canalise the river Rother to the port of Romney.

A number of authors believe that the Rhee Wall was built to reclaim Romney Marsh proper (Dugdale, 1662; Lewin, 1867; Robertson, 1880; Stokes, 1930; Green and Avery, 1955; Williamson, 1959). A greater number of authors, however, propose that the Rhee Wall was actually built to canalise the Rother to Romney (Elliot, 1847,1862; Holloway, 1849; Drew, 1864; Robertson, 1880; Livett, 1930; Homan, 1938; Ward, 1952; Green, 1968; Cunliffe, 1980; Eddison, 1983b). Livett (1930) suggested that if the intention was to 'in' Romney Marsh proper, only one wall would have been required. In addition, a canal would become necessary once a natural channel was no longer open to shipping. Green (1988) suggested that the Rhee Wall was not a wall but an artificial channel to feed the Rother water to the tidal inlet at Romney due to the deterioration of a natural channel through Walland Marsh. Eddison (1983a) disagreed with the suggestion that the Rhee Wall was a sea bank, and suggested that there is no difference between the sediments or altitudes on either side of the wall. Therefore, the Rhee Wall must have been built to replace a natural river course.

It has been demonstrated by Green (1968) that channel silts exist in the sediments between the two walls, and that the Rhee Wall was built to channel water. Homan (1938) illustrated that the word 'Rhee' is from the generic word for stream in late English. Evidence that the Rhee once held water is presented by Robertson (1880), who cited an inquiry by the commissioners of Queen Elizabeth in 1565 "the lande betweene the walles... ...occupies the entire site of a crike or waterway, sewared or dryed upp". Drew (1864) agreed, stating that the river gradually deposited sediment until the bed was raised to the height of the land on both sides.

Much debate concerns the date of construction of the Rhee Wall. Green (1968) noted that most authors considered the wall to be built in Roman times (Dugdale, 1662; Somner, 1693; Elliot, 1847; Lewin, 1862; Robertson, 1880; Gulliver, 1897; Holmes, 1907; Ballard, 1910; Livett, 1930; Piper, 1950). However, some authors have argued that the wall may have been constructed at an earlier date.

It would appear that at some time, the Rother waters were diverted away from the natural channel and down the Rhee Wall in order to keep open the port of Romney. Lewis (1932) and Williamson (1959) proposed that even after the diversion of the Rother to the Rhee Wall, a southern arc of the Rother existed in Walland Marsh. The existence of such a channel may have assisted the third major change of the Rother.

It appears, therefore, that at some time before AD 741 the Rother changed its channel to exit the marsh in the region of Romney. It is not known whether or not this was as a consequence of the blocking by gravel of the Hythe inlet, or, indeed, if the two outlets were in existence at the same time. However, at some time it became necessary to canalise the Rother waters from Appledore down the Rhee Wall to Romney in order to keep the port open. This was probably due to the silting up of the channel running through Walland Marsh. Eventually, though the Rother changed channel again to its present location, south of Rye.

2.1.5.2iii The Rother Estuary at Rye.

The third and most recent change of the course of the river Rother was to its present location south of Rye. The circumstances and dating of this change are a subject of less debate than the outlets at either Hythe or Romney. It was proposed by Dowker (1897) and Cunliffe (1980) that an outlet for the rivers Brede and Tillingham may have existed for some time at Rye. The existence of a marine inlet prior to the 13th century is also suggested by Green (1988) who believes that the histories of Rye and Winchelsea indicate the proximity of a sheltered inlet from at least the 11th century. Green (*op. cit.*) identifies some debate on this issue, as 13th century documents suggest a direct road from Winchelsea to Broomhill and, thus, no inlet.

However, the historical links of Rye and Winchelsea are with Hastings, to the west, rather than with the Rother-dependent towns such as Lydd, Romney and Small Hythe. Therefore, it is likely that Rye and Winchelsea were located on the estuary of the rivers Brede and Tillingham, whose discharge would have been needed to keep the estuary free of gravel. According to Tatton-Brown (1988), an inlet at Rye must have existed before the 13th century due to the rapid growth of Rye and Old Winchelsea from relative obscurity in the Domesday AD 1086 to full membership of the Cinque Ports by AD 1197. It is well documented by many authors that in AD 1287, following a series of catastrophic storms, the Rother found a new outlet south of Rye (Drew, 1864; Furley, 1880; Gulliver, 1897; Lewis, 1932; Piper, 1950; Ward, 1952; Edmunds 1954; Green and Avery, 1955; Williamson, 1959; Green, 1968; Parkin, 1973; Cunliffe, 1980; Tatton-Brown, 1984).

The reason for the change of course of the Rother has been debated by Gulliver (1897) and Piper (1950), who proposed that due to the silting of the Romney outlet, the Rother was forced to seek a new channel. Drew (1864) and Williamson (1959), however, proposed that the storm of AD 1287 blocked the New Romney outlet, and so the river was forced to seek a new outlet. The other authors believe that the great storm breached the gravel barrier south of Rye and the Rother was captured. In addition, Piper (1950) illustrated that the historian Lambarde (1576) cited an 'earthquake' for the change of river channel.

The change of river channel to Rye had a number of consequences. For example, the old Rother channel or channels (including the Rhee Wall) silted up rapidly (Bishop of Dover, 1880; Furley, 1880; Dowker, 1897; Tatton-Brown, 1984).

2.1.6 The Control of Local Versus Regional Processes in Coastal Sedimentation in Romney Marsh.

Changes in the protective gravel barrier are a local control on the coastal evolution of Romney Marsh, whereas, changes in sea-level (section 2.1.3) or storm magnitude / frequency are regional controls. The relative importance of these (and other) local and regional controls will be considered here.

During the course of recent research into the evolution of Romney Marsh and the coastline of East Sussex, debate has existed as to whether local or regional processes have dominated coastal sedimentation. It has been argued, particularly by Jennings and Smyth (1982; 1985;

1987; 1990), that shingle barriers (local processes) have been the major influence on coastal sediments in Romney Marsh and East Sussex. However, Burrin (1982) and Tooley and Switsur (1988) have stated that Jennings and Smyth have over-stated the role that local processes have played, and that the evidence for barrier construction and breakdown lacks stratigraphic underpinning (Jennings and Smyth, 1990). Tooley and Switsur (1988), having established a concurrence of dates for regressive overlaps in the back- and fore-marshes suggested that the conclusions of Jennings and Smyth (1982; 1985; 1987) needed some re-examination.

It was argued by Long (1992) that local processes have controlled the chronology of sea-level tendencies in south-east England in the last 7500 cal. yrs. BP. However, Long (*op. cit.*) has also identified that the altitude of sea-level index points from this area strongly reflect regional time / altitude changes in sea-level. Long (1992), therefore, concluded that the coastal stratigraphy of south-east England appears to be the product of both local and regional processes.

In addition, Long and Innes (1993) identified that a similarity existed between the age of peat accumulation at Midley Church Bank and the minimum ages of barrier emplacement of Tooley and Switsur (1988) and Needham (1988). However, Long and Innes (*op. cit.*) noted that it is difficult to determine whether the two environments are related, and thus, they concluded that a combination of protective gravel barriers and a slowing in the rate of sea-level rise allowed peat accumulation to occur at Midley Church Bank. It was identified by Long and Innes that the altitudes of the sea-level index points at Midley Church Bank were similar to those of other parts of East Sussex, which would appear to indicate a regional control on coastal sedimentation. However, Long and Innes also demonstrated that the ages of the initiation and cessation of peat accumulation across Romney Marsh are variable and, thus, no synchroneity of sea-level tendencies appears to exist, maybe indicating some influence from local processes.

Long and Hughes (1995) have suggested that the changes in depositional features of the Dungeness foreland have probably not been caused by sea-level changes, but are more likely to be due to changes in storm incidence and gravel supply. Consequently, they believe that the predominant influence on the gravel of the Dungeness complex was not regional sea-level changes but may have been an increase in storm incidence, as has been identified by Lamb (1977) for the period between ca. 2765-2741 and 1918-1867 cal. yrs. BP and also between ca. 460-436 and 315-304 cal. yrs. BP.

It has been suggested by Long *et al.* (1996) that the predominant control on the sediments of the Romney Marsh area may have changed through time. They suggest that until ca. 6868-6792 cal. yrs. BP regional sea-level controlled coastal sedimentation. Conversely, after ca. 6868-6792 cal. yrs. BP they propose that barrier dynamics and stability in combination with changes in sea-level, controlled coastal sedimentation.

It can be seen, therefore, that a number of views are held concerning whether regional or local processes have dominated coastal sedimentation in the Romney Marsh and East Sussex areas. The balance of argument would appear to point to an interaction of both regional and local processes having a controlling influence on the coastal sediments, but with differential relative influences both spatially and temporally.

2.2 Barrier and Back-barrier Processes.

Comprehensive reviews of the processes interacting in barrier / back-barrier environments are provided by Roy *et al.* (1994) and Cooper (1994). Cooper (*op.cit.*) also cites a number of case studies from a wide range of geographical locations and reviews a series of evolutionary models that have been developed regarding barrier / back-barrier evolution. Due to the wide-ranging characteristics of both barrier and back-barrier environments, many of the processes discussed in the literature are not directly applicable to this study as they do not concentrate on drift aligned systems. As a consequence, only the facters potentially influential on the evolution of the Romney Marsh barrier / back-barrier system will be considered here.

It is suggested that large scale behaviour of wave-dominated coasts can be understood in terms of two concepts (Roy *et al.*, 1994). The first is the geological inheritance of the coastline. The second concept considers the processes operating over shorter time periods, *i.e.* hundreds to thousands of years, and relates to the large-scale morphodynamics of the coastal sediments. Roy *et al.* (*op.cit.*) suggest that three major groups of coastal deposits exist, sediments deposited under marine transgression, under stable sea-level and under marine regression.

The major controlling processes on the coastal barriers deposited under the above regimes are suggested to be sea-level and sediment supply (Roy *et al.*, 1994). The inter-relationship of sea-level and sediment supply is demonstrated by Ireland (1987), who suggests that breaching of the barriers and the subsequent inundation of the back-barrier environments in Rio de Janeiro State was due to rising sea-level. In contrast, a similar back-barrier sedimentary record in south-east Ireland which experienced increased marine influence due to barrier breaching was demonstrated by Carter *et al.* (1989) to have been controlled by low sediment supply, resulting in the reworking of barrier sediments.

Other changes in the nature of the barrier, such as closure of the back-barrier outlet causing back-barrier isolation, will have a significant influence on the evolution of the system. Cooper (1994) suggests that outlet closure will tend to occur due to a combination of low fluvial discharge and barrier sediment build up. Indeed, it is suggested that the evolution of the Dutch Waddens was due to the closure of the inlet (Carter, 1988).

Cooper (1994) demonstrates that much variation exists in the evolution of back-barrier lagoons. The evolution of back-barrier lagoons is summarised by Kjerfve and Maghill (1989) as: 'the combined action of marine and fluvial

> processes....[leading to].... trapping and infilling of semi-enclosed coastal systems, including coastal lagoons, and the re-shaping of seaward boundaries'.

It is also noted that the scale of coastal lagoons varies, with the deposits ranging from 40m thick in Natal and Zululand, whereas in Maine and Brazil, lagoon deposits tended to be only 3-4m thick. One of the processes affecting back-barrier coastal sedimentation is considered by Cooper (1990a) to be sediment supply. Cooper (1990a) compared the sites of Lake St. Lucia and Kosi Bay, the former is infilled by 40m of fluvially derived sediment, whereas the incised valley bathymetry of the latter is largely preserved due to the fact that no rivers discharge into the lagoon. The interaction of sediment supply and sea-level rise is suggested by Cooper (1994) to account for the observed differences in the stratigraphies of the back-barriers of Maine, and Brazil and New South Wales. Transgressive back-barrier sequences are recorded in many lagoons in Maine, where sediment supply was low and sea-level rise rapid. Conversely, in Brazil and New South Wales sediment supply was relatively high and, thus, prevented the inundation of back-barrier sediments.

The response of barriers to sea-level rise has been discussed by Cooper (1994), who suggests that the response of a barrier to sea-level rise has a direct bearing on the evolution of the back-barrier sediments. Cooper (*op.cit.*) has identified that if a transgressive barrier, under conditions of sea-level rise, is stabilised and accretes upwards, then the volume of the back-barrier lagoon is increased, as long as the rate of sea-level rise is greater than the rate of sediment supply. Conversely, the landward migration of a barrier under conditions of sea-level rise may keep the volume of the back-barrier lagoon constant.

The inter-relationship between barrier and back-barrier environments is further demonstrated by Duffy *et al.* (1989) who suggest that barrier grain size limits washover into the back-barrier environment. Duffy *et al.* (1989) suggest that only limited overwash sediments are present in the relatively coarse grained lagoons of Maine. A contrasting process is recorded in areas of sand barriers, where washover fan sediments can extend some way into the back-barrier environment (Andrews, 1970; Cooper and Mason, 1986), constituting a significant sediment source (Bartberger, 1976). An explanation for this difference is provided by Carter and Orford (1984) who suggest that swash percolates into coarse grained barriers, and, thus, any entrained material is deposited, enlarging the barrier and, therefore, reducing the possibility of washover.

In considering the processes that influence lagoon evolution, Cooper (1994) identified that in some cases single processes may dominate, whereas in other situations a number of processes may be of differing importance. Some of the processes identified as influencing barrier and back-barrier sedimentation are: segmentation, fluvial delta progradation, chemical precipitation and lateral accretion.

Vertical accretion in the back-barrier environment is controlled by the rate and nature of sediment supply. The potential sources of sediment are the land, sea or biogenic sediments which form in the lagoon. Deposition tends to be of silts and clays in quiet water conditions, possibly aided by flocculation of fluvial water entering the more saline lagoon (Hobday, 1976). Increased sedimentation can also be caused by anthropogenic activity (Finklestein and Ferland, 1987). Marine sedimentation in back-barrier environments can occur *via* tidal delta progradation, barrier overwash or deposition from suspension. Barrier overwash is suggested to be associated with storm events, and may introduce large quantities of sediment into the back-barrier lagoon (Andrews, 1970).

Carter *et al.* (1989) stated that they were unable to identify a generalised process of barrier and back-barrier lagoon evolution. They suggested that such a generalised process may exist, but that it is masked by local-scale influences. Cooper (1994) suggests, therefore, that lagoon or back-barrier evolution is the result of the balance between the processes acting to reduce or increase the size of the lagoon, and that variations in the relative importance of the evolutionary processes may, in turn, be influenced by macro- and micro-scale controls. Examples of macro-scale controls are; sea-level changes and climatic and tectonic stability, whereas examples of micro-scale controls are sediment supply, longshore drift, coastal morphology, wave energy and tidal range.

2.2.1 A Record of Sea-level Change from Barrier / Back-Barrier Environments?

It was noted by Carter (1988) that most barrier and back-barrier environments exist due to Holocene sea-level rise and that these systems are good locations for the palaeoenvironmental reconstruction regarding barrier and back-barrier relationships, and the influence of sea-level change on their evolution. Jennings *et al.* (1995) suggested that barrier and back-barrier environments could be viewed as two systems coupled through an energy gradient and sediment transport networks. Clearly, the two systems are inter-related. Indeed, Orford *et al.* (1991) stated that barriers have a strong influence on the depositional environments in the back-barrier, and that the back-barrier environment was subject to control by the seaward barrier. In addition, Orford *et al.* (1991) proposed that sediments deposited behind a gravel barrier may confuse sea-level reconstruction. Orford *et al.* (*op. cit.*) suggest that during the Holocene, the initiation and breakdown of gravel barriers has clear implications for the back-barriers sediments that may hold a record of sea-level changes, as they believe that all back-barrier environments are subject to control by the seaward barrier.

It is suggested by some authors (e.g. Duffy *et al.*, 1989; Shaw and Forbes, 1987) that a gravel barrier may exert such control on the back-barrier environment that variation in the marine and freshwater influence may be recorded independent of sea-level tendency, and, thus, may present a problem for sea-level reconstruction. However, Rampino and Sanders (1981) believe that the most important variable in determining the evolution of back-barrier sediments is the rate of sea-level rise.

It was identified by Long *et al.* (1996) that, in recent years, much research has been completed concerning the morphodynamics of gravel barriers over a variety of timescales (Roy *et al.*, 1980; Carter and Orford, 1984, 1993; Carter *et al.*, 1987; Carter *et al.*, 1989; Carter *et al.*, 1992; Forbes *et al.*, 1995; Jennings *et al.*, 1993; Orford *et al.*, 1991; Orford and Carter, 1995). However, Long *et al.* (*op. cit.*) demonstrate that much of this research has focused on the behaviour of gravel barriers in the mesoscale (1-100 years). Some investigations of long-term or megascale gravel barrier dynamics are described in the studies outlined above and also by Carter and Orford (1993), Jennings and Smyth (1990), Boyd and Penland (1984).

Mason and Jordan (1993), and Schulmeister and Kirk (1993). It is suggested by Long *et al.* (1996) that the reconstruction of megascale gravel barrier dynamics is complicated by a number of factors that either cannot be quantified or that are irrelevant to shorter time periods. In the Romney Marsh area, for example, much of the evidence of former gravel barriers is not preserved, as older ridges have been destroyed or re-worked. In addition, the dating of barrier sediments is problematic and any dates obtained may bear an equivocal age relationship to the formation of the barrier.

As a consequence of the problems experienced in investigating gravel barriers over the megascale time period, many workers have utilised the back-barrier sediments simply as a record of the evolution of the barrier (Long *et al.*, 1996). Examples of this are the studies of Carr and Blackley (1973) at Chesil Beach, Jennings and Smyth (1990) from the East Sussex coast and Long and Innes (1993, 1995a&b) and Long *et al.* (1996) from the Romney Marsh area. Indeed, it has been suggested by Carter *et al.* (1989) that the most extensive gravel barrier in the United Kingdom is the Dungeness Foreland, and that this system provides the best site in the United Kingdom to examine the inter-dependence of gravel barriers and tidal back-barrier sedimentation.

Some models of barrier development do exist. Orford *et al.* (1991) noted that coastal geomorphologists have paid relatively little attention to coarse clastic barriers, and that as a consequence, it is not clear how gravel barriers were initiated, stabilised and eventually broken down. By studying both drift- and swash-aligned barriers from the paraglacial coast of east Canada, Orford *et al.* (1991) presented a three phase model for gravel barrier evolution; the initiation phase, the established phase and the breakdown phase. The established phase is characterised by an influx of sediment to the coastal zone. Once sedimentation has begun, the barrier becomes enlarged *via* distal extension and thickening. In order that the established phase should occur, a continued or increasing supply of sediment is necessary, without which updrift reworking of the barrier may occur. The breakdown phase is suggested to involve the segmentation and re-orientation of the barrier into a new structure, which may be caused by a reduction in the sediment supply and / or an increase in the rate of sea-level rise or storm incidence in drift-aligned barriers.

Long and Innes (1995a) suggest that the simple three phase model of barrier evolution (Orford *et al.*, 1991) should leave a predictable pattern of sedimentary and vegetational changes in the back-barrier sediments. However, Long and Innes (*op.cit.*) acknowledge that it may be difficult (if not impossible) to link changes in the back-barrier environment to changes in barrier conditions due to the influence of other processes, such as the migration of back-barrier tidal channels, changes in sediment supply or sea-level changes. The three phase model of barrier evolution was, thus, utilised by Long and Innes (1995a) as a point from which to examine the linkage between back-barrier and barrier sedimentation in Romney Marsh. It was concluded that the three phase model was identifiable from the back-barrier sediments of Romney Marsh. However, they suggest that the system was more complex than the simple tripartite model due to the complex interplay of local and regional factors controlling both gravel barrier dynamics and also back-barrier sedimentation.

Long and Innes (1995a) have utilised the conclusions of Orford *et al.* (1991) to suggest that the emplacement of a gravel barrier in the Romney Marsh area would be recorded in the back-barrier sediments. Long and Innes suggest that these changes would be a reduction in the marine influence, continuing upward and then eventually a return of marine conditions as the barrier degraded. However, it was illustrated that this model is an oversimplification and that vegetational and sedimentary changes may be caused by other environmental changes, such as changes in sea-level, the juxtaposition of tidal channels or sediment supply. Therefore, studies that have attributed a sedimentary change to changes in the barrier merely due to the existence of a seaward barrier may be erroneous. It was also identified by Long and Innes (*op. cit.*) that the sedimentary variability observed in Romney Marsh, *i.e.* two peat units in the back-marsh and only one in the fore-marsh, demonstrates that the sedimentary history is more complex than the three phase model proposed by Orford *et al.* (1991). It is suggested that such variability is to be expected due to the interplay of local (gravel barrier) influences and regional (sea-level change) influences.

2.2.2 Models of Barrier / Back-barrier Evolution.

A number of authors (Kjerfve, 1986; Nichols, 1989; Kjerfve and Magill, 1989) have attempted to assess lagoonal variability and to place lagoons within a conceptual framework. Nichols (1989) proposed that three types of lagoon exist and assessed the relative importance of sedimentation rate and sea-level change on lagoonal evolution. Nichols termed the first of these lagoon types is a deficit lagoon, in which the rate of sea-level rise exceeds the rate of sedimentation and, therefore, the lagoon deepens. The opposite is recorded in a surplus lagoon, with sediment supply exceeding the rate of sea-level rise and, thus, the lagoon is infilled. Nichols (*op. cit.*) termed the third type of lagoon an equilibrium lagoon, in which the morphology was maintained by a balance of sea-level rise and sediment supply. However, it is suggested that many case studies from back-barrier environments show sedimentary variation and, therefore, no single sedimentary model can be applied to all circumstances.

A problem that exists in attempting to model the evolution of back-barrier sediments is that in many studies the barrier and the back-barrier environment are treated

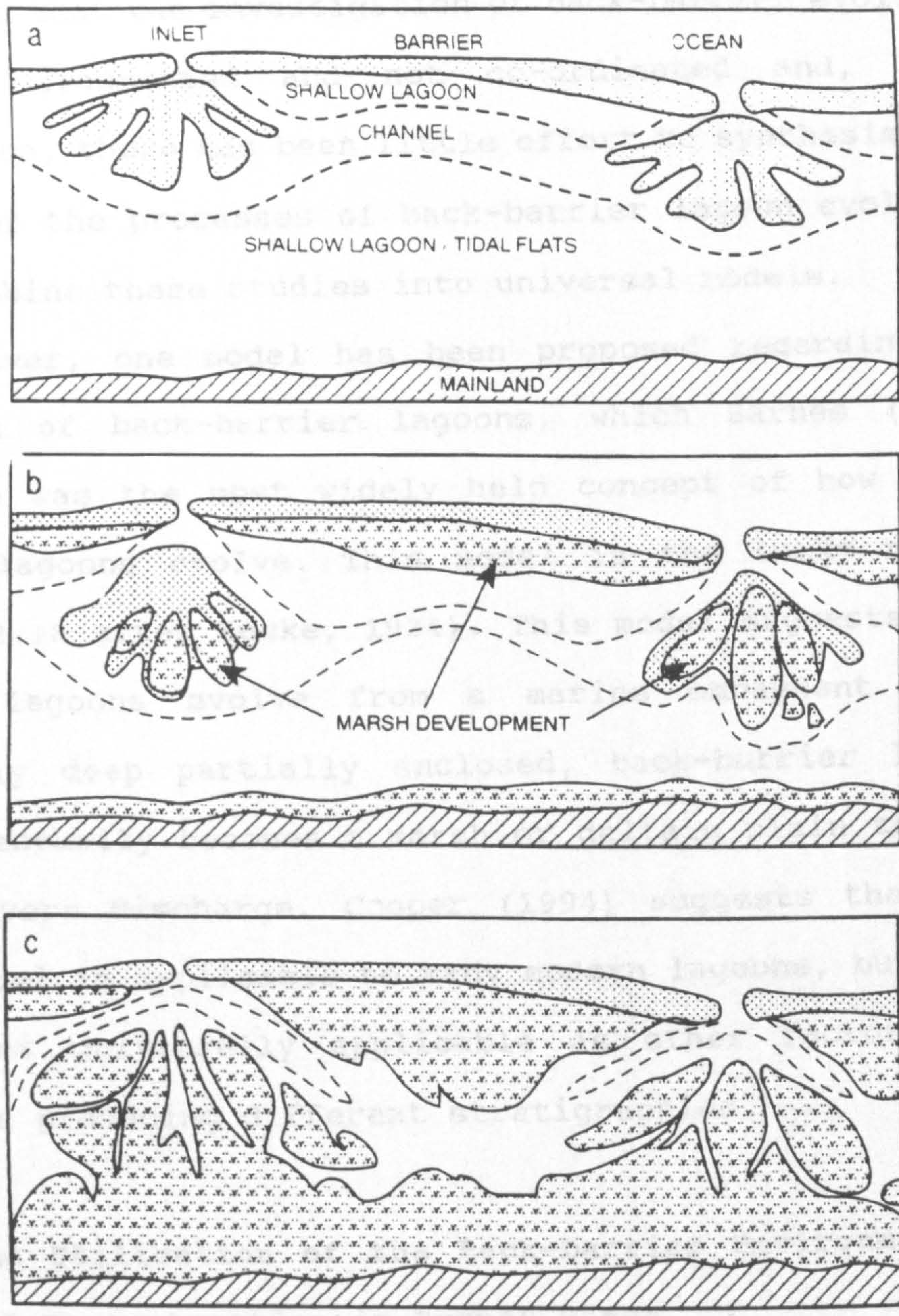


Figure 2.18: A summary diagram of the 'Lucke model' of lagoon evolution.

Source: Cooper (1994), after Oertal *et al.* (1989).

separately (Nichols, 1989). In addition, Cooper (1994) has identified that the investigation of back-barrier evolution has been fragmented and not co-ordinated and, as a consequence, there has been little effort to synthesise the studies of the processes of back-barrier lagoon evolution or to combine these studies into universal models.

However, one model has been proposed regarding the evolution of back-barrier lagoons, which Barnes (1980) suggested was the most widely held concept of how back-barrier lagoons evolve. This model is the Lucke model, (figure 2.18 after Lucke, 1934). This model suggests that coastal lagoons evolve from a marine embayment to a relatively deep partially enclosed, back-barrier lagoon which eventually becomes a marsh or deltaic plain through which rivers discharge. Cooper (1994) suggests that the Lucke Model is applicable to many modern lagoons, but that it is not universally applicable as other factors may intervene producing different stratigraphies.

2.2.3 The Utilisation of the Back-barrier Environment of Sea-level Reconstruction in Romney Marsh.

In spite of the apparent problems, a number of authors have utilised the back-barrier sediments of the Romney Marsh area in order to reconstruct past sea-levels (Tooley and Switsur, 1988; Long and Innes, 1993, 1995a&b; Long et *al.*, 1996). Long *et al.* (1996) noted that the sediments of the Romney Marsh area record the widespread removal of marine conditions through much of the mid-Holocene. This contrasts with the pattern of coastal sedimentation in other, smaller back-barrier regions of southern England, for example, Coombe Haven (Jennings and Smyth, 1987) Willingdon Levels (Smyth and Jennings, 1990) and Stansore Point (Long and Tooley, 1995). Equally, this pattern of coastal sedimentation differs from that recorded by Devoy (1979) on the more open coastline of the Thames Estuary. Long *et al.* (1996) suggest that these differences in the record of coastal sedimentation indicate that the size of a barrier may influence its ability to adapt to changes in sediment supply or sea-level rise.

3 Methodology.

In this chapter the methodologies utilised in the palaeoenvironmental reconstruction of this study will be outlined. A brief background to each of the techniques will be presented followed by an explanation of the methodology used in the sampling and the analysis the sediments. In addition, the approach utilised in the interpretation and presentation of the data will be outlined. Table 3.1 illustrates which techniques of palaeoenvironmental reconstruction have been applied to each of the typecores sampled from Scotney Marsh.

Typecore / Analysis	AY17	A-B27	G60	AW63	AW-AX67
Lithostratigraphy	*	*	*	*	*
Diatoms	*	*	*	*	*
Pollen	*	*	*	*	*
Particle Size Analysis	*		*		
Environmental Magnetism	*		*		
Radiocarbon Dating	*	*	*	*	*

Table 3.1 : Analyses completed on the Scotney Marsh typecores.

3.1 Lithostratigraphy.

The recording of the characteristics of the sediments sampled, both in the field and the laboratory, was completed using the classification scheme of Troels-Smith (1955). This scheme has been widely applied to the study of coastal marshlands in Romney Marsh, e.g. Tooley and Switsur (1988) and Long and Innes (1995a), and also in other coastal lowland areas, e.g. Shennan (1980; 1986a) and Tooley (1978). Therefore, this scheme has the advantage of providing a consistency in approach to the recording of the sedimentary characteristics. Another advantage of the Troels-Smith sediment classification scheme is that it is objective and, therefore, does not necessitate any interpretation. In the field, in addition, the features recorded from the sediment encourage the analyst to investigate the sediment at close hand and, thus, to obtain a detailed understanding of the sediments component parts and structure.

3.1.1 Field Stratigraphy.

The preliminary stratigraphic data from the Scotney Marsh area were obtained *via* the establishment of a sampling grid with boreholes sunk at 25m intervals. Indeed, Shennan (1986) has suggested that a borehole density of at least 30m may be required to avoid errors of +/- 0.30m in recording transgressive and regressive stratigraphic contacts. To determine the detailed morphology of the basal gravel an additional borehole was sunk between points on the sampling grid (at a 12.5m interval) where the differential altitude of the basal gravel was >2.00m. The orientation of the sampling grid was established with the long, south-west to north-east axis parallel to Jury's Gut Sewer.

In contrast to the above, a larger sampling interval was utilised in completing stratigraphic transects I and II, which established the relationship between the site-specific work at Scotney Marsh and other studies from within Romney Marsh. Stratigraphic transect I was sampled at 100m intervals between Denge Marsh (TR058193) and Brookland Church (TQ990257), whereas stratigraphic transect II was sampled at 100m and 150m intervals dependent on the sediments encountered between Guldeford Lane Corner (TQ959228) and Hammonds Corner, New Romney (TR052248).

The borings were completed by hand augerings. A 3cm diameter gouge auger was used to obtain samples for stratigraphic analysis. Following the completion of the field stratigraphic work in Scotney Marsh, specific areas were identified to be studied in greater detail. This was achieved *via* the collection of 'type' boreholes for laboratory-based granulometric, micropalaeontological, sediment source and chronostratigraphic analysis. These boreholes were collected using a 5cm diameter piston corer.

All boreholes and type boreholes were levelled to Ordnance Datum (OD) Newlyn from benchmarks in the local area.

3.1.2 Diagrammatic Presentation of Lithostratigraphy.

The presentation of the stratigraphic transects was completed by utilising the computer package TSPPlus (see

Waller *et al.*, 1995). The transects selected for presentation here are those from which the typecores were selected, primarily because these sediments are characteristic of the sedimentary sequences recorded in Scotney Marsh, but also because they provide the necessary stratigraphic context for the typecores. These transects trend perpendicular to the general trend of the gravel ridges that are orientated approximately south-west to north-east.

In addition to the perpendicular sections, a longitudinal transect was selected, trending parallel to the gravel ridges of Scotney Marsh, in order to establish any stratigraphic variability throughout the length of the study area. The stratigraphic transects I and II were also presented using TSPPlus.

To assist with the visual presentation of the stratigraphy of Romney Marsh a series of schematic diagrams have been completed that represented all of the stratigraphic work that has been completed on the marsh. In these diagrams the areas in which peat was encountered are indicated by black shading, whereas the areas in which the channel sediments were recorded are represented by a stippled shading.

3.1.3 Three-dimensional Representations of the Lithostratigraphy.

The lithostratigraphic data from Scotney Marsh were utilised to produce three-dimensional figures of the buried gravel and peat surfaces. The altitude of the buried gravel surface was plotted utilising the computer package SURFER. To plot the peat surface, the altitude of the upper contact of the peat unit was used. Where the peat unit was absent, the contact between the underlying blue-grey silts and the overlying oxidation mottled silts was used as it marks the discontinuity between the lower and upper sedimentary facies. Three-dimensional surfaces for the buried gravel and peat surfaces were produced for the area surrounding each of the palaeoenvironments from which the typecores were selected.

In addition, the stratigraphic data from the whole of Scotney Marsh were used to produce a contour diagram of the gravel surface. This was carried out to demonstrate the gravel morphology of the entire Scotney Marsh.

3.2 Diatom Analysis.

Diatom analysis can provide precise information regarding the sedimentary conditions at a site, from both within organic units (where pollen analysis can also be utilised) and, more importantly, minerogenic sediments. Alderton (1994) has identified three characteristics that make diatoms a useful tool for palaeoenvironmental analysis. First, diatoms occur in a wide range of environments, e.g. from freshwater conditions to saline springs, and therefore have a wide salinity range; secondly, diatoms are readily preserved and, therefore, are generally present in sediments; and thirdly, due to the minute size and frequent abundance of diatoms, only a relatively small sample is required to complete the necessary analysis.

3.2.1 Diatom Sample Selection.

Diatom analysis was carried out on all of the typecores from Scotney Marsh. A complete profile, with sampling at regular intervals coupled with finer resolution sampling resolution across the transitions between organic and minerogenic strata, was completed for typecores AY17 and G60. However, a more limited sampling strategy was completed for typecores A-B27, AW63 and AW-AX67. In these cores, the sediments from above, below and within the peat units were analysed, again with a finer sampling resolution across the stratigraphic transitions.

3.2.2 Technique of Diatom Analysis.

Each of the samples selected for diatom analysis comprised of ca. 0.5cm^3 of sediment, which was digested *via* heating in a 20% solution of 100 volume hydrogen peroxide (H_2O_2). An aliquot was evaporated onto a cover slip and then mounted in Naphrax (refractive index 1.74). Microscopic analysis was completed on the aliquots at magnifications of x400 and x600 using a Nikon Laboplot microscope. Diatoms were identified with reference to van der Werff and Huls (1958-74) and Hendy (1964).

Minimum counts of 200 diatom valves were completed. However, when one species type (usually *Paralia sulcata*) constituted >50% of the total diatom count, then at least 100 other diatom valves were counted. In addition, at least 20 different species types were counted for each aliquot. For this study, the salinity tolerance of the diatoms was recorded using the scheme established by van der Werff and Huls (1958-74).

3.2.3 Graphical Representation of the Diatom Analysis.

The data from the diatom analysis were plotted against sample altitude utilising the TILIA computer package (Grimm, 1991), with species types and salinity groups expressed as a percentage of total diatom valves (TDV). The resulting diagram was then sub-divided into local diatom assemblage zones (LDAZ) on the basis of core sections of relative consistent trends of diatom characteristics.

3.2.4 Interpretation of Diatom Analysis.

In interpreting the results of diatom analysis, the schemes of Vos and de Wolf (1988 and 1993) were utilised. These schemes classify coastal diatoms into ecological groups, and relate these groups to specific sedimentary environments. In addition, the work of Denys (1994) was also used to assist in the interpretation of the diatom assemblages. This study assigned specific diatom species to particular coastal environments, *i.e.* lower mudflat to fen environment.

It was noted by Vos and de Wolf (1988 and 1993) that the relative proportions of autochthonous diatoms (species that have lived at the place of deposition) and allochthonous diatoms (species that have been transported from elsewhere to the place of deposition) were a major problem in palaeoenvironmental diatom research in tide-influenced coastal areas. Therefore, an attempt was made in this study to quantify the contributions made by autochthonous and allochthonous assemblage components.

In undertaking the above investigation a number of methodologies were adopted. First, the percentages of epiphytic / benthic and planktonic diatoms relative to TDV were calculated. This was completed as planktonic diatoms are, by definition, allochthonous (Simonsen, 1969). Secondly, an attempt was made in Beyens and Denys (1982) to quantitatively distinguish between autochthonous and allochthonous diatoms. This approach has been adapted and utilised in this study.

Vos and de Wolf (1988 and 1993) believed that the model of Beyens and Denys (1982) was too simplistic as it ignored both the epiphytic and any euryhaline taxa and only utilised benthic taxa. In this study an attempt has been made to improve on the method of Beyens and Denys (1982) by including both epiphytic and benthic diatoms together. The salinity group from the classification of van der Werff and Huls (1958-74) with the greatest number of epiphytic / benthic diatoms (as a percentage of TDV) is the 'optimal diatom group'. This optimal diatom group also includes the percentages (of TDV) of the two neighbouring salinity groups. The percentages of epiphytic / benthic diatoms to the freshwater 'side' of the optimal diatom group were added together and divided by the optimal diatom group to produce a ratio, *i.e.* the fresh allochthonous ratio. Similarly the epiphytic / benthic diatoms to the marine 'side' of the optimal diatom group were calculated to produce the marine allochthonous ratio.

For example -:

optimal group : brackish-fresh		25.8%
neighbouring groups :		
fresh-brackish		+ 19.4%
brackish		+ 18.7%
autochthonous component		= <u>63.9%</u>
Fresh allochthonous group : fresh		12.4%
Fresh allochthonous ratio	12.4 / 63.9	= <u>0.194</u>
Marine allochthonous groups :		
brackish-marine		9.4%
marine-brackish		+ 8.1%
marine		+ 6.2%
		= 23.7%
Marine allochthonous ratio	23.7 / 63.9	= <u>0.371</u>

Finally, the criteria described by Vos and de Wolf (1988, 1993) as used in palaeoecological diatom studies by the Diatom Department of the Geological Survey of the

Netherlands, were also considered in order to assess the allochthonous and autochthonous diatom contributions. This methodology considers a number of diatom-related criteria, *i.e.* the occurrence of relatively rare taxa, and non-diatom- related criteria, *i.e.* palaeogeographic location.

3.3 Pollen Analysis.

The organic sediments from the typecores were sampled for pollen analysis, which is a standard method for the study of vegetational history (Waller, 1994b). The utilisation of pollen analysis can provide information on the past vegetation at or close to the site, and also may provide a regional pollen signal. Waller (1994b) has suggested that fen and marsh sites derive much (almost certainly most) of their pollen from the local environment. As a consequence, Waller suggests that it should be possible, with some confidence, to reconstruct the local vegetation and, thus, draw inferences concerning the nature of local environmental conditions, e.g. the proximity of a site to the sea. A general account to the background and the application of pollen analysis is provided by Moore and Webb (1978).

3.3.1 Pollen Sample Selection.

Aliquots were selected for pollen analysis from the organic-rich units (peats and peaty-clays) from the typecores selected from Scotney Marsh. Initially these units were sampled at regular intervals, but with a finer resolution sampling at stratigraphic transitions to provide a skeleton pollen diagram, as suggested by Moore and Webb (1978). Following this initial analysis, further sampling was completed at the levels that required increased detail to resolve the palaeovegetational history of the typecore site.

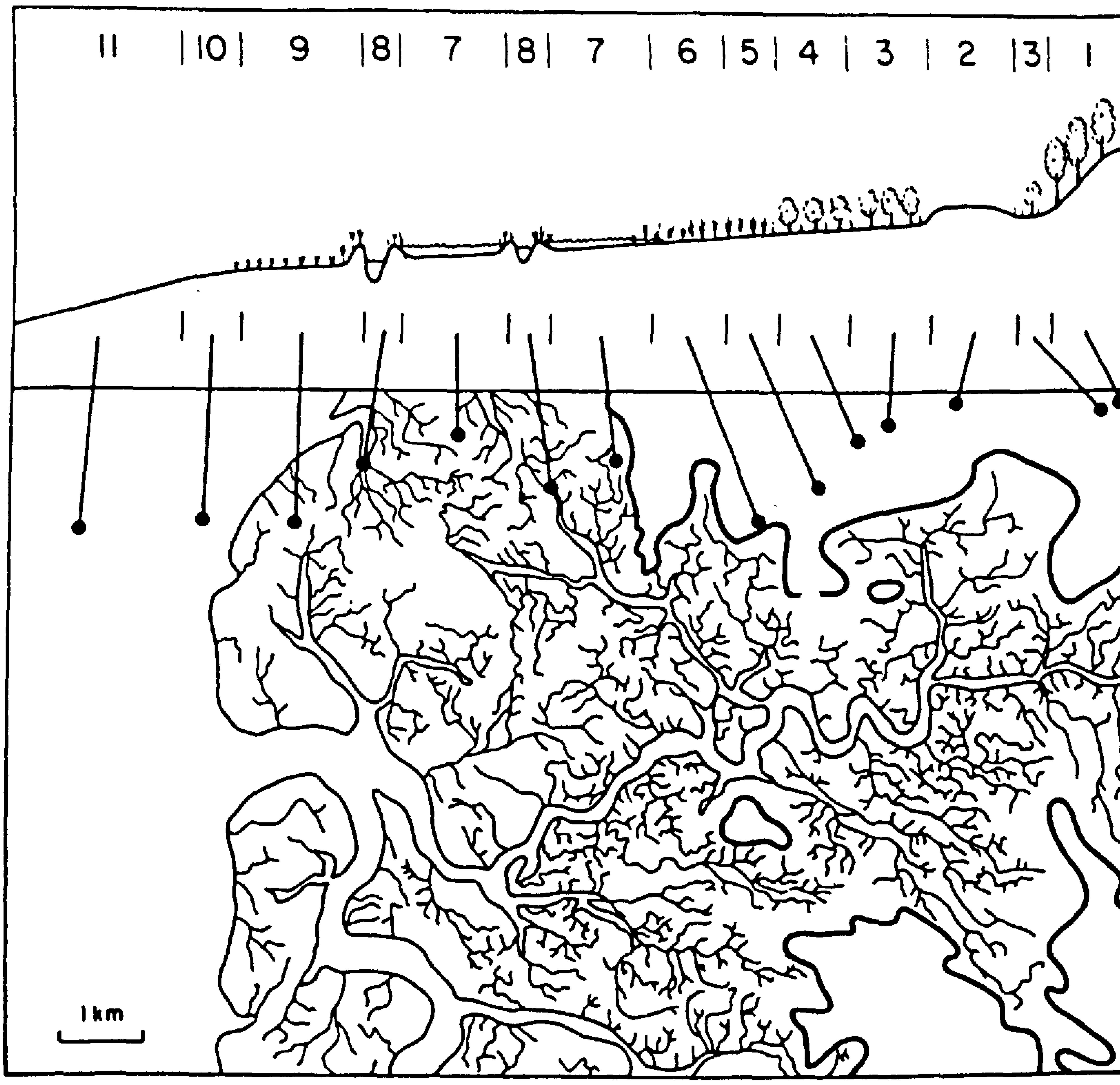
2.3.2 Techniques of Pollen Analysis.

The aliquots selected for pollen analysis, *i.e.* ca. $1.0cm^3$, were prepared in the laboratory for microscope analysis. The samples were sieved through a $125\mu m$ sieve, following heating in a 10% solution of sodium hydroxide (NaOH). The samples were centrifuged, decanted and washed in a 10% hydrochloric acid solution. The soluble cellulose was removed by acetolysis. First, 10ml of glacial acetic acid was added, centrifuged and decanted. Following this, a solution of 9ml of acetic anhydride and 1ml of concentrated sulphuric acid was added to each sample and then heated in a water bath. After two minutes in the water bath, 40ml of glacial acetic acid was added and the samples were centrifuged and decanted. Additional washes of double distilled water and 10% sodium hydroxide were also completed. The samples were then stained with safranine and mounted in glycerol.

The counting of the pollen was completed using a Nikon microscope at magnifications of x400, x600 and, where necessary, x1000 with immersion oil. In an attempt to eliminate error due to the non-random distribution of pollen grains on a slide, at least two slides were counted at regularly spaced traverses, as suggested by Waller (1994b). The pollen grains were identified using the keys of Moore and Webb (1978) and also the reference collection at the University of Liverpool. A minimum of 200 land pollen grains were counted for each aliquot and, where one species was particularly dominant and comprised >50% of total land pollen (TLP), at least 100 land pollen grains of other species were counted.

3.3.3 Graphical Representation of the Pollen Analysis.

The results of the pollen analysis were presented in a graphical form utilising the computer package TILIA, as for the diatom analysis. The totals of the pollen species and pollen groups (trees, shrubs, herbs, aquatics, pteridophytes) were expressed as a percentage of total land pollen (TLP), this excludes ferns, spores and aquatic pollen types. The pollen diagrams were sub-divided into local pollen assemblage zones (LPAZ) containing zones of relatively consistent trends of pollen and spore assemblages. This method of sub-division was selected for this study as many of the coastal vegetational assemblages are by their nature transitional, and it is these transitional changes, *i.e.* from saltmarsh to coastal reedswamp, that are identified as distinct zones. The alternative methods of sub-division, *i.e.* CONISS (Grimm, 1987), would not have determined the somewhat subjective, transitional zones as described above.



1. Upland or regional forest.
2. Raised bog.
3. Oak fen woodland.
4. Alder carr.
5. Sedge fen.
6. Coastal reedswamp.
7. Inter-creek area.
8. Creek.
9. Saltmarsh.
10. Mudflat.
11. Intertidal sandflat.

Figure 3.1: Sedimentary and vegetational succession utilised in this study, adapted from Shennan (1986). Source: Waller 1995.

3.3.4 Interpretation of Pollen Analysis.

It has been noted by Waller (1994b) that theoretical accounts of pollen analysis tend to concentrate on more conventional 'basins', *i.e.* lakes and bogs. Consequently, fenland or marshland, such as the deposits from Scotney Marsh, are rarely considered. The pollen encountered in a particular deposit does not necessarily directly represent the vegetation of the source area. Waller (1994b) has illustrated that this is due to a number of factors; some taxa demonstrate a higher pollen frequency than their abundance in the vegetation, equally other pollen types are scarce and even absent from the pollen assemblage. In addition, a dispersal bias of the pollen may exist. Waller (*op. cit.*) also identifies that the lack of taxonomic precision in pollen identification also creates a problem for the reconstruction of fenland and marshland vegetational histories. This is because it is rare that a pollen grain can be identified such that the environment of origin can be unequivocally determined.

In this study, the vegetation zonation presented by Waller (1994b) was utilised to assist in the interpretation of the pollen data from Scotney Marsh (figure 3.1). This model describes the typical succession of vegetation types from the unvegetated tidal mudflats to saltmarsh and eventually reaching fen woodland to upland vegetation.

3.4 Particle Size Analysis.

Friedman and Sanders (1978) note that the movements of air and water commonly separate particles according to their size. The frequency distribution of particle sizes in sediments relates primarily to two factors; first, the availability of different particle sizes in parent material, and secondly, processes operating at the site of deposition, particularly competence of flow (Friedman and Sanders, 1978). Statistical analysis of the particle size distributions leads to a better understanding of the distribution in terms of physical processes. Detailed accounts of the background to the study of particle sizes are provided by Folk (1966), Friedman and Sanders (1978) and Leeder (1982), whereas, Postma (1967) discusses the sedimentary characteristics of various estuarine systems.

3.4.1 Particle Size Analysis: Sample Selection.

Aliquots were selected for particle size analysis from two of the typecores (cores AY17 and G60). It was decided to select these aliquots from the same levels that were utilised for mineral magnetic analysis. This decision was taken as it has been demonstrated by Long *et al.* (1996) that data from the two techniques combined can provide enhanced detail on sediment source and processing for palaeoenvironmental reconstruction.

3.4.2 Technique of Particle Size Analysis.

Each aliquot was digested in a 20% solution of 100 volume hydrogen peroxide (H_2O_2) and heated to remove any organic matter from the sediments. The hydrogen peroxide was then boiled off and the samples cooled before calgon was added and the sample disaggregated in an ultrasonic bath. Before each sample was analysed, it was again placed in an ultrasonic bath and then decanted into a round bottomed glass beaker with a side baffle, and stirred with a mechanical stirrer to maintain the sediment in suspension prior to sampling.

The samples were then analysed utilising a Coulter LS130 laser sizer. The principles and the uses of this method of particle size analysis have been described by Harfield (1982), and the validity of the use of this technique has been discussed by Wass (1987). Wass (1995) suggests that the use of an electronic Coulter Counter is particularly suitable for small samples of fine-grained sediments.

To determine the particle size of each sample a number of pipetted sub-samples were taken from the round bottomed beaker, containing the sediment in suspension, and run through the Coulter LS130. This process was completed at least twice for each sample to avoid any unrepresentative results that can be generated due to the presence of air bubbles that can occur within the apparatus during analysis. From the data produced *via* the laser analysis, a cumulative frequency curve was produced and the percentiles required for statistical analysis assessed *i.e.* phi (ϕ) sizes at 5%, 16%, 25%, 50%, 75%, 84%, 95%. The statistical parameters calculated from these percentiles were: mean grain size, standard deviation (sorting), skewness and kurtosis, as proposed by Folk (1974) and presented below.

$$\text{Mean grain size}: \frac{16\% + 50\% + 84\%}{3}$$

$$\text{Standard deviation}: \frac{84\% - 16\%}{4} + \frac{95\% - 5\%}{6.6}$$

$$\text{Skewness}: \frac{(84\% + 16\% - 2\text{x}50\%)}{2(84\% - 16\%)} + \frac{(95\% + 5\% - 2\text{x}50\%)}{2(95\% - 5\%)}$$

$$\text{Kurtosis}: \frac{(95\% - 5\%)}{2.44(75\% - 25\%)}$$

Within this study ϕ units are utilised, in preference to mm or μm, as they are based on a logarithmic scale, *i.e.* $\phi = \log_2 d$ (d = diameter of grain in mm, Krumbein, 1939). This scale was proposed due to the fact that the comparison of particle size populations necessitates the consideration of a wide range of grain sizes and, thus, the utilisation of a geometric scale rather than an arithmetic one.

The **mean grain size** of the particle size distribution provides an indication of the energy required for sediment transportation. The larger the mean grain size, the higher the energy of the environment of deposition.

The **standard deviation (sorting)** provides a measure of the degree of uniformity that was produced by sediment processing. For example, the continued winnowing of sediment by wave action would tend to reduce the standard deviation, thus, improving the sorting. Conversely, deposition of fines at the slack water period of the tidal cycle tends to lead to poorly sorted sediments. The values for the sorting of the samples were derived utilising the scale developed by Folk (1974).

The **skewness** of a sample is used as it is extremely sensitive to the sample's depositional environment as it indicates the preferential addition or removal of grain sizes at the coarse and fine 'tails' of the particle size distribution. An illustration of this is the winnowing away of finer grains *via* the processing of the sediment by wave action, which tends to leave an excess of coarse particles. The resulting sediment is, therefore, coarse- or negatively-skewed. To evaluate the skew of a sample's particle size distribution, the scale of Folk (1974) was utilised.

The **kurtosis** of a sample provides a measure of the 'peakedness' of a particle size distribution. Kurtosis, therefore, provides another measure of the degree of sediment processing. The classes used to determine the kurtosis of the particle size distribution of each sample are those outlined by Folk (1974).

3.4.3 Graphical Presentation of Particle Size Analysis.

The results of the statistical analysis of the particle size data are presented graphically using the 'bplot' computer programme, developed at the University of Liverpool.

3.4.4 Interpretation of Particle Size Analysis.

In interpreting the particle size data, the classes described by Folk (1974) from the statistical calculations completed on the particle size distributions were utilised. Comparisons were also made to similar studies of tidal sediments, e.g. Postma (1961, 1967) and Plater (1992), and also the discussions of particle size characteristics provided by Freidman and Sanders (1978) and Leeder (1982). In addition, the statistical data was applied to the model presented by Tanner (1991) which utilises mean grain size and skewness to evaluate the environment of deposition of the sediments.

3.5 Mineral Magnetic Analysis.

In this study the technique of environmental magnetism is utilised. This measures the response of sediments to a range of artificially applied magnetic fields in an attempt to determine the source of the marsh sediments. Indeed, it has been demonstrated (Walling *et al.*, 1979; Oldfield *et al.*, 1985a; Yu and Oldfield, 1989; 1993) that mineral magnetic analysis is an effective method in sediment source determination.

3.5.1 Mineral Magnetic Properties.

A number of magnetic measurements are used to investigate the magnetic properties of samples. These measurements are outlined below.

Magnetic Susceptibility (χ) - This is a measure of the degree to which a sample can be magnetised (Thompson and Oldfield, 1986). Susceptibility tends to be dominated by the presence of ferrimagnetic minerals, *i.e.* magnetite, and is roughly proportionate to the concentration of magnetic minerals. Magnetic susceptibility is the ratio of the magnetisation in any sample to the intensity of the magnetising field applied to it. This property is affected by grain size and shape.

Anhysteric Remanent Magnetism (ARM) - This is the remanence produced by the smooth decay of a strong field in the presence of a weak steady field (Thompson and Oldfield, 1986). ARM is sensitive to the concentration of grain sizes of ferrimagnetic minerals in a sample. ARM may also be related to the concentration of magnetic minerals of a finer stable single domain size.

Saturation Isothermal Remanent Magnetism (SIRM) - SIRM is the magnetic remanence attainable by a sample is produced *via* the application of a powerful magnetic field (Thompson and Oldfield, 1986). SIRM can be used in a similar way to

χ, as they are both dependent on the magnetic concentration. SIRM is also strongly influenced by magnetic grain size, the values being highest in the superparamagnetic to stable single domain sizes (Hartstra, 1982).

Soft Isothermal Remanent Magnetism (Soft) - This is the remanent magnetism following magnetisation in a reverse field of -20mT. The soft remanence component is that part of a sample demagnetised by the first and weakest reverse field. This parameter can indicate the importance of magnetite in the magnetic signal (Thompson, 1986).

Hard Isothermal Remanent Magnetism (HIRM) - HIRM is a measure of the remanent magnetism following demagnetisation in field of -300mT. Therefore, the HIRM component is that part of a sample that remains unreversed following the application of the strongest reverse field. This parameter can be used as a measure of the relative importance of haematite and goethite in the magnetic signal of a sample.

Utilising the values obtained for the magnetic measurements listed above, a number of percentages, ratios and quotients are then calculated.

SIRM/χ - Thompson and Edwards (1982) have demonstrated that SIRM/χ is sensitive to the mixture of magnetic minerals and the magnetic grain size with SIRM/χ values varying inversely with magnetic particle size, when the dominant magnetic mineral is magnetite.

χ_{ARM} - This is primarily a measure of the concentration of ferrimagnetic material within a sample, and is strongly influenced by magnetic grain size.

$\chi_{ARM/SIRM}$ - This parameter is suggested by Maher (1988) to be particularly sensitive to magnetic grain size in the stable single domain to multidomain magnetic grain size range.

χ_{ARM}/χ - This parameter can provide an indication of changes in magnetic grain size, particularly when the mineralogy is dominated by magnetite.

Recently the typical detrital magnetic model (which is that the sediments deposited at a site will contain a magnetic signal indicative of their parent material) has been questioned due to the presence of authigenic and diagenetic minerals. Therefore, it has been suggested that the presence of authigenic minerals, e.g. greigite (Fe_3S_4), that form in the sediments post-depositionally, can hinder magnetism based sediment source studies (Oldfield *et al.*, 1989; Snowball and Thompson 1990a; Snowball, 1991). Greigite has been identified from mineral magnetic analyses in a range of sediments e.g. in freshwater muds in contact with sulphide rich marine sediments (Snowball and Thompson, 1988; 1990a) and also in tidal lagoon sediments from Denge Marsh in the Romney Marsh region (Spencer, 1992).

Greigite was first identified by Skinner *et al.* (1964). It is suggested that hydrogen sulphide is released as the organic component of the sediment decomposes and combines with heavy metals to form authigenic metal sulphides (Snowball and Thompson, 1990a). The existence of greigite tends to occur when specific physiochemical conditions, in particular reducing conditions (Hilton, 1990), have been present.

3.5.2 Mineral Magnetic Analysis Sample Selection.

Two of the typecores, *i.e.* cores AY17 and G60, were selected for mineral magnetic analysis. These two cores were selected as the former would be likely to provide information regarding the main back-barrier environment, whereas G60 would probably provide information relating to the Scotney Marsh trough. From these two cores 10ml aliquots were selected at regular intervals and subjected to the complete suite of magnetic measurements.

3.5.3 Mineral Magnetic Measurements.

Each aliquot was packed into a pre-weighed 10ml plastic pot, taking care to ensure that no air remained in the sample. A series of magnetic measurements (high and low frequency susceptibility, ARM, SIRM and IRM backfields) was completed on these wet samples. The samples were then oven mass specific values for the magnetic analysis.

The χ of each sample was measured using a single sample well, dual frequency susceptibility sensor, connected to a Bartington digital meter. To obtain an ARM, each of the samples was placed into an adapted Molspin AF demagnetiser. The ARM is applied to each sample by subjecting it to an alternating field of up to 95mT, which steadily increases and then decreases. In addition, a direct current field of 0.04mT is superimposed on to the sample. For the ARM and the subsequent IRM measurements, a Minispin Slow Speed Spinner Fluxgate Magnetometer was utilised to measure the magnetic remanence of each sample. To apply a SIRM field to each sample, they were placed into a Molspin pulse discharge magnetiser in a field of 1.0T, and the remanence of each sample was then measured. Following the application of the SIRM, a number of backfield IRMs (ranging from -20 to -300mT) were applied and the remanent magnetism measured.

3.5.4 Graphical Representation of the Mineral Magnetic Analysis.

The results of the mineral magnetic analysis were entered into a spreadsheet from which the various quotients and ratios were calculated. Following this, the data were plotted against sample depth / altitude using the 'bplot' computer programme, and the mineral magnetic graphs were produced. Zones of the sediment cores with similar mineral magnetic properties were then identified as mineral magnetic zones.

3.5.5 Mineral Magnetic Interpretation.

By utilising the results of the mineral magnetic measurements, along with the quotients and ratios, described in section 3.5.1, it is possible to determine estimates of the variable mass contributions of magnetic minerals to a sample. Most natural samples utilised in the study of environmental magnetism give magnetic signatures that will reflect the mixtures of magnetic oxides and sulphides, along with changes in magnetic grain size. By the determination of the magnetic signatures of the marsh sediments it should be possible to compare the magnetic signature of the marsh sediments with those of the potential source areas. These potential source areas are the river valleys to the west of Romney Marsh, the Wealden sediment to the north, and the marine sediments to the east and south of Romney Marsh (Spencer, 1992).

3.5.6 Particle Size Separation.

To obtain a truly detrital signal, an attempt was made to obtain marsh sediments devoid of over-printing by authigenic greigite (Fe_3S_4). To do this, groups of sediments from both of the typecores were selected for further analysis. First, sediments believed to contain a high proportion of greigite were selected and combined. Similarly, sediments from each of the typecores believed to contain a relatively low proportion of greigite where also combined. The reason for combining the sediments was to obtain samples of sufficient size so that they may still be measured following the separation of the sediments into separate ϕ size classes.

3.5.7 Method of Particle Size Separation.

Each of the four samples was disaggregated in an ultrasonic bath with 25ml of calgon (2g per 1000ml water) and 25ml double distilled water. These samples were then wet sieved through a 4ϕ sieve and the filtrate decanted into a 500ml sedimentation tube. The $<4\phi$ fraction was sampled for ϕ sizes 5 to 10 by decanting from the sedimentation tube by the pipette method after the appropriate settling times using the Stokes equation. Each of the ϕ fractions was then oven dried, at $<40^0$C, and run through the suite of magnetic measurements as outlined in section 3.5.3.

3.5.8 Interpretation of the Particle Size Separation.

Several mineral magnetic parameters were considered in assessing whether or not the particle size separation was successful in identifying any particle size class in which no greigite was present. These parameters were SIRM/χ, IRM_{100mT} - IRM_{40mT} and the backfield IRMs. It was suggested in Spencer *et al.* (1994) that a value of >40 KAm^{-1} for SIRM/χ in combination with the divergence of the backfield IRMs (with the -20mT and -40mT backfields attaining relatively low values and the -1000mT and -300mT backfields relatively high values) was indicative of the presence of greigite. This is represented in the value of IRM_{100mT} - IRM_{40mT} which is derived from the difference between the -100mT and -40mT backfields. Therefore, a high value for this parameter indicates divergence of the backfield IRMs. The opposite of these conditions, *i.e.* a low (<40 KAm^{-1}) value for SIRM/χ, lack of divergence of the backfields and a low value of IRM_{100mT} - IRM_{40mT}, would suggest that a relatively low contribution was made by griegite to the mineral magnetic signal.

3.6 Chronostratigraphic Analysis.

The application of conventional radiocarbon dating has been used widely in environmental reconstruction, e.g. Shennan *et al.* (1983) and Shennan (1986). This includes the study of coastal lowlands, e.g. Shennan (1982) and Waller (1994b) in the Fens, and Tooley and Switsur, (1988); Long and Innes (1995a) and Long *et al.* (1996) in the Romney Marsh area. The dating of the organic sediments of Scotney Marsh would, therefore, aid comparison of the work at Scotney Marsh with the other palaeoenvironmental research on Romney Marsh and also would enable this work to be placed in the wider context of Holocene sea-level studies in south-east England.

3.6.1 Radiocarbon Analysis: Sampling Strategy.

To establish a chronology for the evolution of Scotney Marsh certain organic samples were selected for radiocarbon dating. In determining which of the units from the typecores should be dated, typecores a consideration of the palaeoenvironments of each site was made. Once regressive and transgressive contacts had been identified in the typecores, units were selected to establish the ages of these regressive and transgressive events in the main back-barrier (cores AY17 and A-B27), the sheltered gravel inlet (cores AW63 and AW-AX67), and also the Scotney Marsh trough (core G60). Consequently, ten units were dated to establish a chronology for the evolution of Scotney Marsh.

3.6.2 Technique of Radiocarbon Analysis.

Approximately 50.0g of organic rich sediment from each of the units selected for radiocarbon analysis was cleaned by removing the outer layer of the typecore and removal of any suspected contamination from adjacent sediment horizons. The sample was then securely wrapped in aluminium foil, clearly labelled and sent for radiocarbon analysis at Beta Analytic Inc., Florida, USA.

3.6.3 Interpretation of the Radiocarbon Dating.

The radiocarbon dates received were presented as calibrated ages in years before present (cal. yrs. BP) using a 2 sigma analytical error (95% probability). Previously some studies from Romney Marsh have presented radiocarbon dates from organic deposits as radiocarbon years, e.g. Tooley and Switsur (1988). In order to compare these sites with the present study, and with other studies that have used cal. yrs. BP, a calibration was required. For this purpose the CALIB computer program of Stuiver and Reimer (1993) was utilised.

3.7 The Methodology Utilised for Time / Altitude Analysis.

The methodology utilised in the construction of time / altitude plots of sea-level change has been described in detail by Tooley (1978a&b, 1982) and also by Shennan (1982, 1986b). It is suggested that each sea-level index point (SLIP) requires five principal attributes (Long and Tooley, 1995): a location, an age, an altitude, an indicative meaning and a reference tidal level. Long and Tooley (1995) state that the indicative meaning of a SLIP refers to the altitudinal relationship between the index point's environment of deposition and a reference tidal-level, e.g. mean high-water of spring tides (MHWST). In addition, Shennan (1982, 1986b) suggests that the indicative range of transgressive and regressive contacts is ca. 0.20m, and that if factors such as sediment compaction or levelling errors are considered, transgressive and regressive contacts may have a typical altitudinal error of ca. +/-0.50m.

Long and Tooley (1995) note that spatial variability in the altitude of MHWST exists and that, as a consequence, the altitude of each SLIP should be reduced to mean sea-level (MSL) by subtracting the present-day height difference between MHWST and MSL from the altitude of each SLIP, which for this study is 4.03m at Dungeness. In addition, it has been suggested by Long and Tooley (*op.cit.*), that changes in palaeotidal range demonstrated in the English Channel (Austin, 1991) are not a source of potential inaccuracy for the study of sea-level changes during the last 7500 cal. yrs. BP, and, thus no correction is necessary.

3.8 Synthesis of Palaeoenvironmental Information.

It has been illustrated by Shennan (1994) that in the study of coastal evolution and relative sea-level change during the Holocene, no formal methodology has been adopted. Consequently, studies have depended largely on inductive inferences and reasonings, where extrapolations are made in an attempt to develop generalisations of sea-level changes. Shennan (*ibid.*) proposes that the method of multiple reasoning hypothesis be utilised, where as many explanations as possible are considered and tested. Along with this approach, Birks and Birks (1980) have proposed two other methodological principles that must be considered; first, methodological uniformatism, and, secondly, that the most simplistic explanation should suffice until such time that further evidence necessitates a more complex explanation. Throughout this study, the above principles have been adopted.

In addition to the principles outlined above, the methodology of synthesis of palaeoenvironmental information utilised in a number of previous studies, e.g. Tooley (1978), Shennan (1986a), Plater (1992) and Long and Innes

(1995a), has been adopted. The results of the diatom, pollen and, where applicable, particle size analyses, along with the geomorphological location of each typecore, are considered in order to complete a site-specific palaeoenvironmental reconstruction. The site-specific palaeoenvironmental reconstructions, in addition to other palaeoenvironmental studies completed in Romney Marsh, are then utilised to attempt a reconstruction of the palaeoenvironment of the Scotney Marsh area, and to formulate a model of the evolution of Romney Marsh. In addition, considerations will be made here of the various processes, e.g. sea-level change, storm magnitude and frequency, sediment budget etc., that have influenced the evolution of Romney Marsh during the Holocene.

4 Results.

In this chapter the results of two extensive stratigraphic transects across Walland Marsh will be presented, as an understanding of the broad stratigraphy across Romney Marsh will assist in determining the most effective location to conduct a detailed site-specific palaeoenvironmental investigation. Following this the results of the site-specific stratigraphic and palaeoenvironmental investigation in the region of Scotney Marsh will be presented.

4.1 Large-scale Stratigraphic Transects : Introduction.

First in this study, an extensive stratigraphic transect was completed across Walland Marsh. This transect ran approximately north-west from Boulderwall Farm on Denge Marsh (TR058193) to Brookland (TQ990257). The completion of this stratigraphic transect established a stratigraphic context for the consideration of studies from Denge Marsh (Plater, 1992; Plater and Long, 1995), Midley (Innes and Long, 1992; Long and Innes, 1993; Long and Innes, 1995b), and Brookland (Long and Innes, 1995a). As the transect also linked up with the stratigraphic transect of Long and Innes (1995a) at Brookland, a large number of the site-specific studies across the marsh were, therefore, stratigraphically integrated. The location of this transect (stratigraphic transect I) is illustrated in figure 4.1.

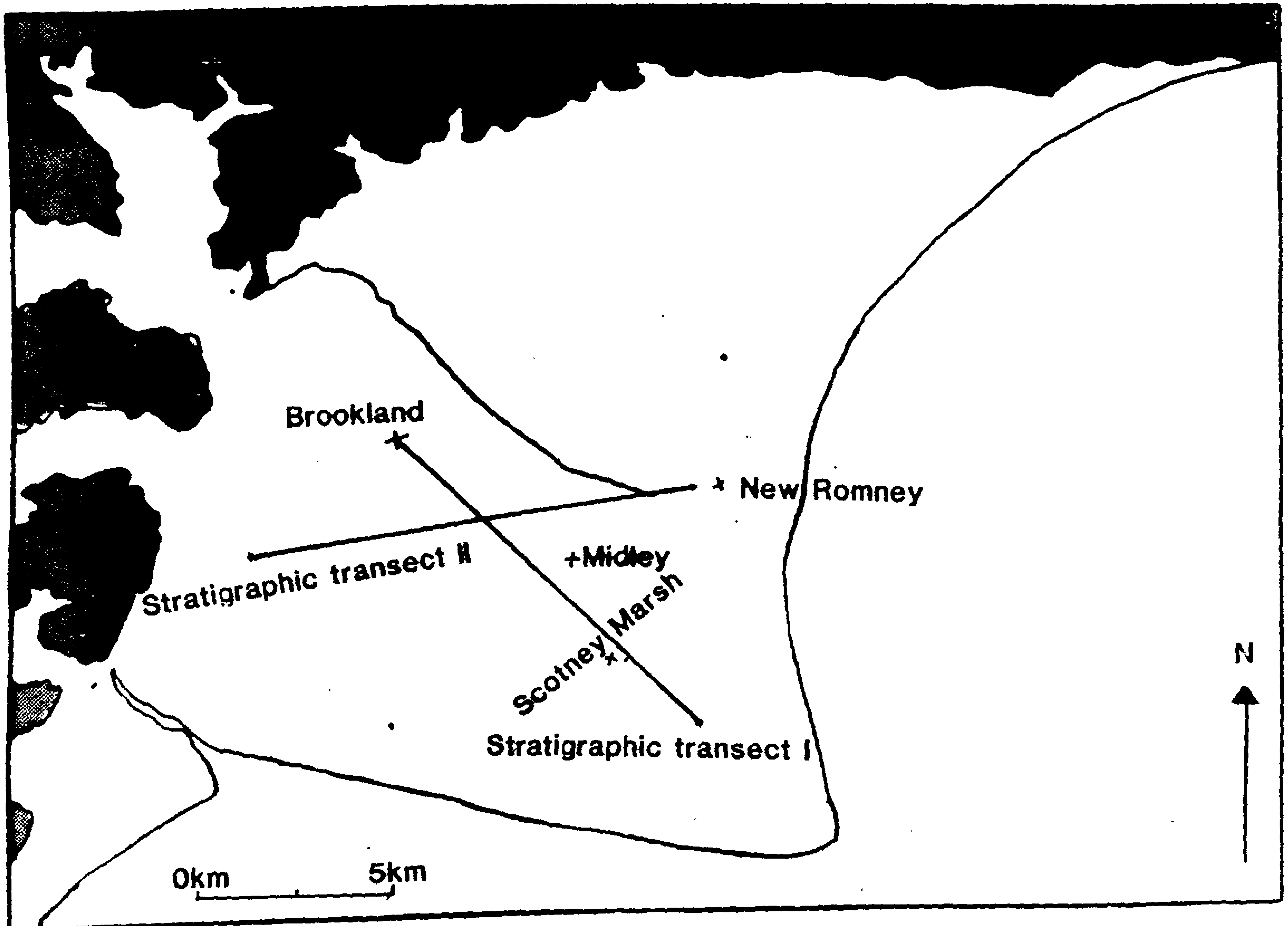


Figure 4.1: Location of stratigraphic transects I and II.

4.1.1 Stratigraphic Transect I : Results.

It was determined that three distinct sedimentary sequences are present across the marsh: the 'south-east of Lydd' sequence in Denge Marsh, found between cores tr1 and tr18 (figures 4.3a-b); the 'channel sands' sequence, found between cores NM1-tr50 and cores tr55-tr66 (figures 4.3c-e); and the 'main marsh' sequence, recorded between cores tr51-tr54 and cores tr67-tr97 (figures 4.3d and 4.3f-g). The location of Scotney Marsh within this transect is illustrated by the sediments of ridge 1, as recorded in cores tr28-tr29. Some difficulty was experienced in sampling the sediment directly to the north-west of Scotney Marsh as access to the land was denied. Consequently, a relatively disjointed series of samples were completed along the roadside in this region. The results of stratigraphic transect I are summarised in figure 4.4.

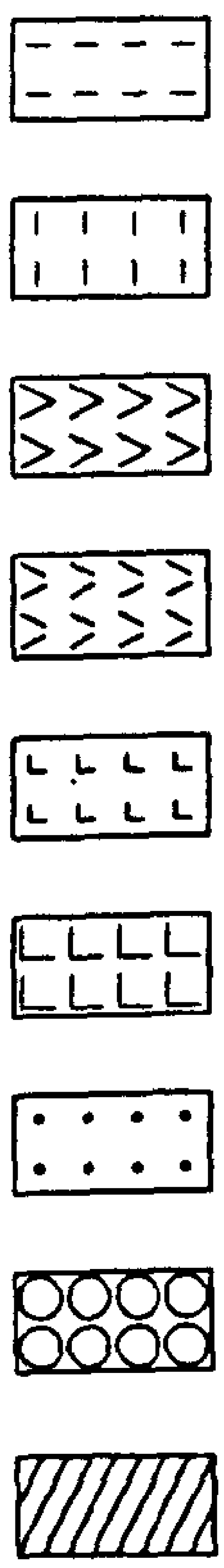
4.1.1.1 The Stratigraphy to the South-east of Lydd (Denge Marsh, DM).

Cores tr1-tr18.

Unit DM1 : The deepest stratigraphic unit recorded is an impenetrable gravel, found at the base of a number of cores (tr1, tr2, tr17) at both the south-east and north-west extremes of Denge Marsh.

Unit DM2 : Unit DM2 is a battleship to dark grey sand with some laminated silt at both the south-east and north-west extremes of Denge Marsh. This becomes impenetrable at ca.

Legend



Sh	**Substantia humosa**
Th	**Turfa herbacea**
Dl	**Detritus lignosus**
Dh	**Detritus herbosus**
Ag	**Silt**
As	**Clay**
Ga	**Sand**
Gg	**Gravel**
Sc	**Disturbed ground**

Variable vertical and horizontal scales are utilised to best demonstrate the recorded stratigraphic changes.

Figure 4.2: Legend for straigraphic figures.

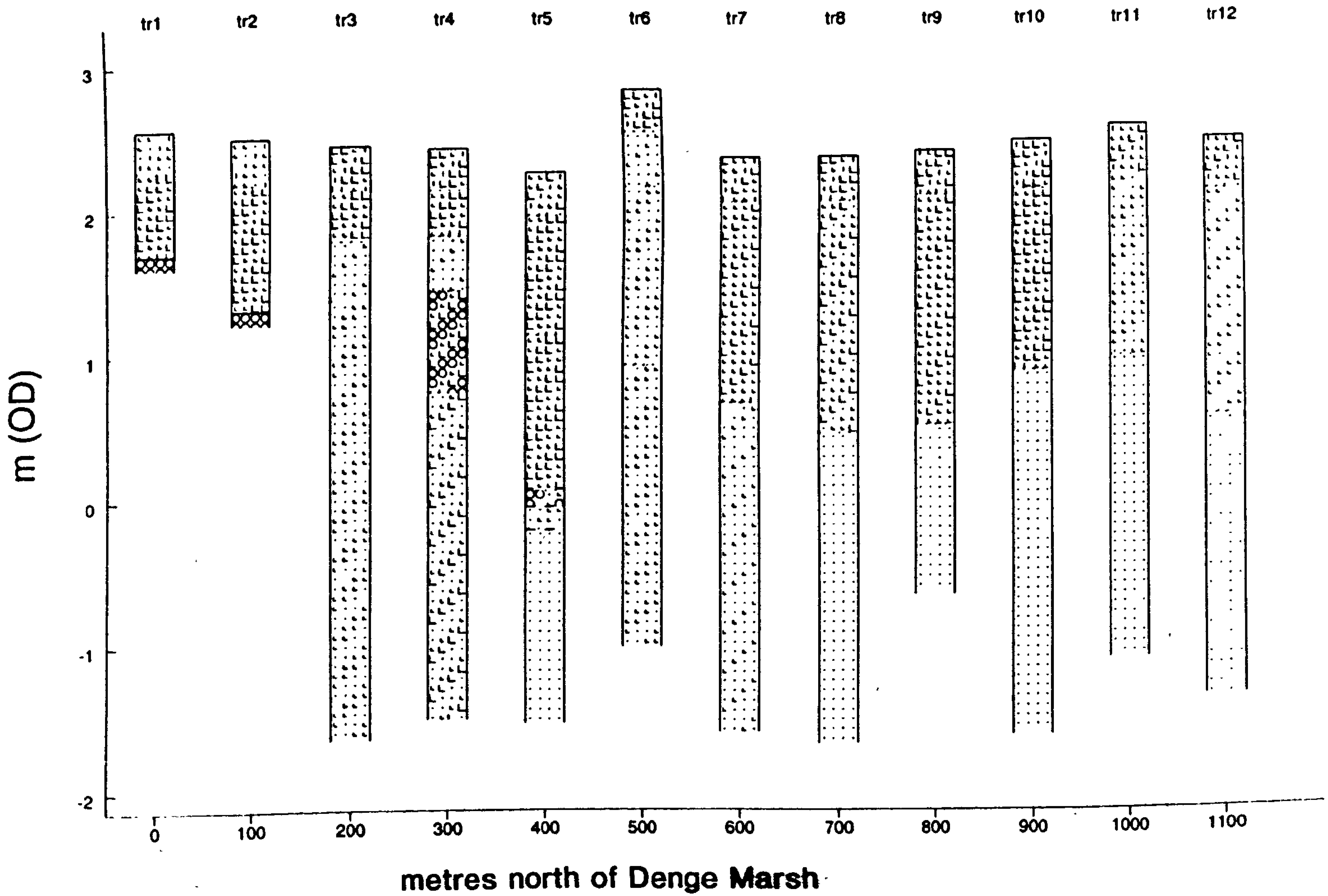


Figure 4.3a: Stratigraphic Transect I (legend figure 4.2).

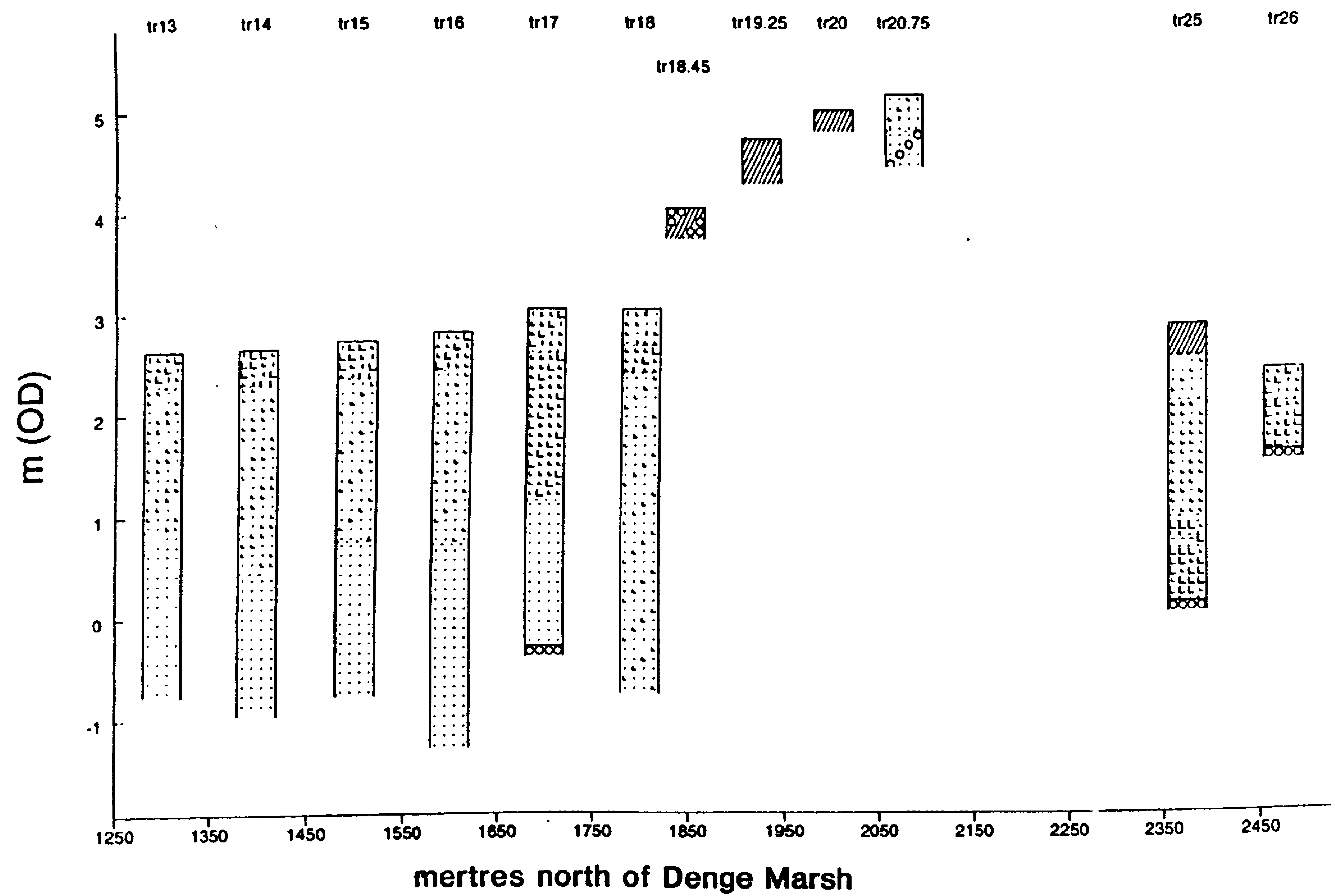


Figure 4.3b: Stratigraphic Transect I (legend figure 4.2).

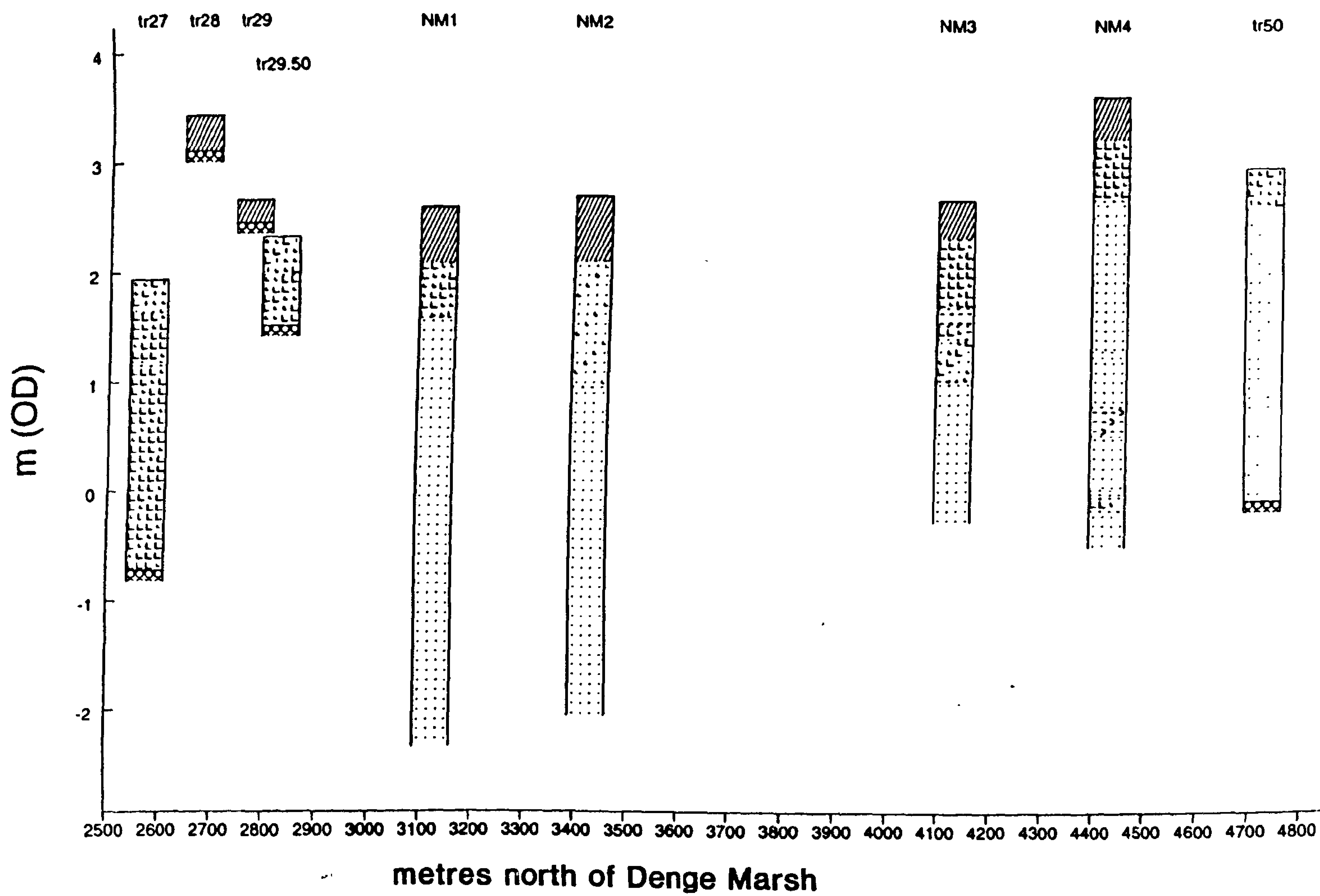


Figure 4.3c: Stratigraphic Transect I (legend figure 4.2).

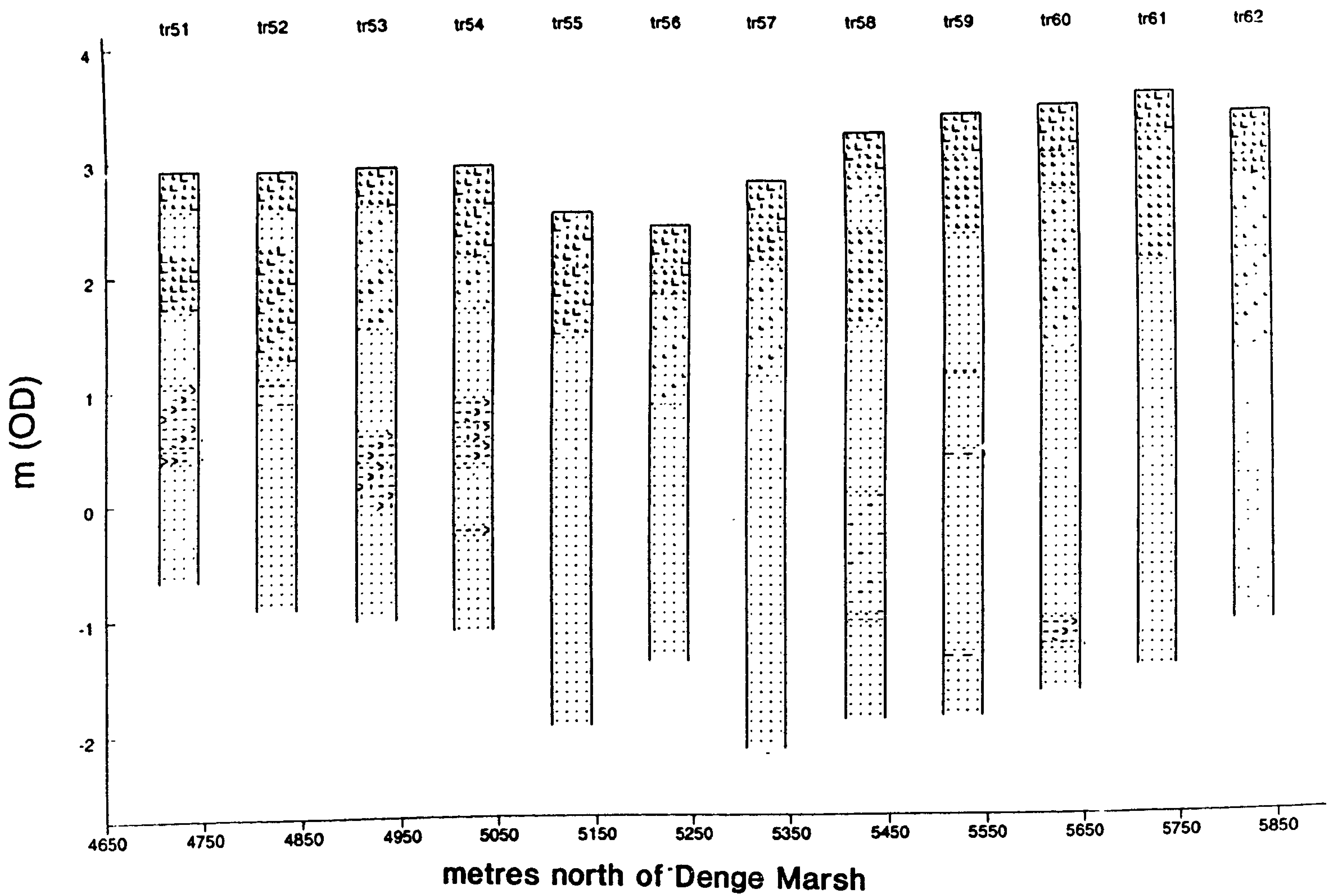


Figure 4.3d: Stratigraphic Transect I (legend figure 4.2).

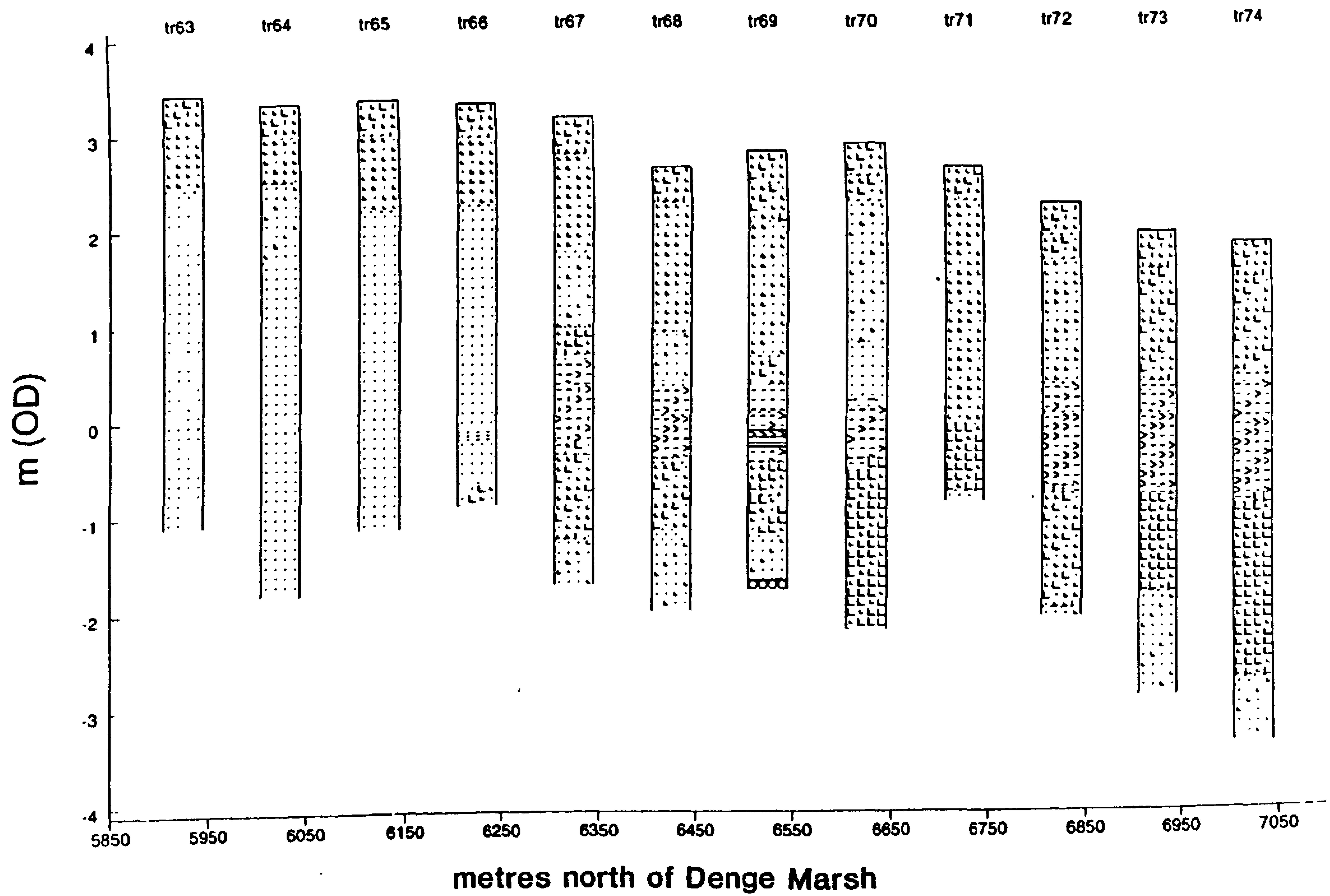


Figure 4.3e: Stratigraphic Transect I (legend figure 4.2).

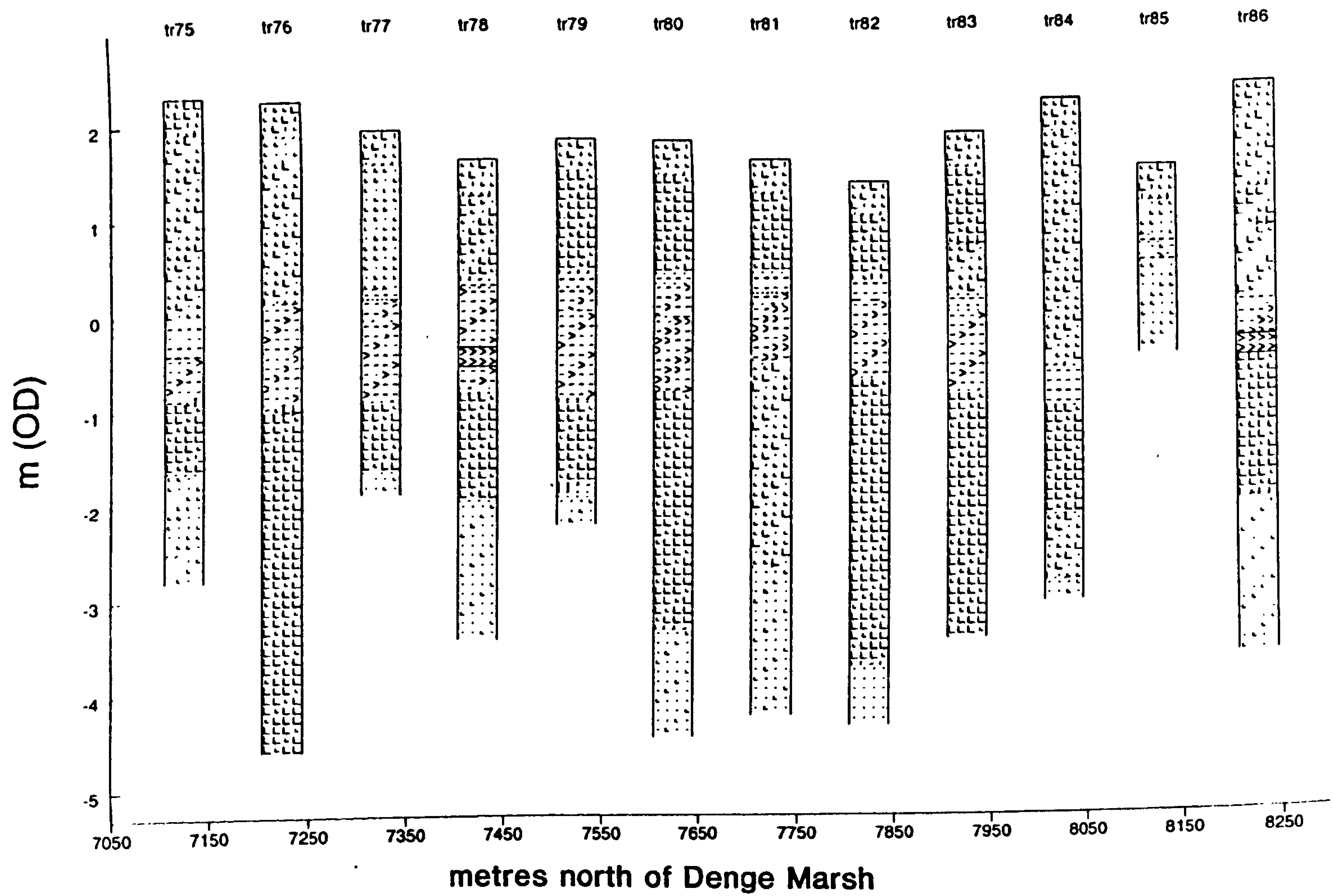


Figure 4.3f: Stratigraphic Transect I (legend figure 4.2).

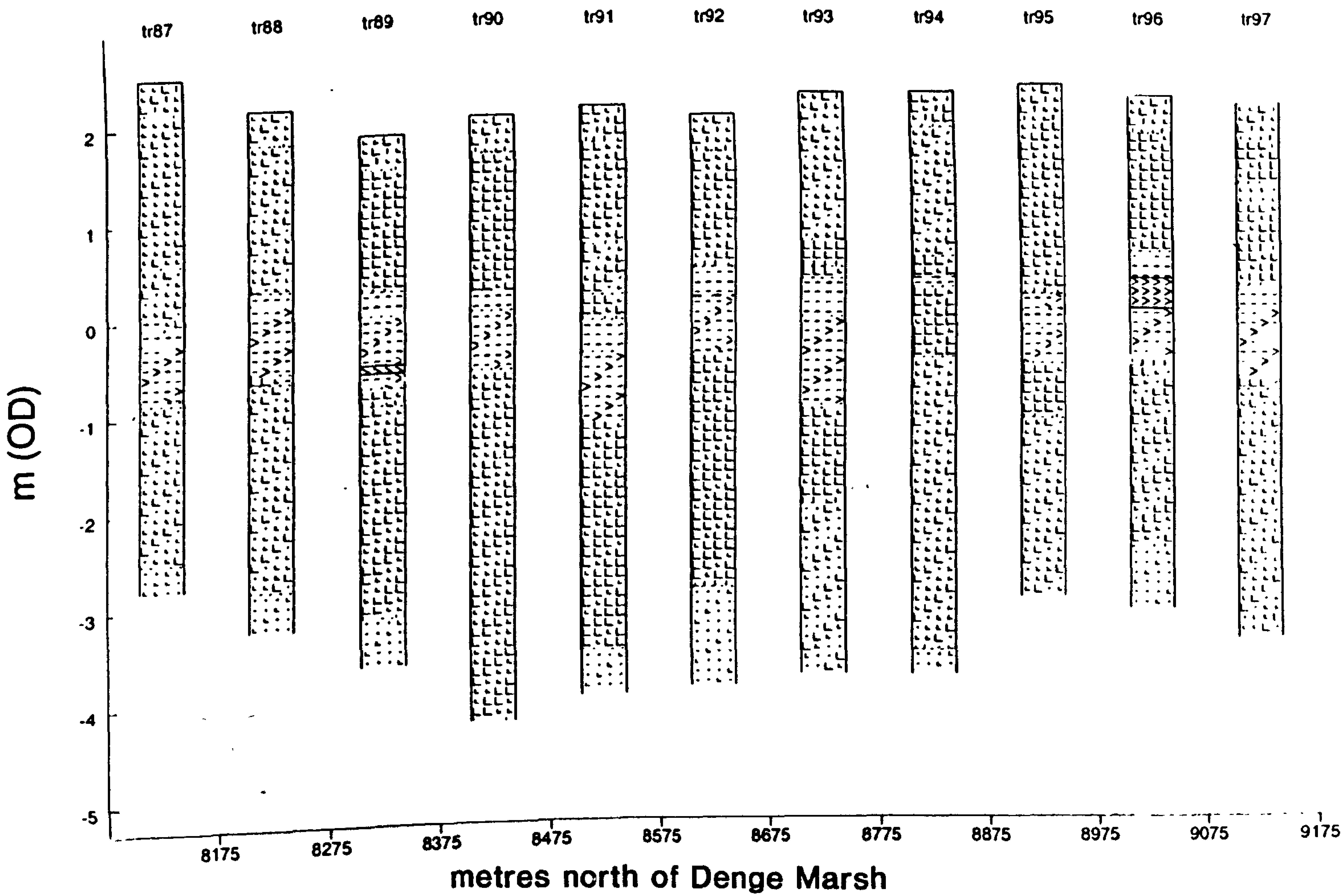


Figure 4.3g: Stratigraphic Transect I (legend figure 4.2).

-1.00m OD, and, in these cores, unit DM1 is not encountered. A general fining upward sequence is recorded in the upper parts of this unit. The upper contact of this unit is at an altitude of ca. +1.00m OD and contains a silty-sandy gravel lens in the upper part of two cores (tr4, tr5).

Unit DM3 : Overlying unit DM2, at a variable depth of ca. +1.00m OD, is a complex sequence of silty-clays to silty-sands, which generally fine upwards. These tend to be grey with significant iron staining throughout.

4.1.1.2 The Stratigraphy of the Channel Sands (CS).

Cores NM1-tr50 and tr55-tr66.

Unit CS1 : The deepest stratigraphic unit identified is an impenetrable grey sand, reaching a depth of ca. -1.50m OD in many cores. Within many of the cores, some well-humified organic sediments are recorded as laminations or as lenses of between ca. 1mm and 0.10m in thickness. These organic sediments are probably detrital and deposited as organic horizons or peat balls. The sediment is predominantly a saturated silty-sand to coarse sand with detrital wood and broken shells. A gravel base is encountered in core tr50.

Unit CS2 : This unit often exhibits a transitional basal contact with unit CS1, which tends to occur at ca. +1.00m OD in most cores. The sediments are sands that generally fine upwards, becoming silty-sands to silty-clays. The uppermost part of this unit shows some variability both laterally and vertically. The sediment tends to be saturated and orangey-grey in colour, with iron staining decreasing variably with depth.

4.1.1.3 The Stratigraphy of the Main Marsh (MM).

Cores tr51-54 and tr67-tr97.

Unit MM1 : The deepest stratigraphic unit is a grey sand to silty-sand which is impenetrable between altitudes of ca. -1.50m OD and ca. -4.00m OD. However, in one core (tr69) the sand was penetrated and a basal gravel recorded. The unit tends to be sand in the south-east, around Midley, whereas, towards Brookland, the sediment becomes more silty-sand. Some laminations and broken shells are recorded throughout the area where these sediments occur.

Unit MM2 : This unit is absent in the south-east, around Midley, where unit MM3 directly overlies unit MM1. Towards the north-west, however, a considerable thickness (ca. >3.50m) of blue-grey silty-clays to clayey-sandy silts are recorded in many of the cores. The basal altitude of this unit is variable; ranging from ca. -1.00m OD to below ca. -4.50m OD, and the upper contact is at an altitude of ca. +1.00m OD. Occasional humified organic laminations and broken shells are also present in some cores. Many cores exhibit a transitional organic-rich upper contact.

Unit MM3 : This unit is a peat that is laterally consistent between cores tr51-tr54 and cores tr67-tr97. However, some variability exists between the two groups. The former is generally ca. < 0.70m in thickness and present between altitudes of ca. 0.00m OD and ca. +1.00m OD. The base of the biogenic unit is a peaty-sand which becomes a well-humified reddy-brown peat. Towards the top of the unit, the peat becomes very dark brown and well-humified. The peat contains variable components of turfa and detritus throughout. The latter example of peat, however, is up to ca. 1.20m in thickness and occurs between altitudes of ca. -1.20m OD and ca. +0.50m OD. The biogenic sediments contain variable amounts of turfa and detritus, as well as significant quantities of wood in many cores. At the base of this unit, in many cores, is a transitional peaty-clay which becomes a reddy-brown peat with a large quantity of wood upward. Above this, the peat becomes dark brown and, eventually, black, with the degree of humification increasing upward. At the upper contact of unit MM3 in many cores is a transitional peaty-clay. The wood content also tends to increase towards the north-west, in the region of Brookland (cores tr73-tr77).

The organic sediments are absent in cores tr71, tr85 and, to some extent, core tr94. These cores are suggested as being the infilling of tidal channels, as has been described by Long and Innes (1995a), hence explaining the lack of biogenic sediment.

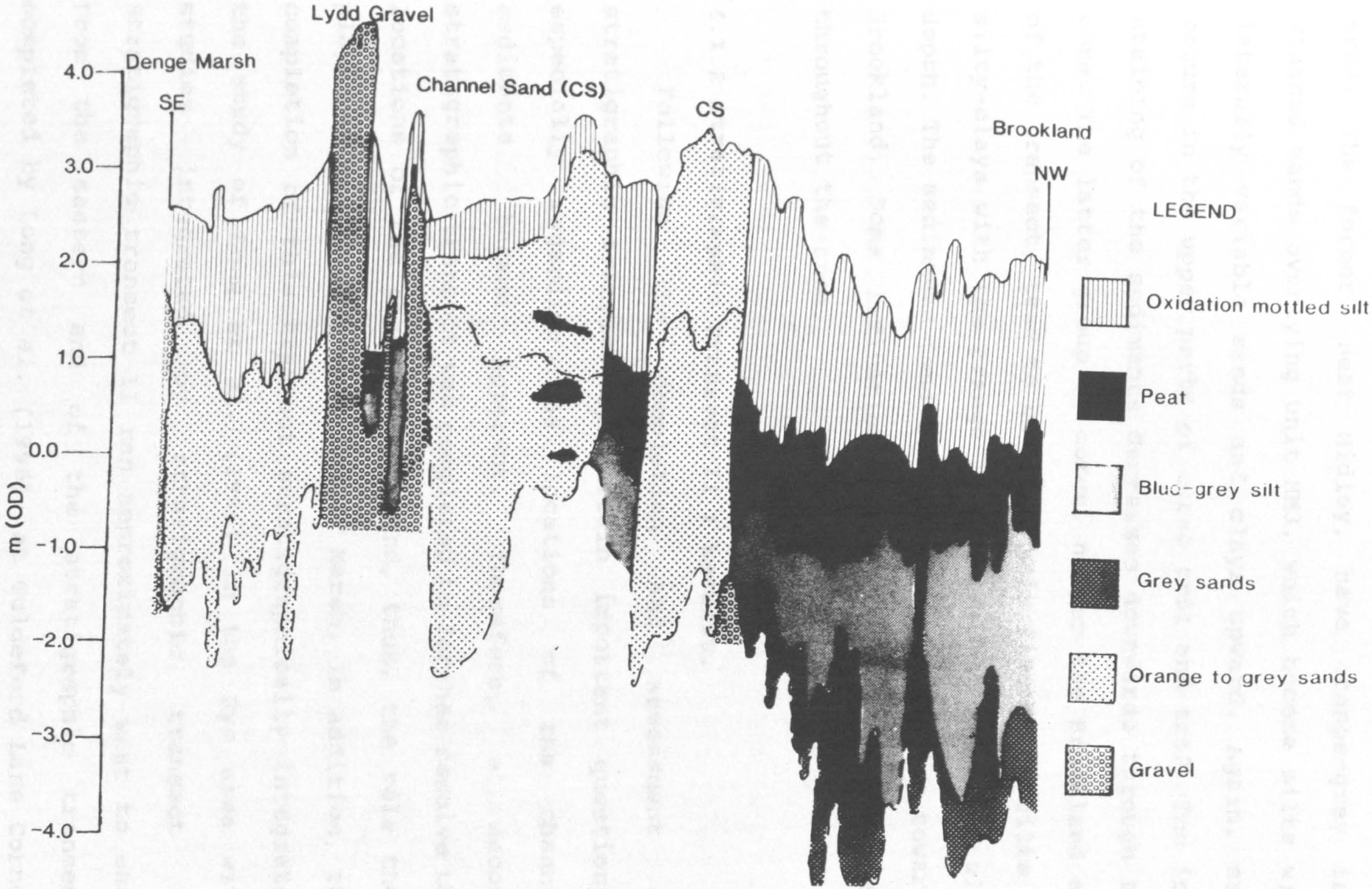


Figure 4.4: Summary diagram of stratigraphic transect I.

Unit MM4 : The deposits of unit MM4 again show variability between the two groups of cores, *i.e.* tr51-tr54 and tr67-tr97. The former, near Midley, have orange-grey iron stained sands overlying unit MM3, which become silts with laterally variable sands and clays upward. Again, sand occurs in the upper parts of cores tr51 and tr52. The iron staining of the sediments decreases downwards through the core. The latter group of cores, nearer the Brookland end of the transect, tend to be relatively fine sandy-silts to silty-clays with the orange iron staining decreasing with depth. The sediments tend to fine upwards and also towards Brookland. Some laminations were present in unit MM4 throughout the area.

4.1.2 Stratigraphic Transect II : Results.

Following the completion and assessment of stratigraphic transect I, certain important questions, especially regarding the locations of the channel sediments, became apparent. Therefore, a second stratigraphic transect was completed to further resolve the locations of the channel sands and, thus, the role they played in the evolution of Romney Marsh. In addition, the completion of this transect stratigraphically integrated the study of Long *et al.* (1996) in the Rye area with studies integrated by stratigraphic transect I. Stratigraphic transect II ran approximately west to east from the eastern end of the stratigraphic transect completed by Long *et al.* (1996), at Guldeford Lane Corner

(TQ959228), to Hammonds Corner (TR052248), west of New Romney (figure 4.1). Across the transect, the sediments essentially fell into two types: Channel Sands and Main Marsh sediments, figures 4.5a-e. The general lithostratigraphy of these two groups of sediments in stratigraphic transect II are similar to the corresponding groups in stratigraphic transect I. The characteristics of the sediments particular to this transect are outlined below. The results of stratigraphic transect II are summarised in figure 4.6.

4.1.2.1 The Stratigraphy of the Channel Sands.

Cores : RT16-RT19, t10-t6, NRT5-NRT11, NRT21-NRT28.

Unit CS1 : The deepest stratigraphic unit identified is a grey impenetrable sand, generally extending to an altitude of ca. -2.00m OD. However, in cores t10-t6, the sands were impenetrable below ca. 0.00m OD. In core NRT11, the sand was penetrated and a basal gravel recorded. The sediments are grey with black staining. In addition, some cores contain peat balls, organic laminations and broken shells.

Unit CS2 : A transitional contact is recorded between this and unit CS1 in most cores at ca. +1.00m OD. The sediment is a grey-orange iron stained silty-sand that generally fines upwards. Laminations are recorded in this unit throughout the area, especially near the Rhee Wall.

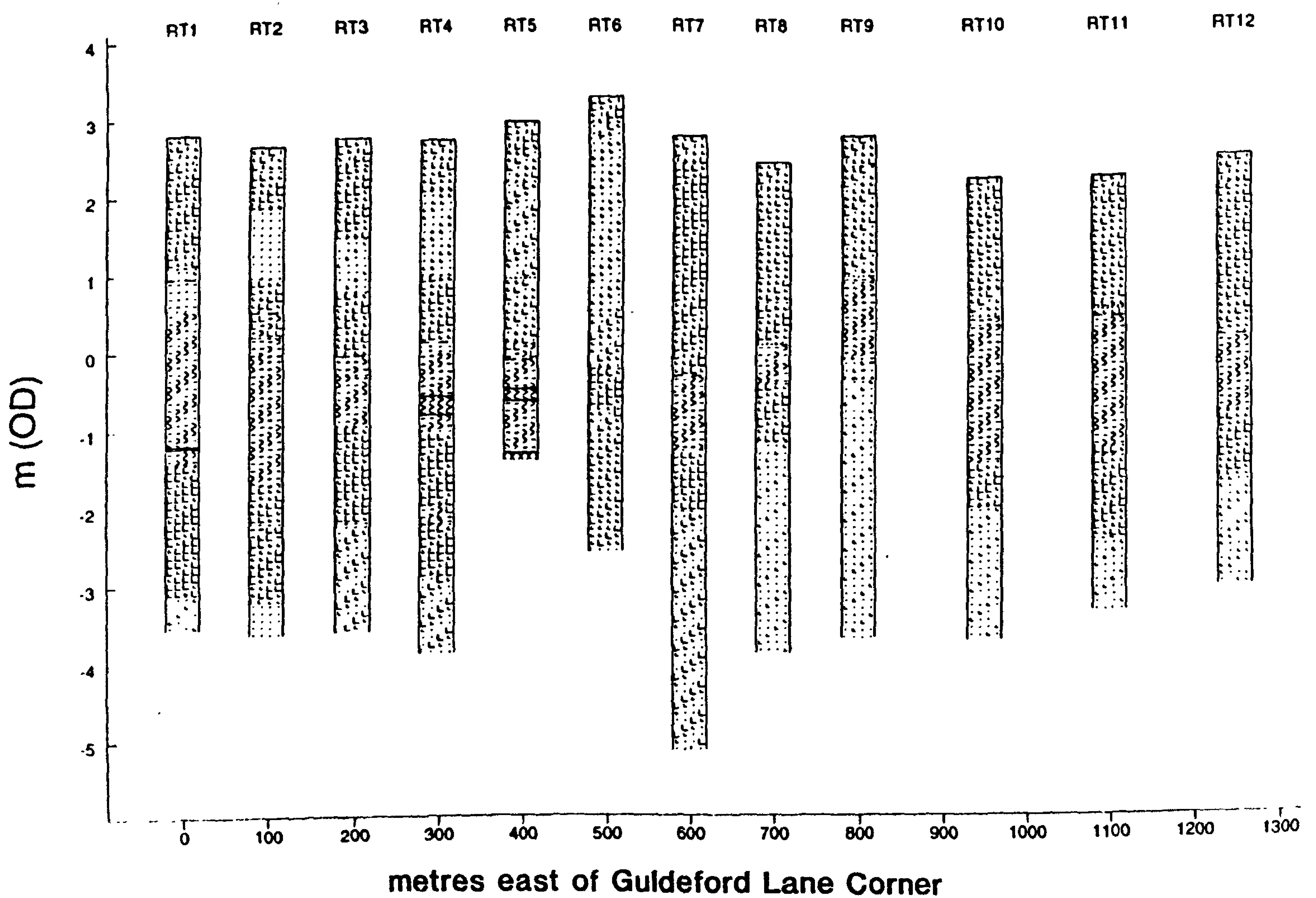


Figure 4.5a: Stratigraphic Transect II (legend figure 4.2).

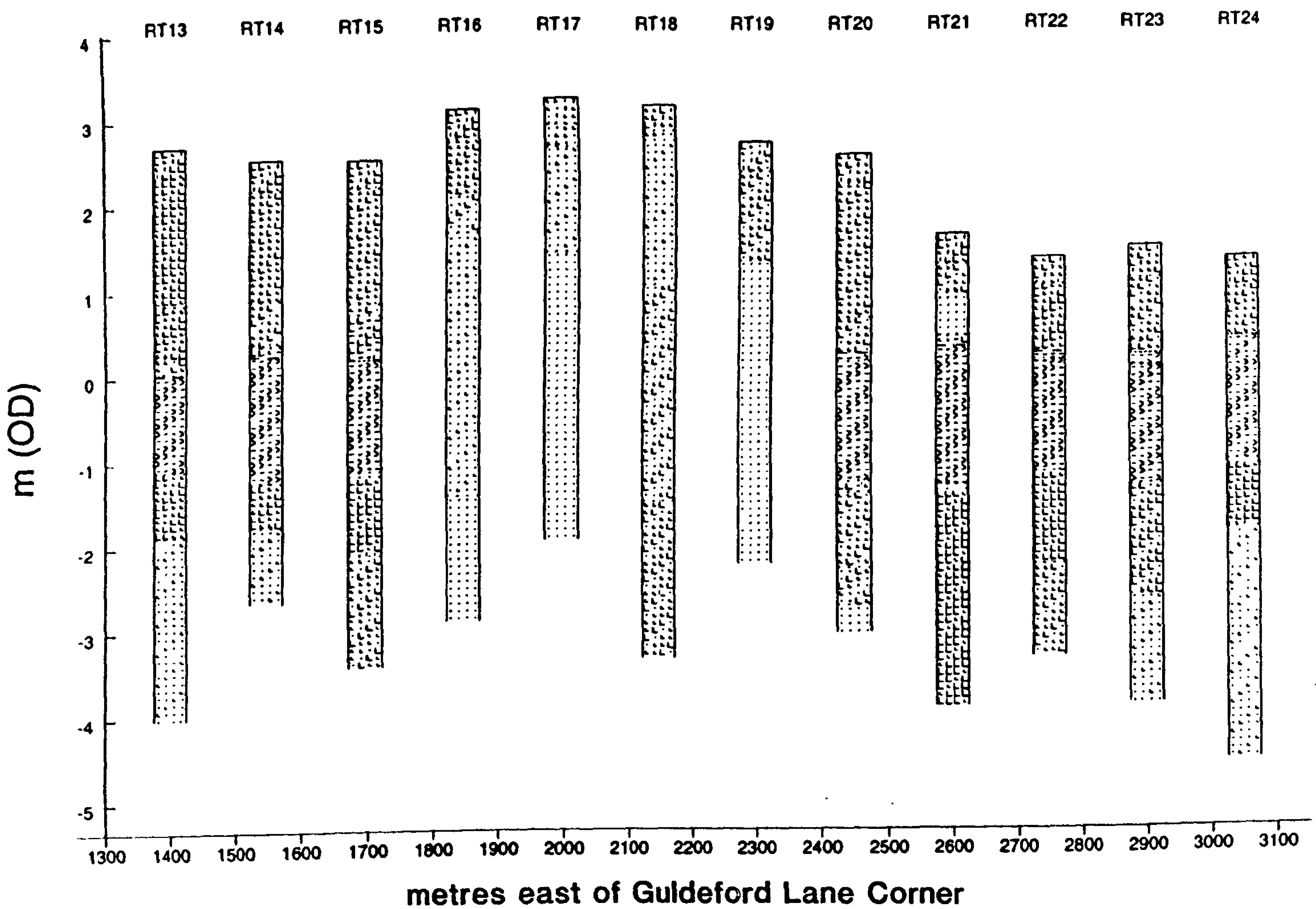


Figure 4.5b: Stratigraphic Transect II (legend figure 4.2).

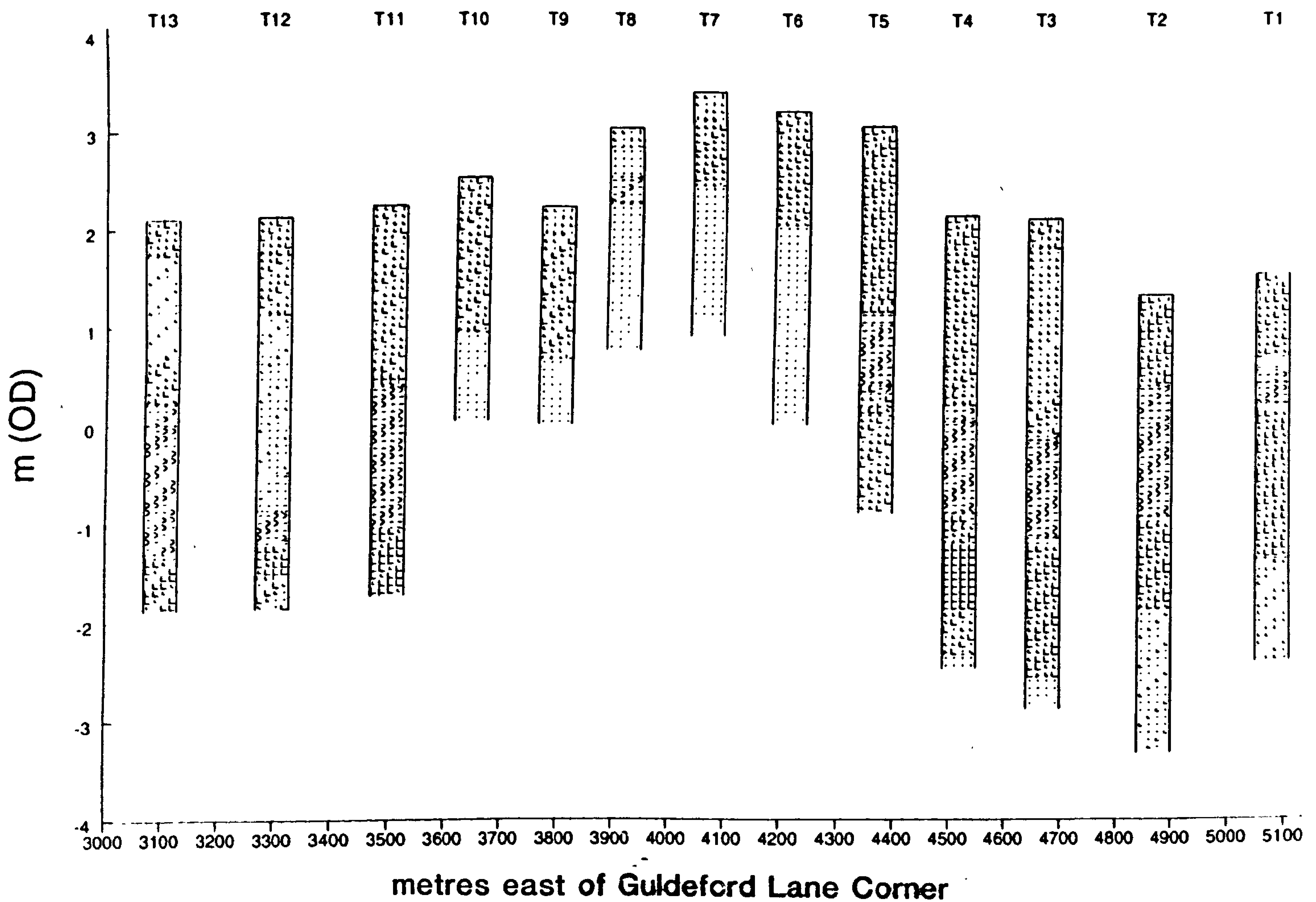


Figure 4.5c: Stratigraphic Transect II legend figure 4.2).

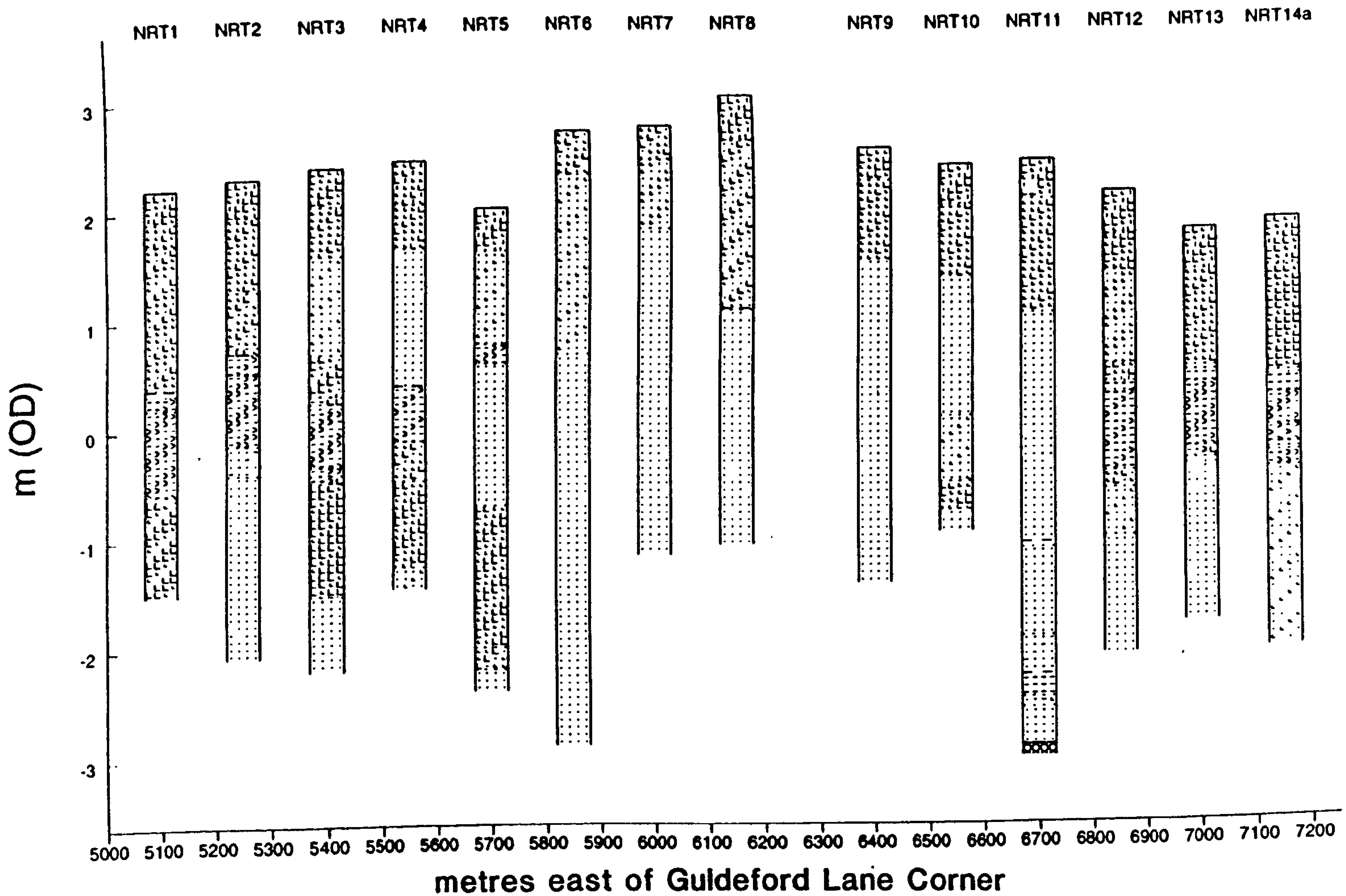


Figure 4.5d: Stratigraphic Transect II (legend figure 4.2).

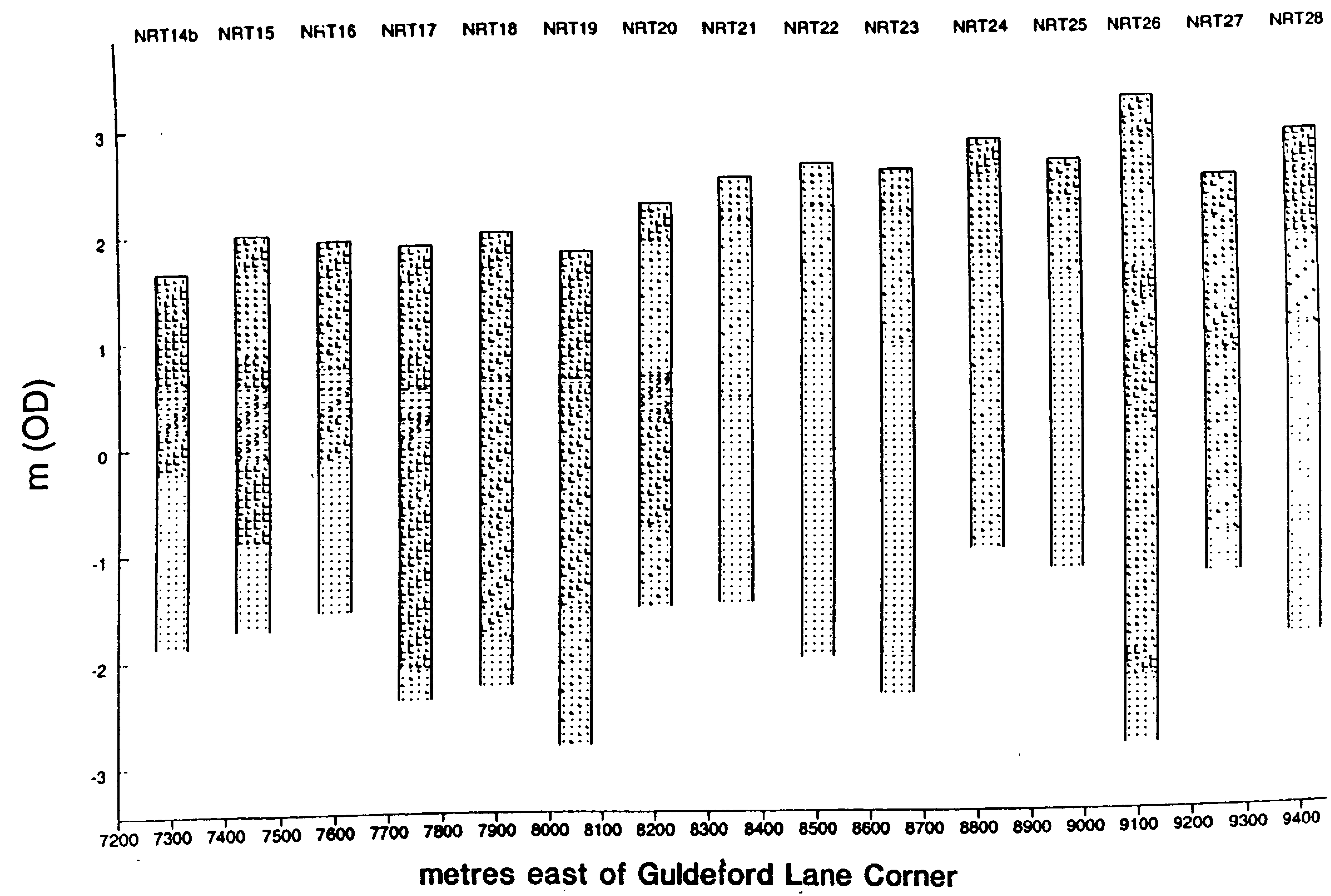


Figure 4.5e: Stratigraphic Transect II (legend figure 4.2).

4.1.2.2 The Stratigraphy of the Main Marsh.

Cores : RT1-RT15, RT20-t13, t5-NRT4, NRT12-NRT20

Unit MM1 / MM2 : The deepest stratigraphic unit extended down to an altitude of ca. -3.50m OD, and is a grey sand which fines upward to a grey to blue-grey silty-sand to silty-clay, becoming organic rich at the top. The unit is generally impenetrable and becomes saturated with depth. However, in core RT5 a basal gravel was reached. Some organic-rich zones were recorded, and also some strongly laminated examples with broken shells. This unit is truncated in core NRT5 by an eroded upper contact with channel sands characteristic of unit CS1.

Unit MM3 : This unit is a peat occurring between altitudes of ca. -1.50m OD and ca. +0.50m OD in the west, and reaching a maximum thickness of ca. 2.00m. Between cores tr5 and NRT4, a thinning of the unit occurs to ca. 0.70m and is found between altitudes of ca. +0.50m OD and ca. -0.50m OD. The thinning of the peat unit continues eastwards between cores NRT12 and NRT20 to ca. 0.20m thickness where occasional clay-rich zones were observed. The unit generally consists of a grey-brown peaty-clay, becoming a reddy-brown peat with wood and, finally, a black peat. The humification tends to increase upwards, and both turfa and detritus are present throughout. In some cores, a transitional peaty-clay upper contact was recorded and the wood content of the peat was reduced as the peat thinned eastwards. In addition, the upper contact of the peat was

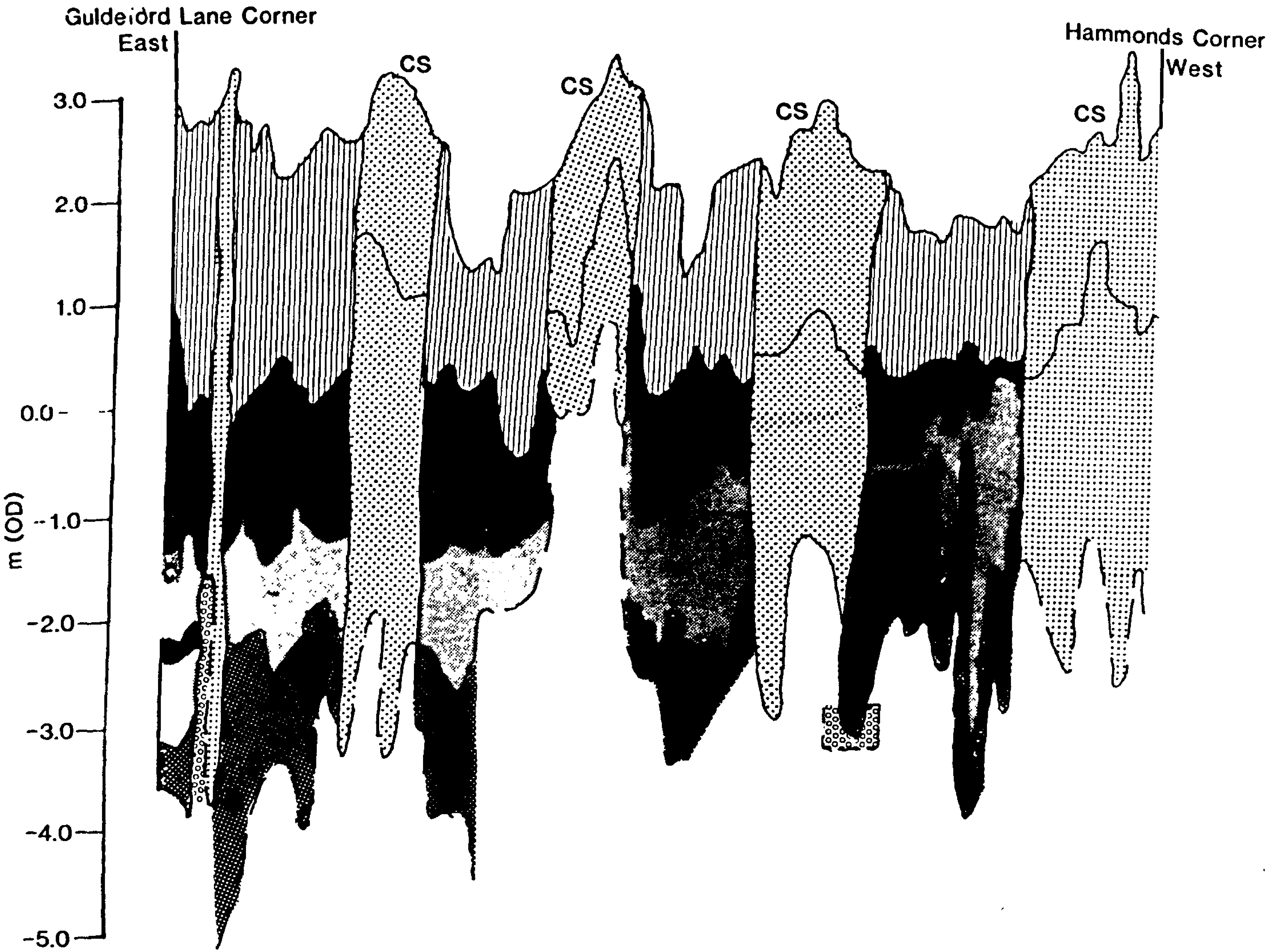


Figure 4.6: Summary diagram of stratigraphic transect II (legend figure 4.4).

truncated in some cores, *i.e.* core t12.

Unit MM4 : The uppermost stratigraphic unit is a grey-orange iron stained silty-sand to silty-clay. This unit generally fines upwards with laminations present in its lower part. The unit is absent were the channel sands truncate unit MM3, depositing units CS1 and CS2 above.

4.2 Palaeoenvironmental Reconstruction of Scotney Marsh : Results.

By considering the results of the extensive stratigraphic transects across Romney Marsh it was possible to determine an area in which to complete a detailed palaeoenvironmental reconstruction. It was decided that the barrier / back-barrier interface at Scotney Marsh would provide the most effective location for this palaeoenvironmental reconstruction. This site was selected as the barrier / back-barrier interface may provide a location that was sensitive to changes in sea-level or to barrier breaching due to storms, and thus, may assist in formulating a model of the evolution of Romney Marsh.

The lithostratigraphy of Scotney Marsh will first be discussed with reference to a number of stratigraphic transects which are considered representative of selected parts of Scotney Marsh. Secondly, the results of the palaeoenvironmental analysis of each of the typecores will be discussed, *i.e.* diatom, pollen and particle size analysis for cores AY17 and G60, and diatom and pollen analysis for cores A-B27, AW63 and AW-AX67. In addition, the chronostratigraphy of 10 samples selected from the typecores will be presented.

4.2.1 Lithostratigraphy of Scotney Marsh : Results.

The stratigraphic investigations carried out in Scotney Marsh demonstrated that an ideal sedimentary succession exists (see table 4.1). This sequence was recorded in all cores, dependent on a number of factors to be discussed below. Each of the sedimentary units are described below, utilising the Troels-Smith (1955) sediment classification scheme (see section 3.1).

Topsoil

3,0,0+,2,-
As3,Ag1
Sh+,Th^2+,Ga+,Gg(maj)+
Brown silty-clay topsoil, locally with sand and gravel.

Oxidation mottled silts

2+,0,0,2,0 to 2+,2,0,2,0
As3,Ag1 to Ga3,Ag1
Lf++,Sh+,Th^1+,Dh+,Gg(maj)+,Ptm+
Orange-grey silty-clay to silty-sand, with oxidation mottling ranging from slight to extreme both laterally and altitudinally. Locally laminated with both *Turfa herbacea* and *Detritus herbosus* present.

Peat

3+,2+,2,2,0 to 4,0,1,2,1+
Sh4 to Sh2,Dh2
Th^2++,As+,Ag+,Ga+,Dl+
Black to dark brown humified to well-humified peat, predominantly *Substantia humosa* with *Detritus herbosus* and *Turfa herbacea*. Locally some minerogenic content and *Detritus lignosus*.

Peaty-clay

3,0,0+,2,0 to 3,1,1,2,0
As2,Sh1,Dh1
Th^2+,Dl+,Ag+,Ga+
Grey to brown unit, ranging down from

	humified peat comprised of *Detritus herbosus* and *Substantia humosa* with some *Turfa herbacea* and minerogenic components to a organic silty-clay with some *Turfa herbacea* and *Detritus herbosus*. Transitional.
Blue-grey silts	2,0,0,2,0 to 2+,1+,0,2+,0 As3,Ag1 to Ga3+,Ag1 Sh+,Dh+,Th^2+,Ga+,Dl+,Gg(maj)+,Ptm+ Bluey-grey to battleship-grey silty-clay to silty-sand, generally coarsening downward. Locally some laminations are present and sand partings are common throughout. Some organics for example, *Turfa herbacea* and *Detritus herbosus* (especially *Phragmites*) are present at many sites. Bluey-black staining of the sediments is common.
Gravel	2+,0,0,2+,4 Gg(maj)4 Gg(min)+,Ptm+ Basal gravel encountered in all cores throughout the study area. Rarely observed in detail due to the difficulty of sampling (though observed in numerous Brett Gravel quarrying sections) but predominantly well-rounded flints. Sharp upper contact.

Table 4.1 : Typical Sedimentary Succession of Scotney Marsh.

4.2.1.1 Sedimentary Characteristics of Scotney Marsh.

The complete stratigraphic data set from Scotney Marsh is presented in Appendix 1 in TSPPlus format, see section 3.1.2. The typical sequence presented in table 4.1 describes almost the entire range of sediments recorded throughout Scotney Marsh, with some notable exceptions which will be examined in more detail later. However, the complete suite of sediments is not present throughout the study area. A major controlling factor appears to be the morphology of the basal gravel unit. As the gravel proved to be impenetrable, it represents the basal depth for each of the 2751 cores investigated in the study area. One of the critical controlling factors on the occurrence of the different sedimentary units in the study area is altitude. Therefore, where the gravel morphology is sufficiently elevated, part or all of the typical stratigraphy is absent. Indeed, in a significant number of cores, the gravel morphology is elevated to the ground surface, with a loose gravely topsoil developed on top. Two major gravel ridges occur in the study area; ridges 1 and 2 (figure 4.7). Gravel features are defined, for the purpose of this study, as the areas where the buried gravel surface is at an altitude of ca. > +1.20m OD. This altitude was selected, first, as it clearly illustrates the location and extent of the gravel ridges of the area, and, secondly, that it is the approximate altitude of the contact between the peat and the oxidation mottled silts. Thus, figure 4.7 also provides an approximation of the ground surface morphology and elevation towards the end of the period of peat formation, and, therefore, provides an insight into the relationship between peat accumulation and gravel morphology.

Ridge 2 forms the south-eastern boundary of the study area and carries the B2075 Lydd to Camber road. The morphology of ridge 1 was mapped more effectively as it occurs through the middle of the study area, trending south-west to north-east (see figure 4.7). Unfortunately, the Brett Gravel plant site is located on ridge 1 and the

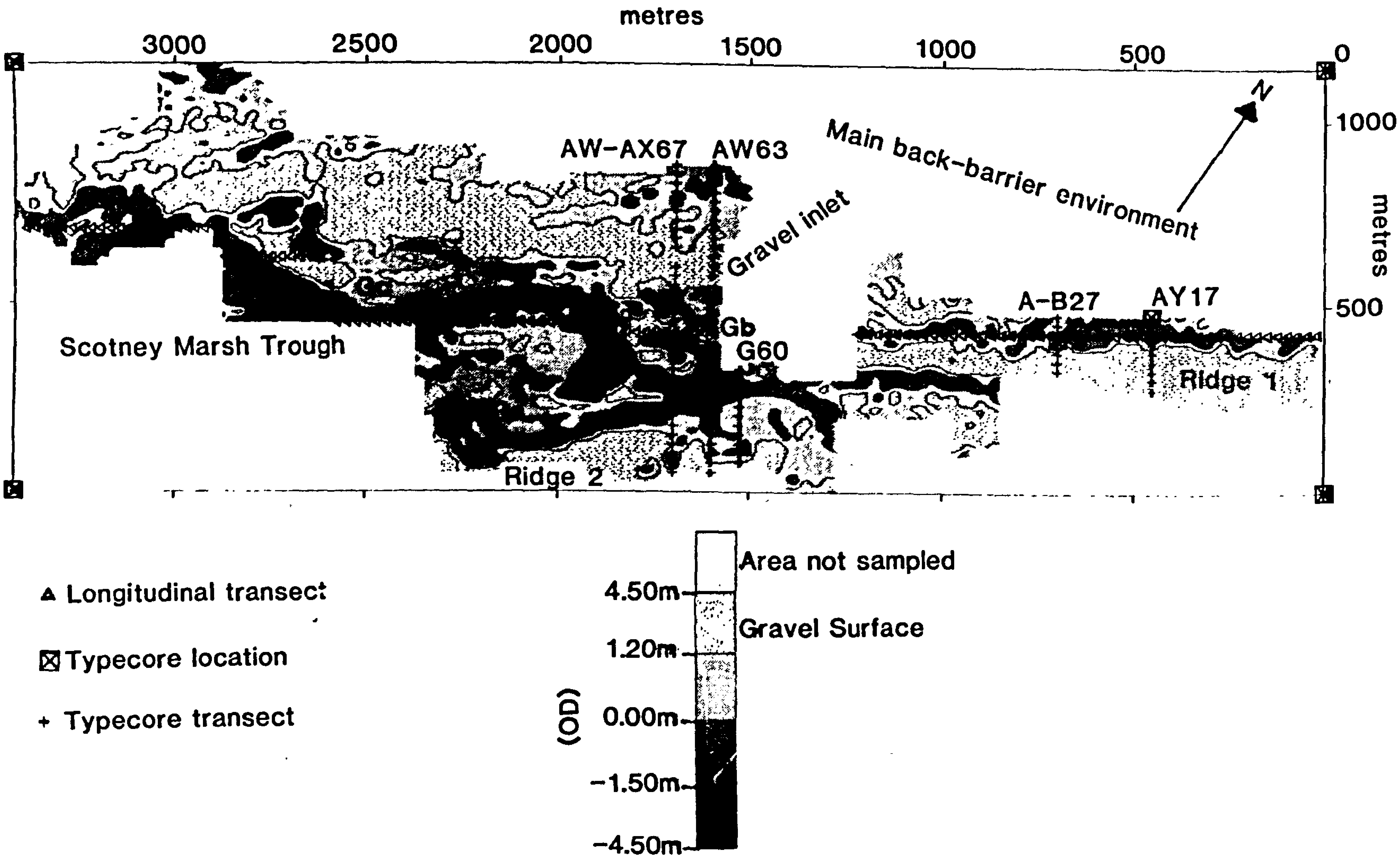


Figure 4.7: Location diagram and contour plot of the Scotney Marsh area.

morphology of the sediments in this area is unknown. A number of less extensive gravel features were also identified between ridges 1 and 2 (figure 4.7), for example Ga, Gb and Gc. These features are smaller gravel ridges and offshoot ridges from the two major ridges.

At many sites the full suite of sediments was not recorded, even though the gravel surface is sufficiently depressed. In particular, the peat and peaty-clay units are absent across wide areas of Scotney Marsh. The absence of organic sediments appears to be controlled partially by the location of ridge 1. The full suite of sediments is typical of the area to the north of ridge 1, *i.e.* the main back-barrier environment, whereas in the Scotney Marsh trough, the blue-grey silt typically passes transitionally into the oxidation mottled silts and topsoil. However, the detailed stratigraphic investigations demonstrated that some peats and peaty-clays do occur outside the main back-barrier environment in Scotney Marsh. These units tend to be located along the margins of the Scotney Marsh trough or in sheltered inlets in the gravel surface, for example, the gravel inlet in which typecores AW63 and AW-AX67 were located. This raises a very important question as to whether peat was deposited across the whole of the Scotney Marsh trough and subsequently removed in places, or was peat only able to form at selected sites where conditions were conducive to colonisation and eventual peat accumulation? This will be addressed in the subsequent discussion of the stratigraphic record.

4.2.1.2 Selected Stratigraphic Transects from Scotney Marsh.

In order to determine the inter-relationships of the sediments of Scotney Marsh a series of stratigraphic transects are presented here. These transects illustrate the lithostratigraphic changes observed across the strike of the gravel ridges, in a north-west to south-east direction (figure 4.7). Additionally three-dimensional images of both the gravel and peat surfaces will be discussed for the areas from which the typecores were sampled, to aid visualisation of the palaeoenvironmental reconstruction. Also, a longitudinal transect illustrating lithostratigraphic changes near-parallel to the strike of the gravel ridges, i.e. south-west to north-east, will be discussed.

4.2.1.3 Lithostratigraphy of Typecores AY17 and A-B27.

Both typecores AY17 and A-B27 were selected from sedimentary sequences typical of the main back-barrier environment in Scotney Marsh (figures 4.8 and 4.9). In this area the full suite of sediments typical of Scotney Marsh is present. These diagrams also illustrate how critical the altitude of the gravel surface is in controlling the nature of the sedimentary record at each site. The transects demonstrate that the sediments of the main back-barrier environment abut the gravel surface of ridge 1, with the blue-grey silts overlying gravel where the gravel surface is sufficiently deep. Above this unit the peaty-clay and

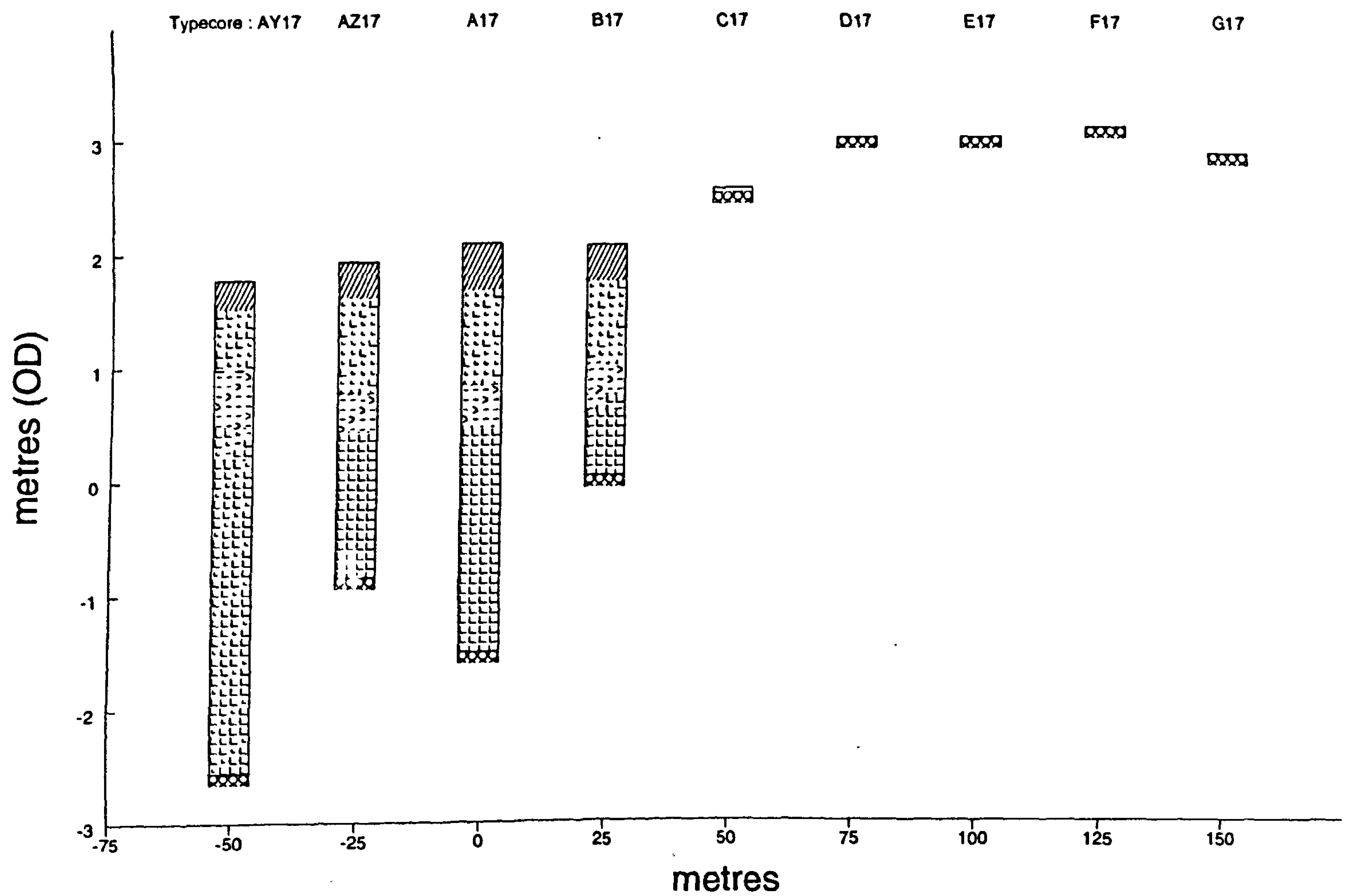


Figure 4.8: Stratigraphic Transect from which typecore AY17 was selected (legend figure 4.2).

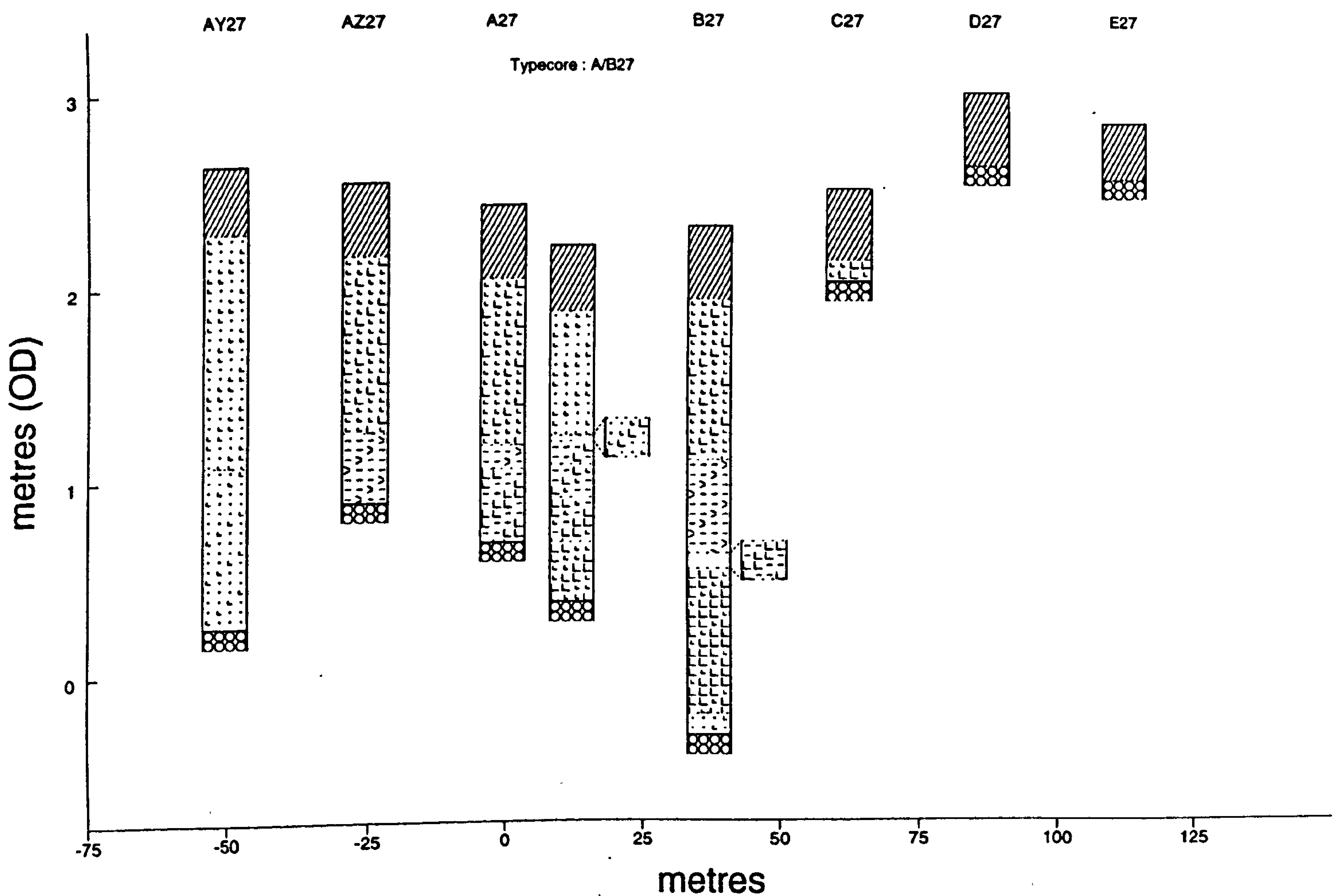


Figure 4.9: Stratigraphic Transect from which typecore A-B27 was selected (legend figure 4.2).

peat units occur between altitudes of approximately +0.50m OD and +1.20m OD. The oxidation mottled silt is present above the organic sediments and passes transitionally upward into the topsoil.

The gravel and peat surfaces are also displayed three-dimensionally, which enables the visualisation of the morphology of the area at the time of deposition of these sediments. The gravel surface (figure 4.10) illustrates that typecore AY17 was sampled where the gravel surface was at a relatively low altitude, and, thus, a greater depth of sediment is present above. Conversely, typecore A-B27 was sited on a rise in the gravel surface.

The peat surface (figure 4.11) illustrates that by the period of peat formation, the ground surface at the two sites was at a similar altitude. This reflects the infilling of the main back-barrier environment with the blue-grey silts preceding peat formation. Figure 4.11 also demonstrates that the peat in this area formed as a semi-continuous layer which abutted ridge 1 in the south-east of the area. The peat surface diagram also illustrates that some of the smaller gravel features occurred as gravel islands protruding through the peat, suggesting that the gravel and peats were not synchronous in their formation.

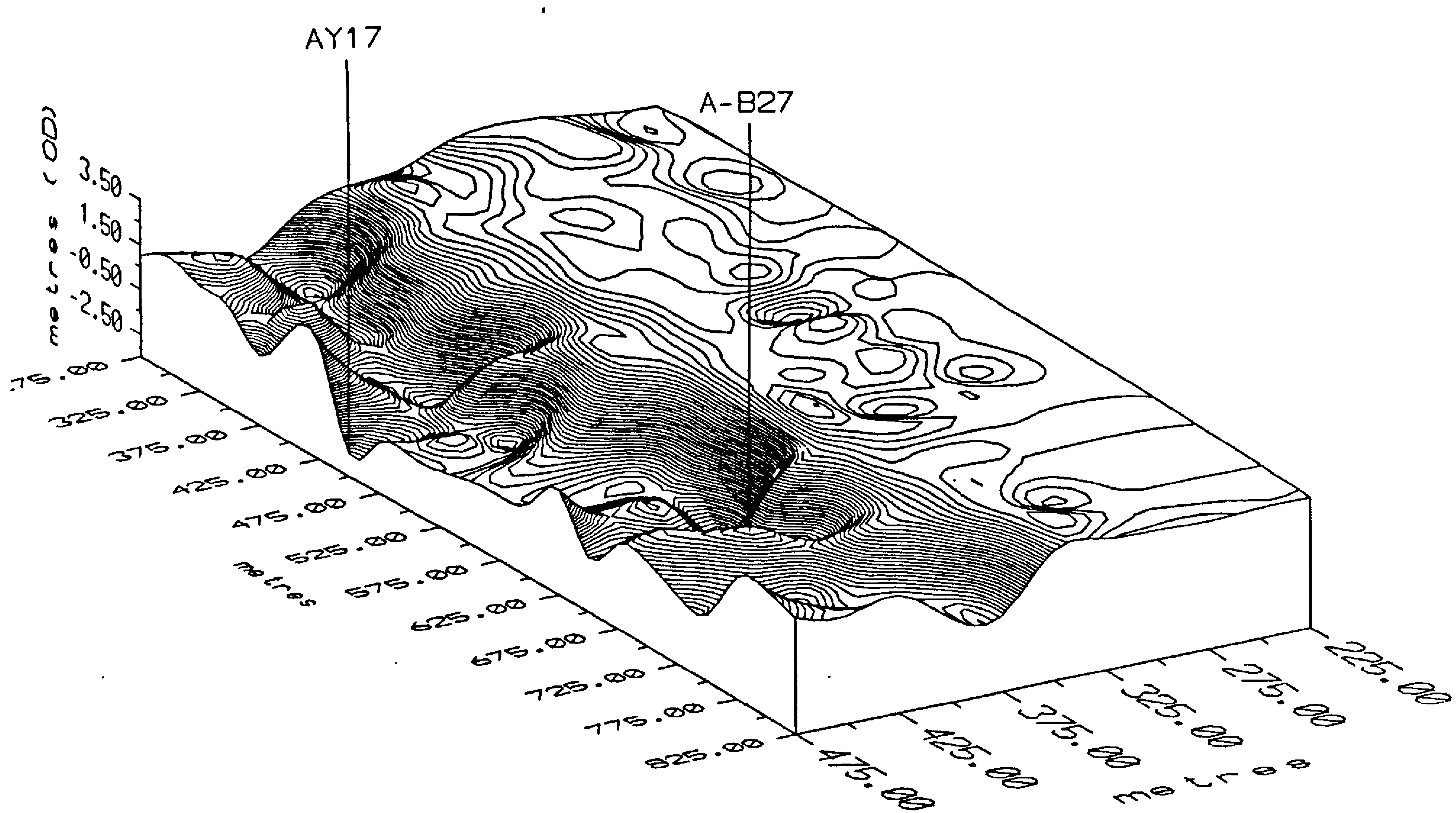


Figure 4.10: Three-dimensional gravel surface, typecores AY17 and A-B27.

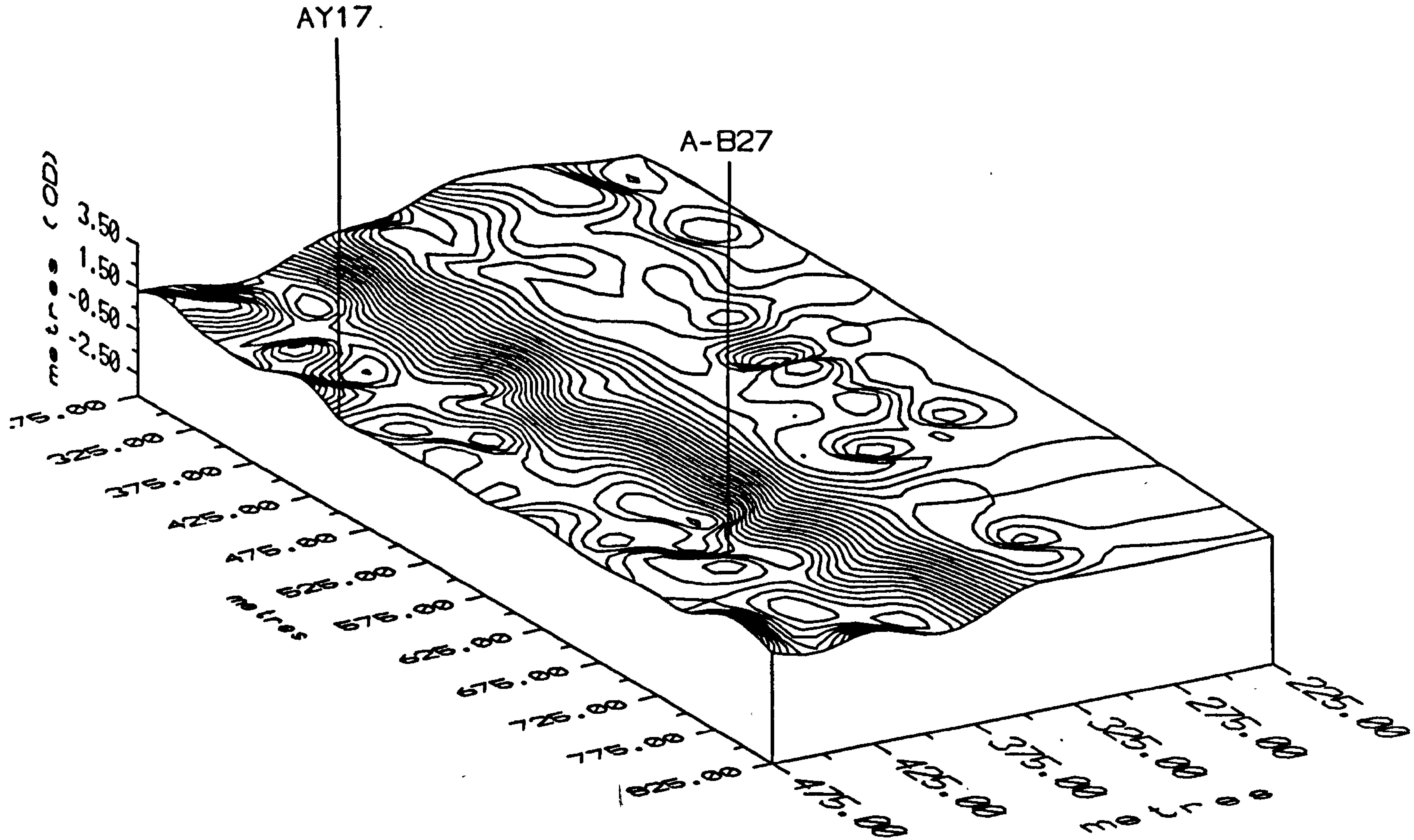


Figure 4.11: Three-dimensional peat surface, typecores AY17 and A-B27.

4.2.1.4 Lithostratigraphy of Typecore G60.

In figure 4.12, the sedimentary sequences characteristic of the Scotney Marsh trough and also the margins of the gravel ridges are presented. Cores E60, F60 and F-G60 exhibit blue-grey silts overlying the gravel surface, above which are, in turn, the peaty-clay, oxidation mottled silts and the topsoil. This sedimentary sequence is typical of gravel margin locations where peat and peaty-clay units are present in contrast to the typical sedimentary record of the Scotney Marsh trough. The cores G60 - N60 have a sedimentary suite typical of the Scotney Marsh trough, in which the gravel surface is overlain by blue-grey silts, oxidation mottled silts and topsoil. It is important to note that no organic sediments were recorded between +0.50m OD and +1.20m OD between cores G60 and N60.

Towards the base of typecore G60, sediments unique to this location were recorded. Organic sediments below -2.00m OD are present above the gravel surface, *i.e.* a peaty-silt overlain by peat. Above the peat, a succession of minerogenic sediments exists, with silty-clays overlain by gravel and then silts and gravels. However, further organic sediments just above -2.00m OD are present, *i.e.* a peat unit overlain by peaty-clay and then another peat. Above this unusual sequence, the suite of sediments typical of the Scotney Marsh trough is present; the blue-grey silts, oxidation mottled silts and topsoil.

Figures 4.13 and 4.14 show the gravel and peat surfaces. The gravel surface (figure 4.13) illustrates that

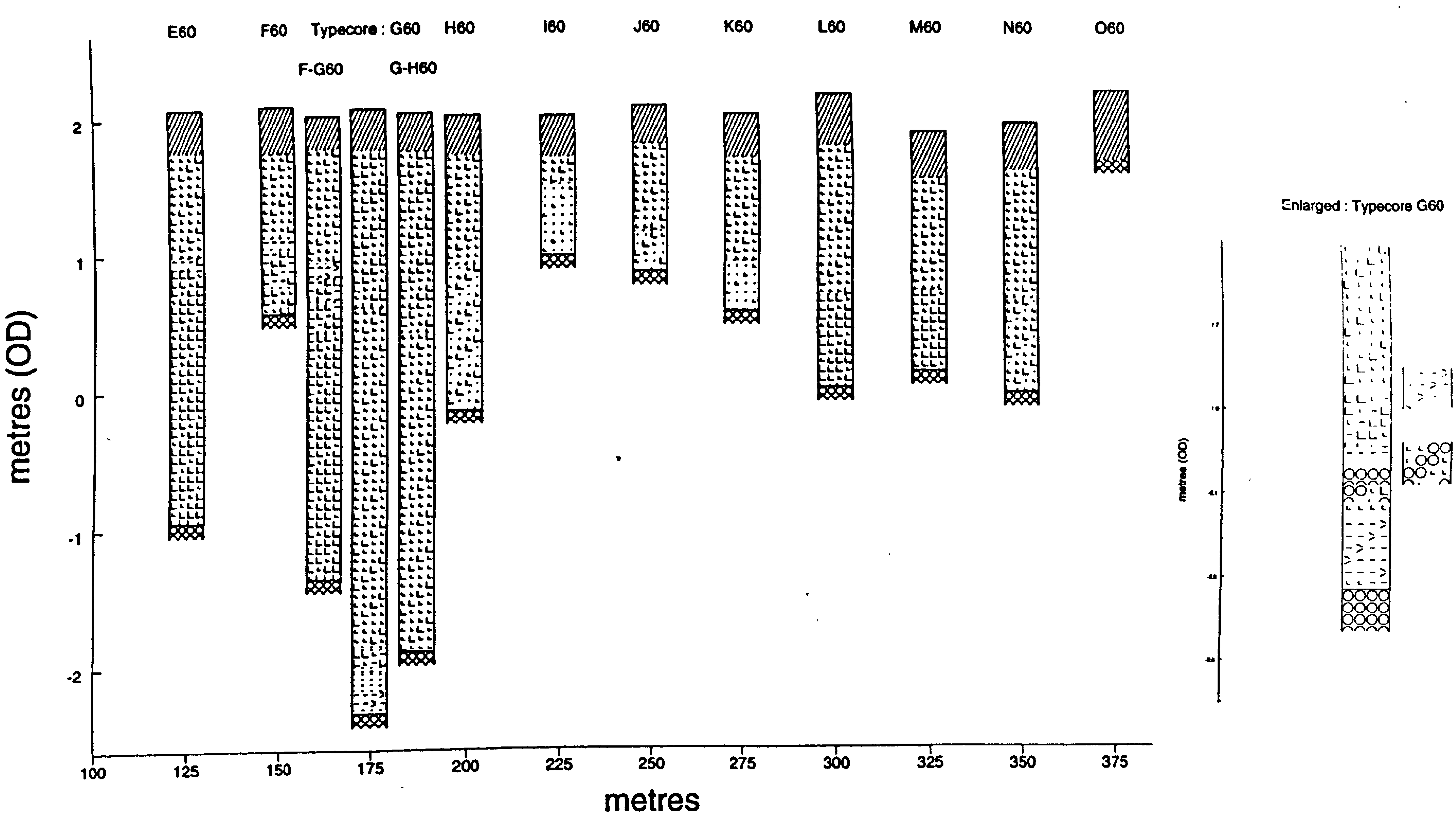


Figure 4.12: Stratigraphic Transect from which typecore G60 was selected (legend figure 4.2).

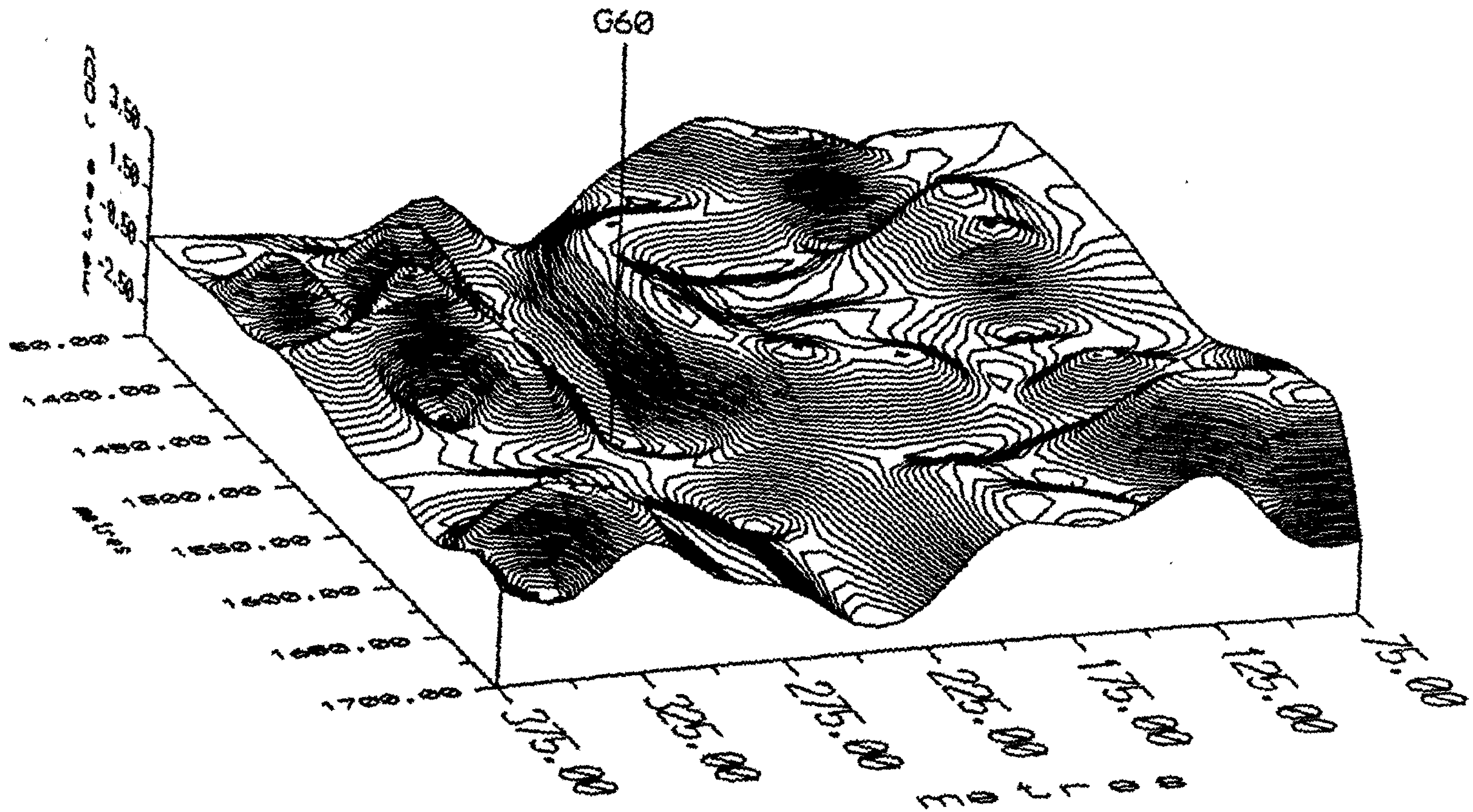


Figure 4.13: Three-dimensional gravel surface, typecore G60.

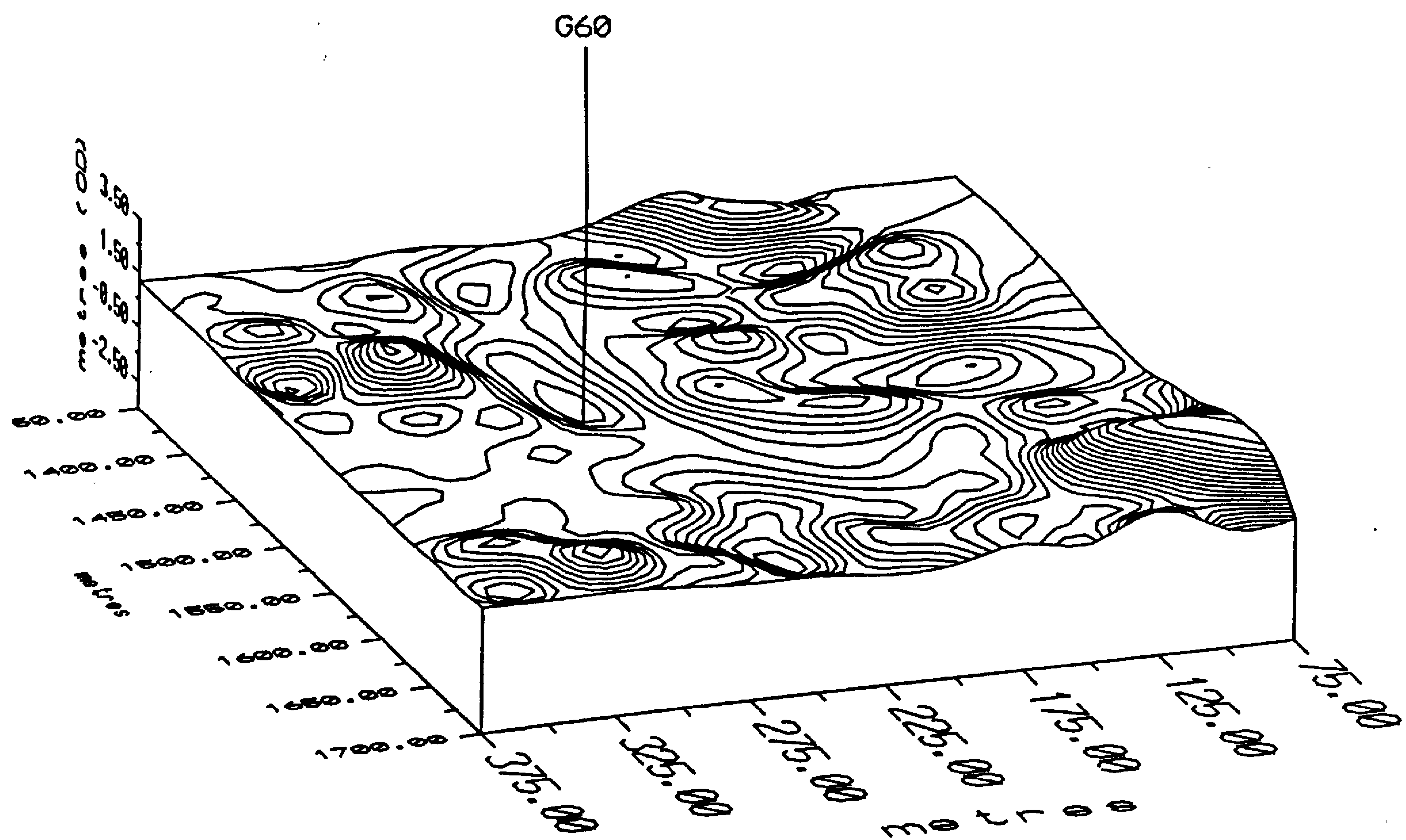


Figure 4.14: Three-dimensional peat surface, typecore G60.

typecore G60 is sited in a trough, which is at the eastern end of the Scotney Marsh trough (figure 4.13). Clearly, typecore G60 was sampled in a hollow in one of the lowermost areas of the Scotney Marsh trough. This may account for the unique suite of sediments. In the south-east of figure 4.13 is ridge 2, and in the north-west of the diagram the southern margin of ridge 1 is present.

The peat surface (figure 4.14) again illustrates that the peat occurred as a layer abutting the gravel ridges. However, it is important to note that where peat was not present, the blue-grey / oxidation mottled silt contact was used instead to create the peat surface (see section 3.1.3).

4.2.1.5 Lithostratigraphy of Typecores AW63 and AW-AX67.

The typecores AW63 and AW-AX67 were selected from the gravel inlet to the west of the Brett Gravel plant site (figure 4.7) as within this sheltered gravel inlet a different succession to that present across the remainder of the main back-barrier environment in Scotney Marsh is recorded. The stratigraphic transects of both the typecores are very similar (figures 4.15 and 4.16), cutting across similar sedimentary sequences across the Scotney Marsh area. Cores AI63 - AM63 and AI67 - AN67 possess a suite of sediments typical of the main back-barrier environment, with the gravel surface overlain, in turn, by blue-grey silts, peaty-clay and peat between +0.50m OD and +1.20m OD, the oxidation mottled silts and the topsoil.

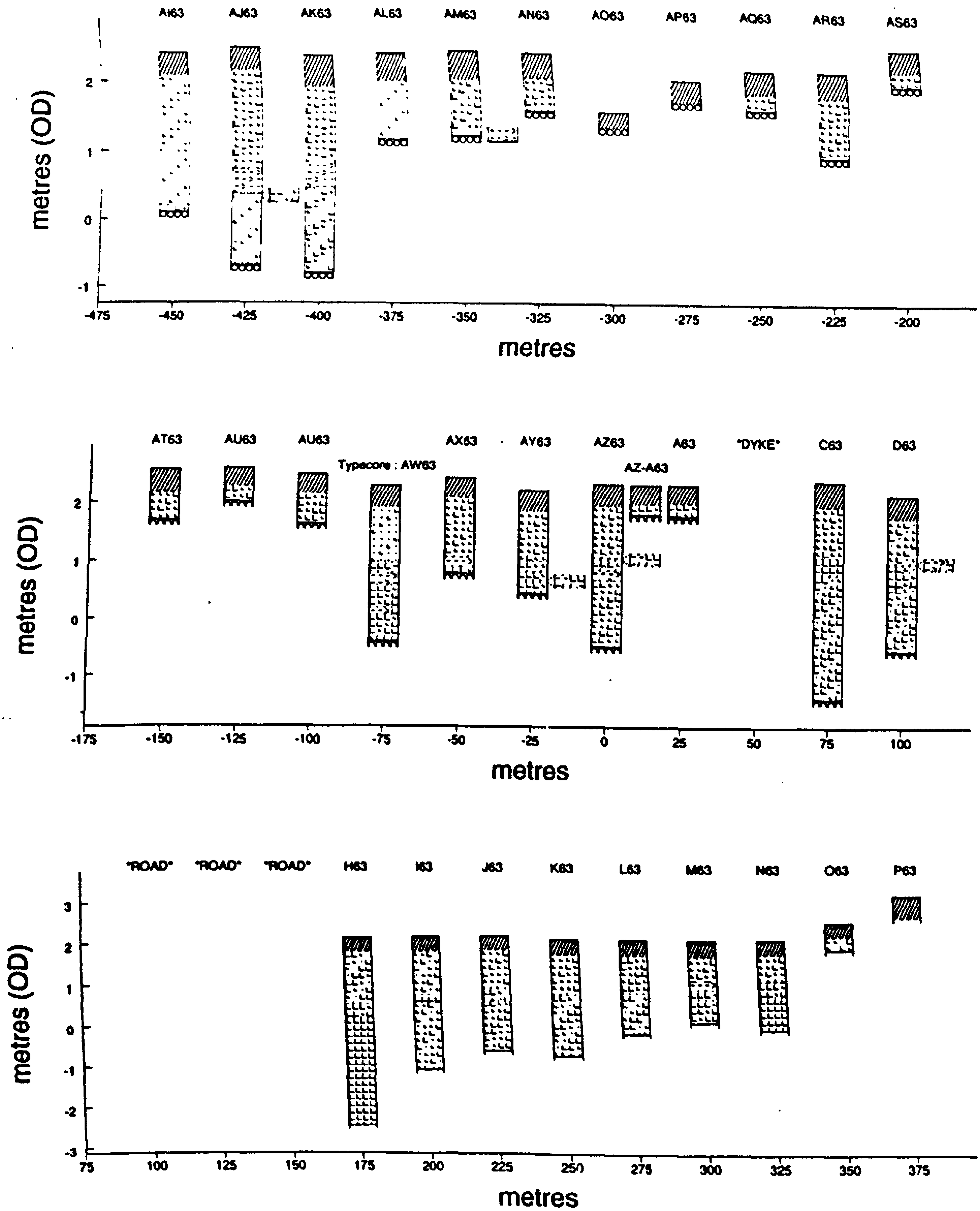


Figure 4.15: Stratigraphic Transect from which typecore AW63 was selected (legend figure 4.2).

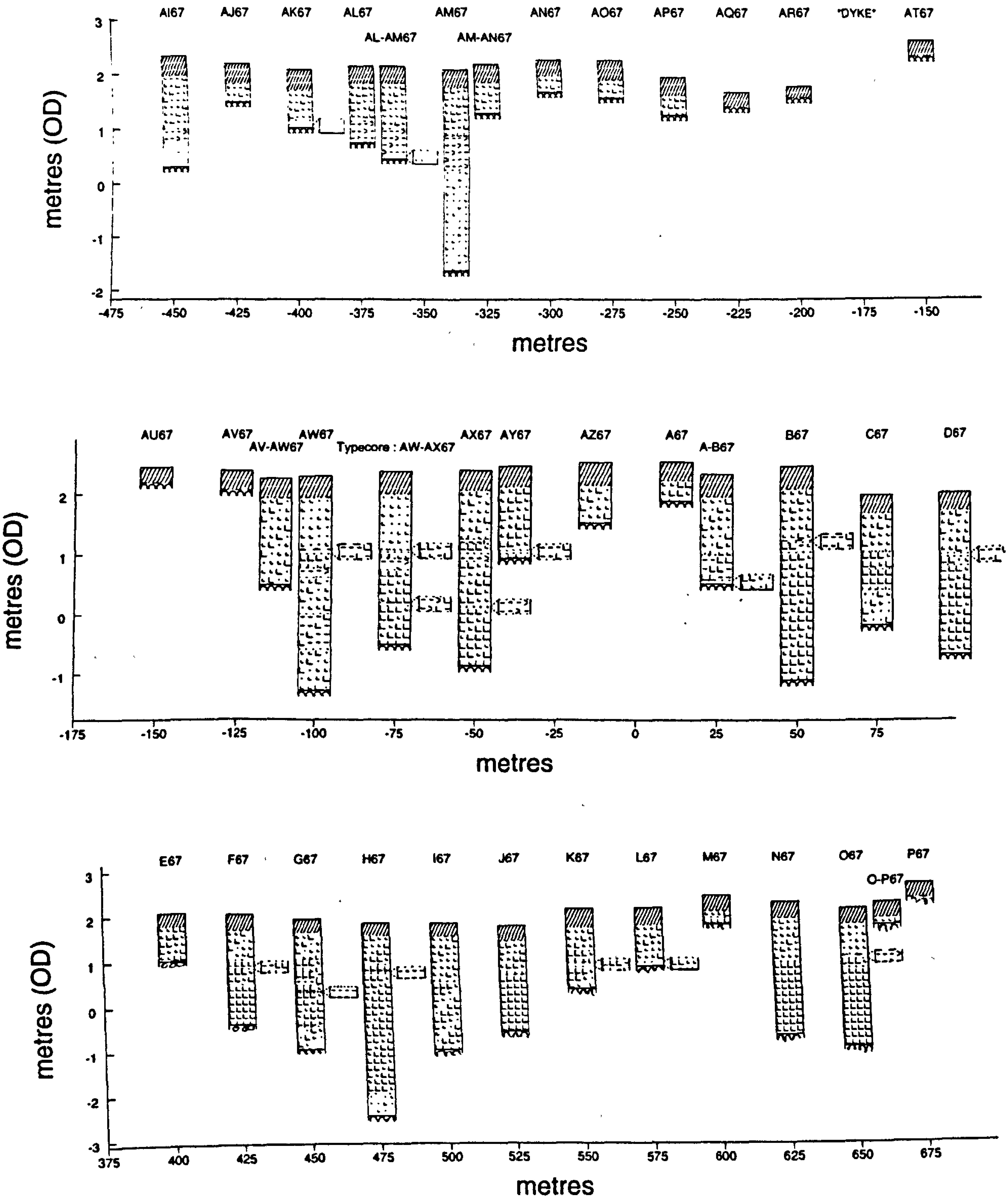


Figure 4.16: Stratigraphic Transect from which typecore AW-AX67 was selected (legend figure 4.2).

South-east along the strike of the stratigraphic transects, ridge 1 is recorded in cores AN63 - AV63 and AO67 - AV67 respectively. The gravel inlet is present parallel to ridge 1 (AW63 - AZ63 and AV-AW67 - AY67), from which the two typecores were selected. This suite of sediments is similar to those of the main back-barrier environment. However, in a number of the cores, a silty-clay unit occurs within the organic sediments.

Further south-east, the relatively smaller gravel ridge (Gb) is recorded in cores AZ-A63 - A63 and AZ67 - A67, adjacent to which are sediments of the Scotney Marsh trough. Cores C63 - N63 and A-B67 - L67 delimit the Scotney Marsh trough with its typical stratigraphic record. However, a number of cores, for example M63 and A-B67, contain organic sediments at ca. +1.00m OD. This is not typical of the Scotney Marsh trough but, as already discussed, organic sediments do occur in sheltered gravel inlets and at the gravel ridge margins. This is the case for the sediments of both cores M63 and A-B67 which exist on the northern margin of ridge 2 and at the southern margin of Gb respectively.

The three-dimensional image of the gravel surface for typecores AW63 and AW-AX67 (figure 4.17) assists in the visualisation of the gravel inlet recorded in the stratigraphic transects. In the north-west of figure 4.17, the surface of ridge 1 is recorded, whereas, to the south-east, the relatively small steep-sided morphology of Gb is illustrated. Between these two gravel ridges there is a

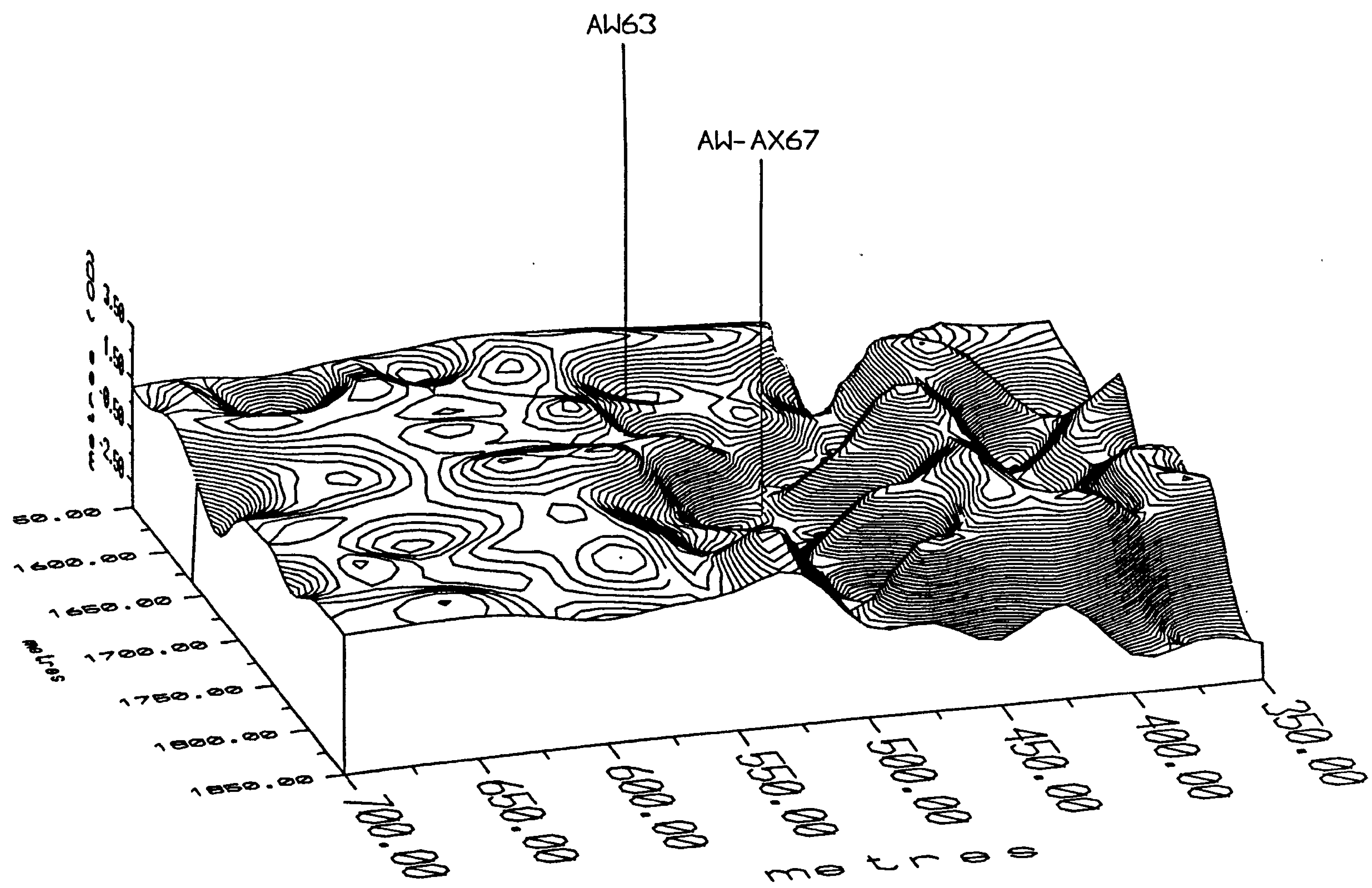


Figure 4.17: Three-dimensional gravel surface, typecores AW63 and AW-AX67.

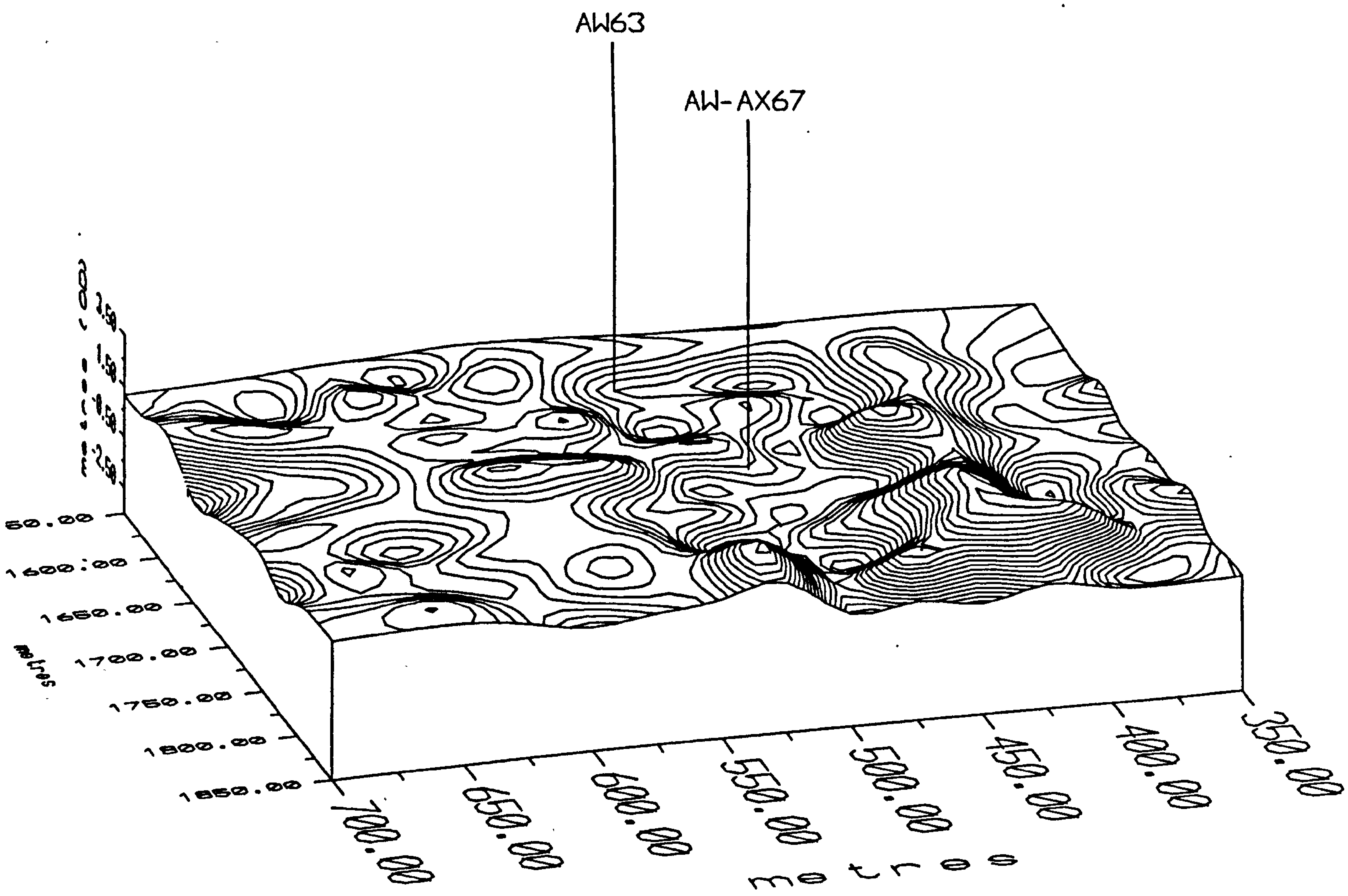


Figure 4.18: Three-dimensional peat surface, typecores AW63 and AW-AX67.

clear depression containing both typecores AW63 and AW-AX67. This is the gravel inlet, *i.e.* a relatively enclosed channel feature.

Typecore AW63 is situated toward the back of the gravel inlet on a relatively elevated gravel surface. This is in contrast to typecore AW-AX67 which is situated toward the centre of the gravel inlet where the gravel surface is relatively depressed. The three-dimensional image of the peat morphology (figure 4.18) illustrates that at the end of peat deposition, the gravel inlet was considerably less well-defined. This represents the infilling of the gravel inlet, first by the blue-grey silts and then by peat. Figure 4.18 also demonstrates that the gravel of ridge 1 and Gb would have protruded through the peat as gravel islands.

4.2.1.6 Longitudinal Lithostratigraphy of Scotney Marsh.

A longitudinal stratigraphic transect, the location of which is illustrated in figure 4.7, follows a line near-parallel with the strike of the gravel ridges in Scotney Marsh. This transect is perpendicular to the stratigraphic transects of the typecores discussed above and, thus, crosses the same sedimentary sequences but gives a different perspective on their inter-relationships. Observations of the stratigraphy of Scotney Marsh parallel to the gravel ridges of the area will enable any changes in the environment of deposition that occur along this transect to be determined.

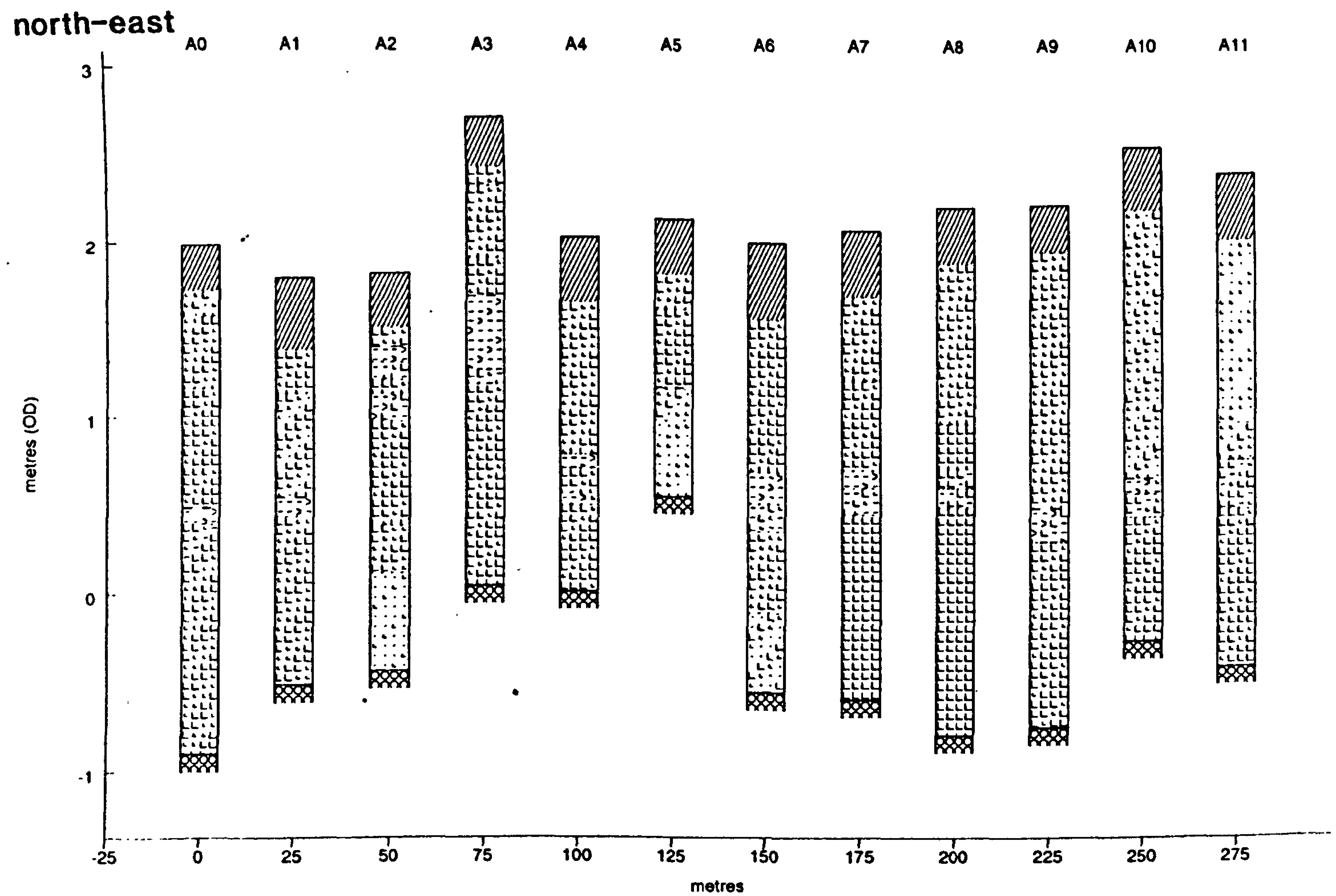


Figure 4.19a: Longitudinal stratigraphic transect through Scotney Marsh (legend figure 4.2).

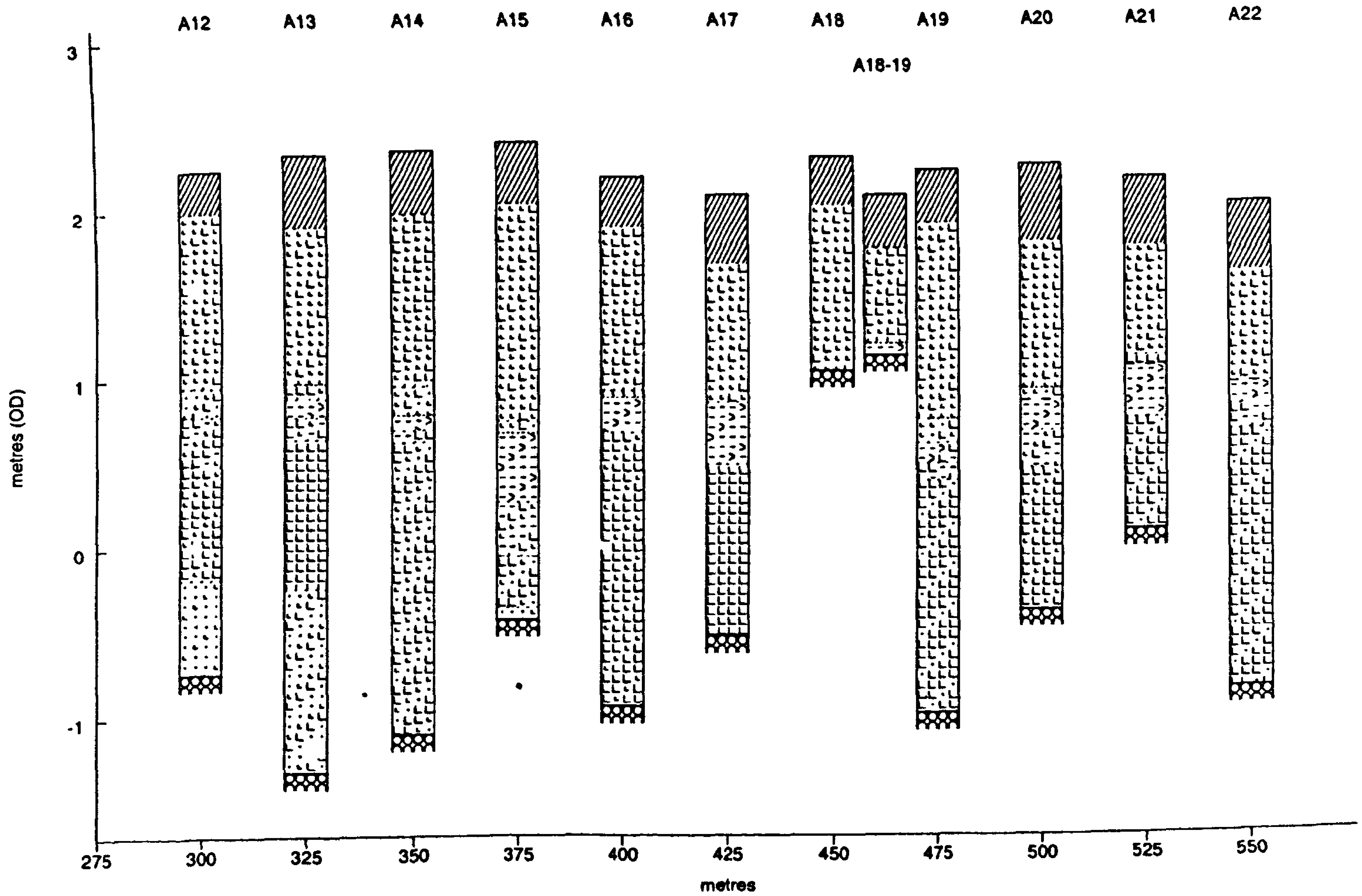


Figure 4.19b: Longitudinal stratigraphic transect through Scotney Marsh (legend figure 4.2).

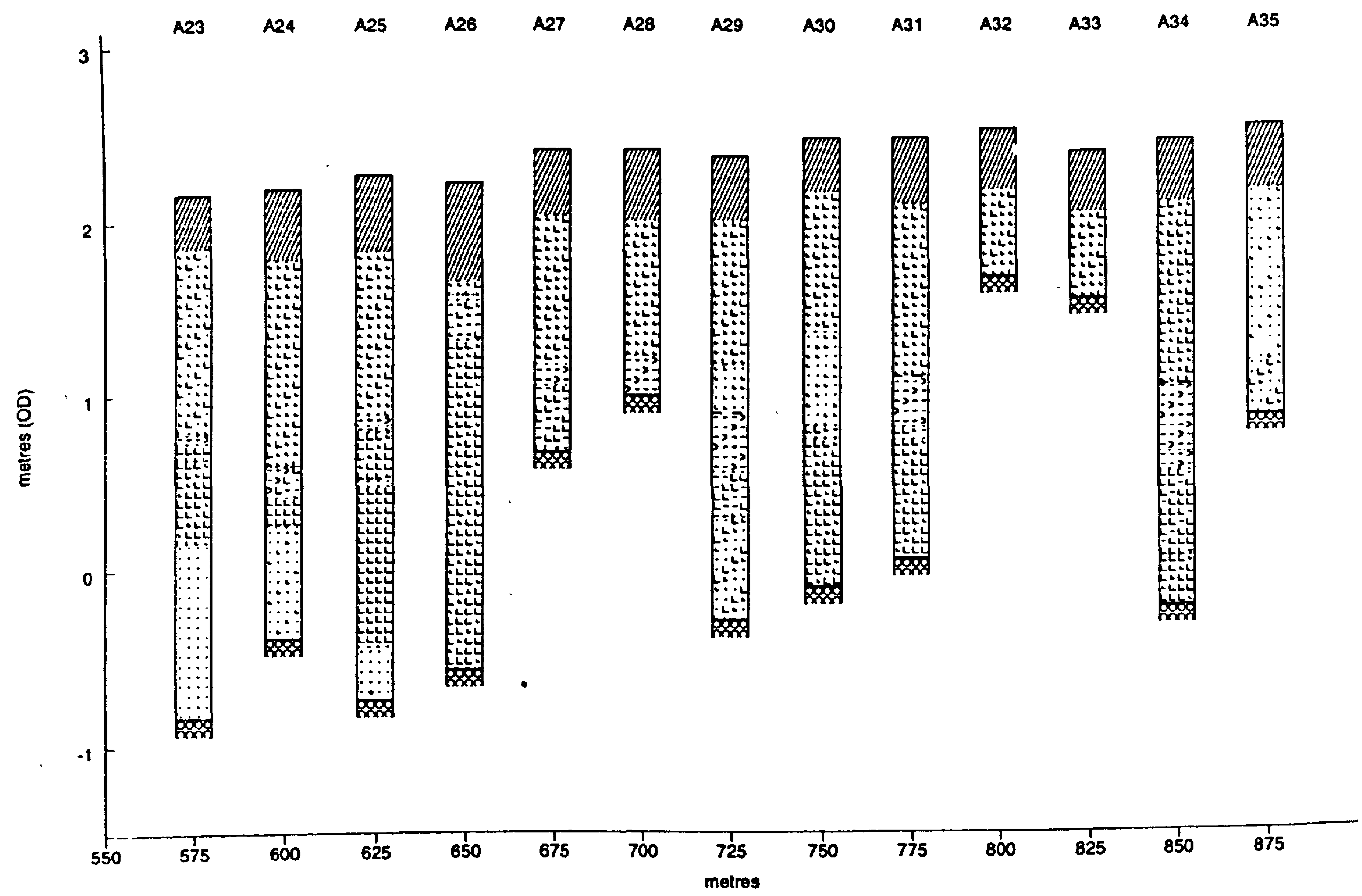


Figure 4.19c: Longitudinal stratigraphic transect through Scotney Marsh (legend figure 4.2).

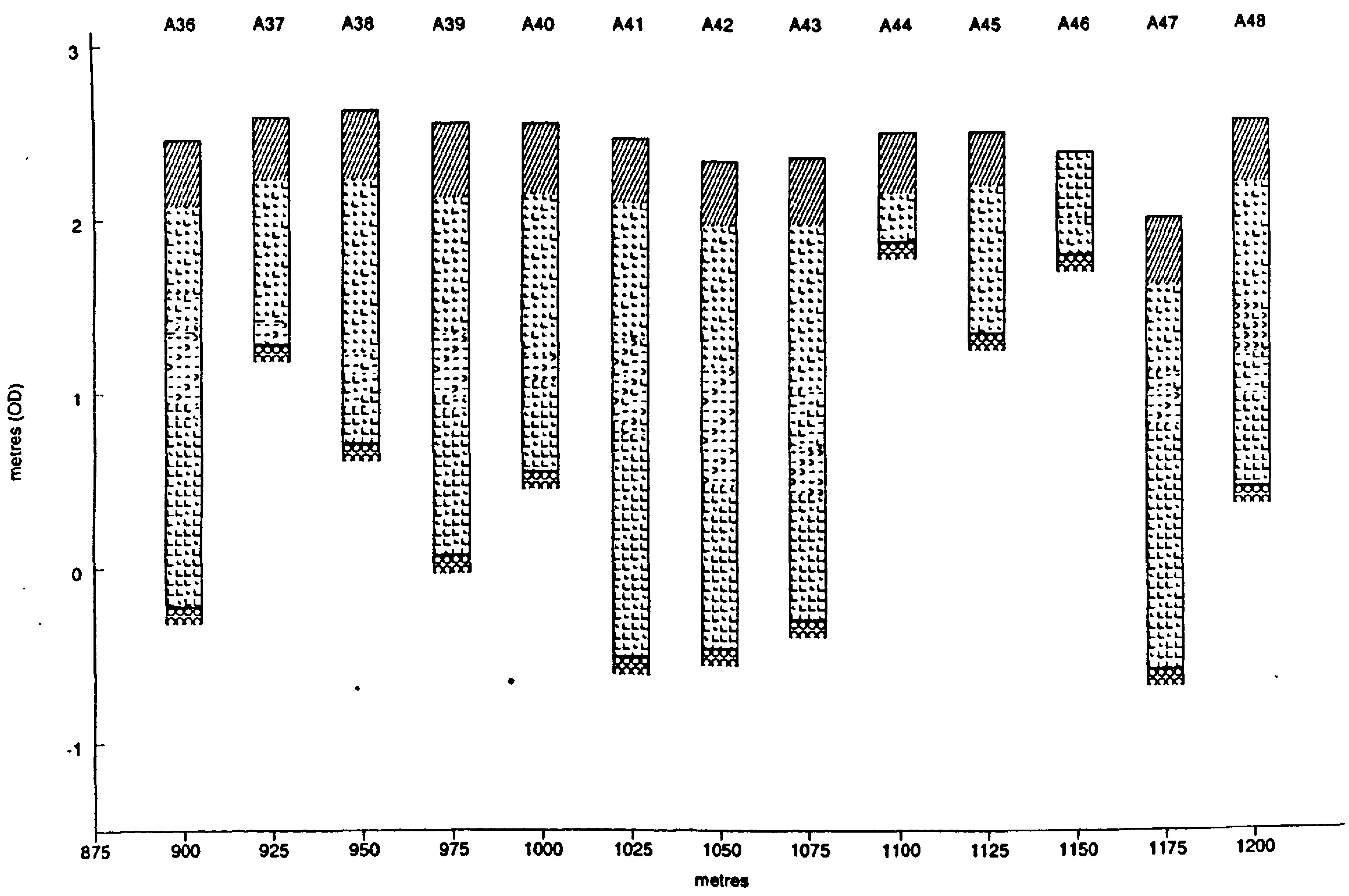


Figure 4.19d: Longitudinal stratigraphic transect through Scotney Marsh (legend figure 4.2).

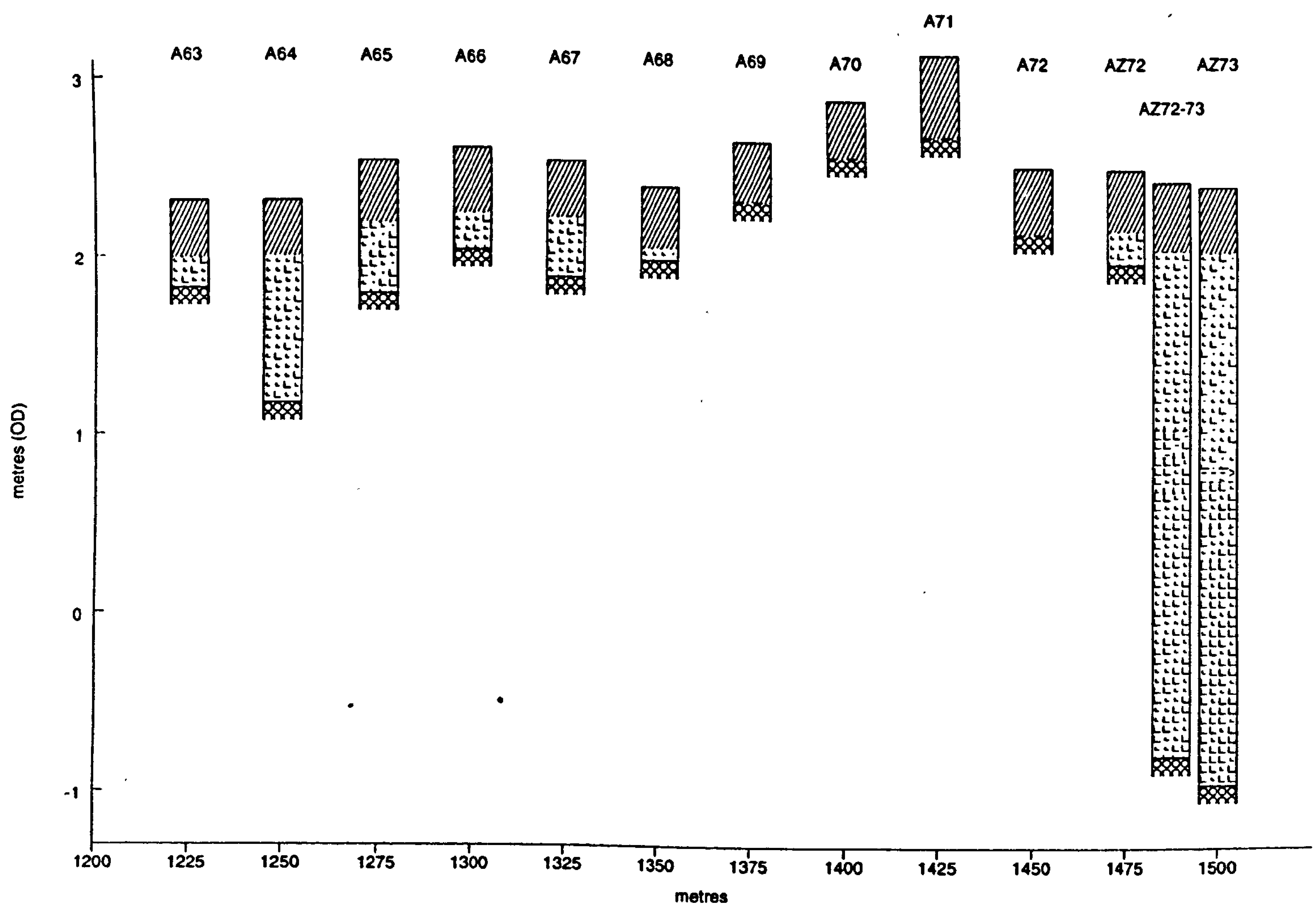


Figure 4.19e: Longitudinal stratigraphic transect through Scotney Marsh (legend figure 4.2).

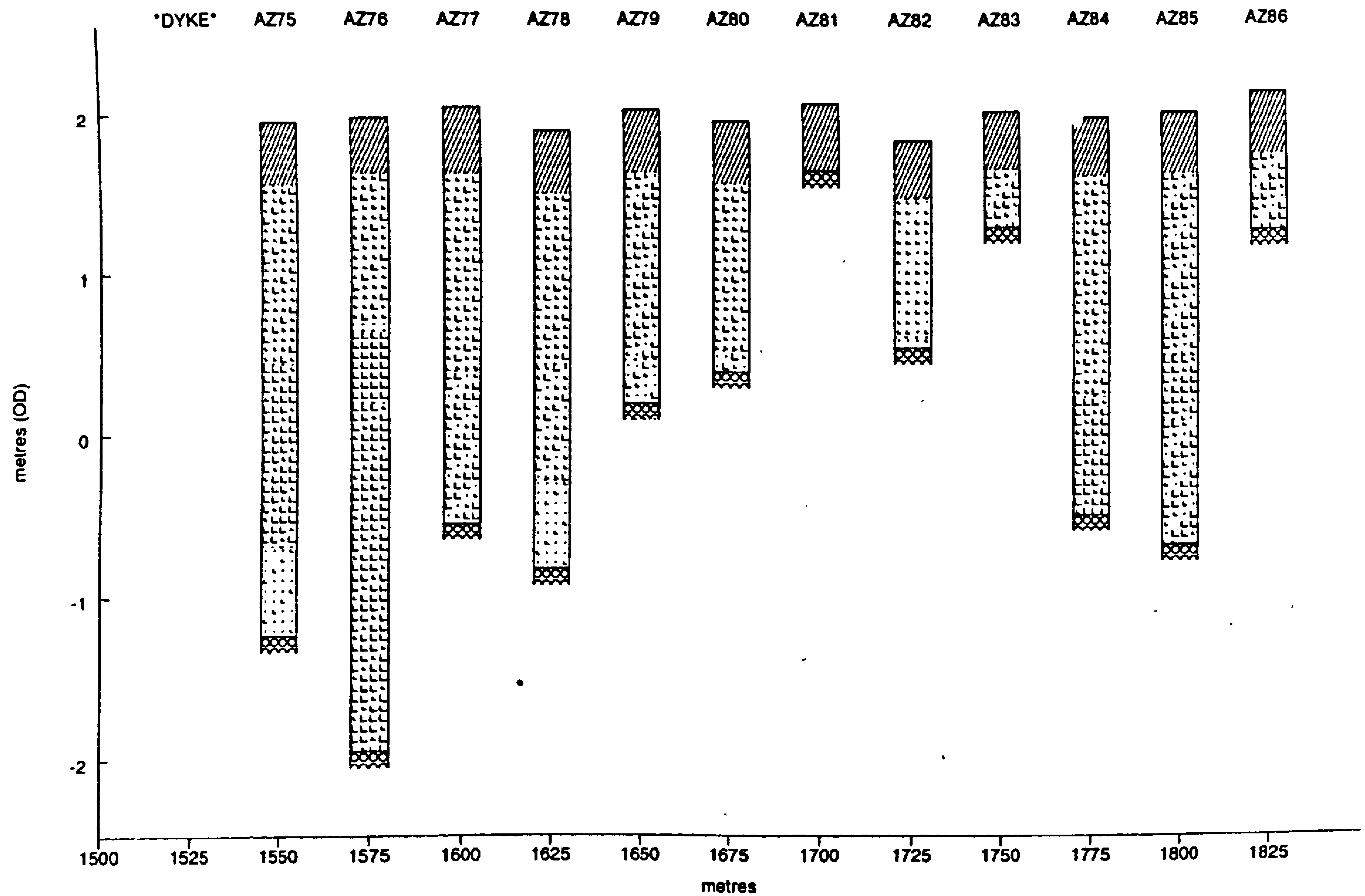


Figure 4.19f: Longitudinal stratigraphic transect through Scotney Marsh (legend figure 4.2).

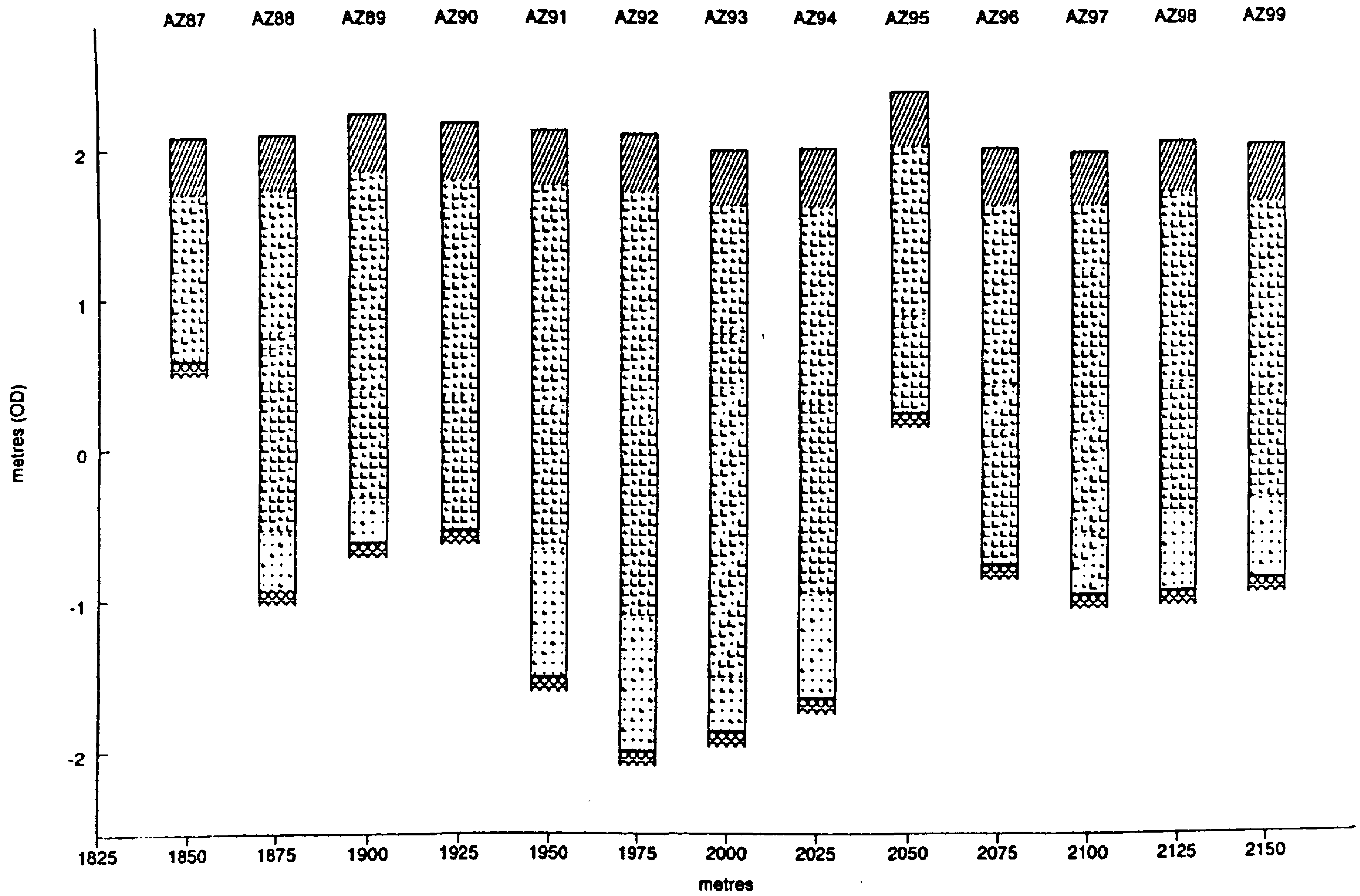


Figure 4.19g: Longitudinal stratigraphic transect through Scotney Marsh (legend figure 4.2).

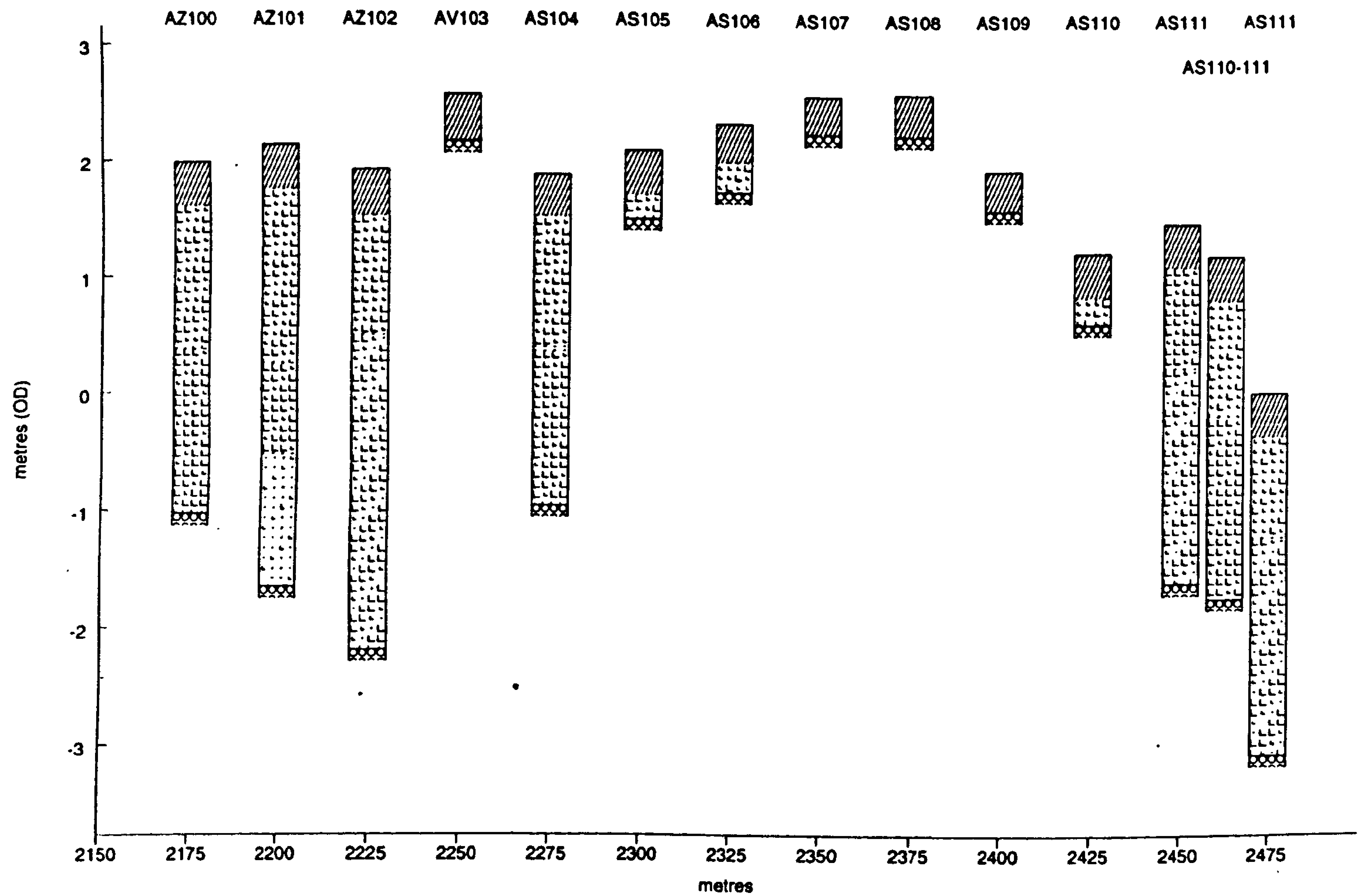


Figure 4.19h: Longitudinal stratigraphic transect through Scotney Marsh (legend figure 4.2).

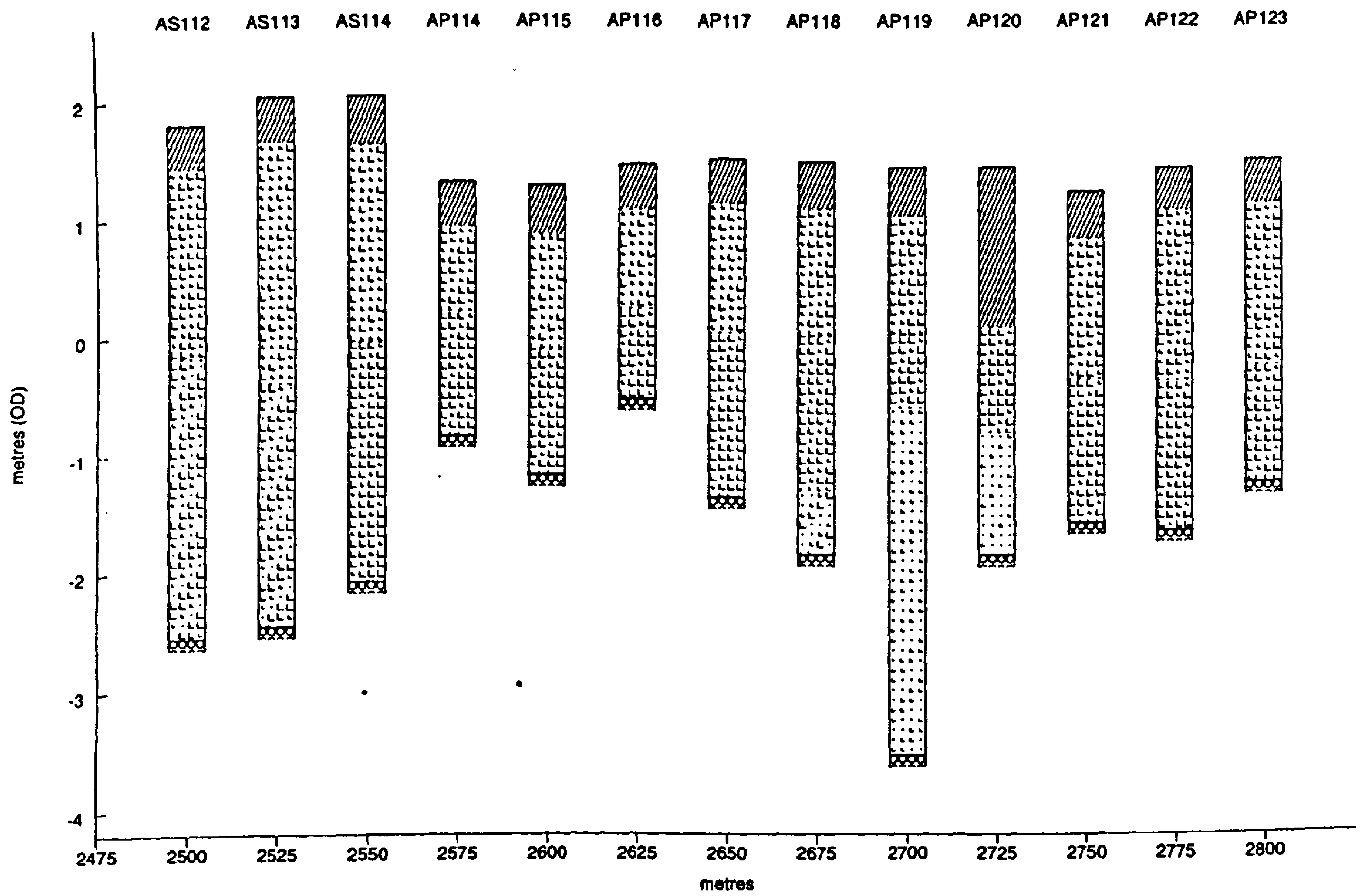


Figure 4.19i: Longitudinal stratigraphic transect through Scotney Marsh (legend figure 4.2).

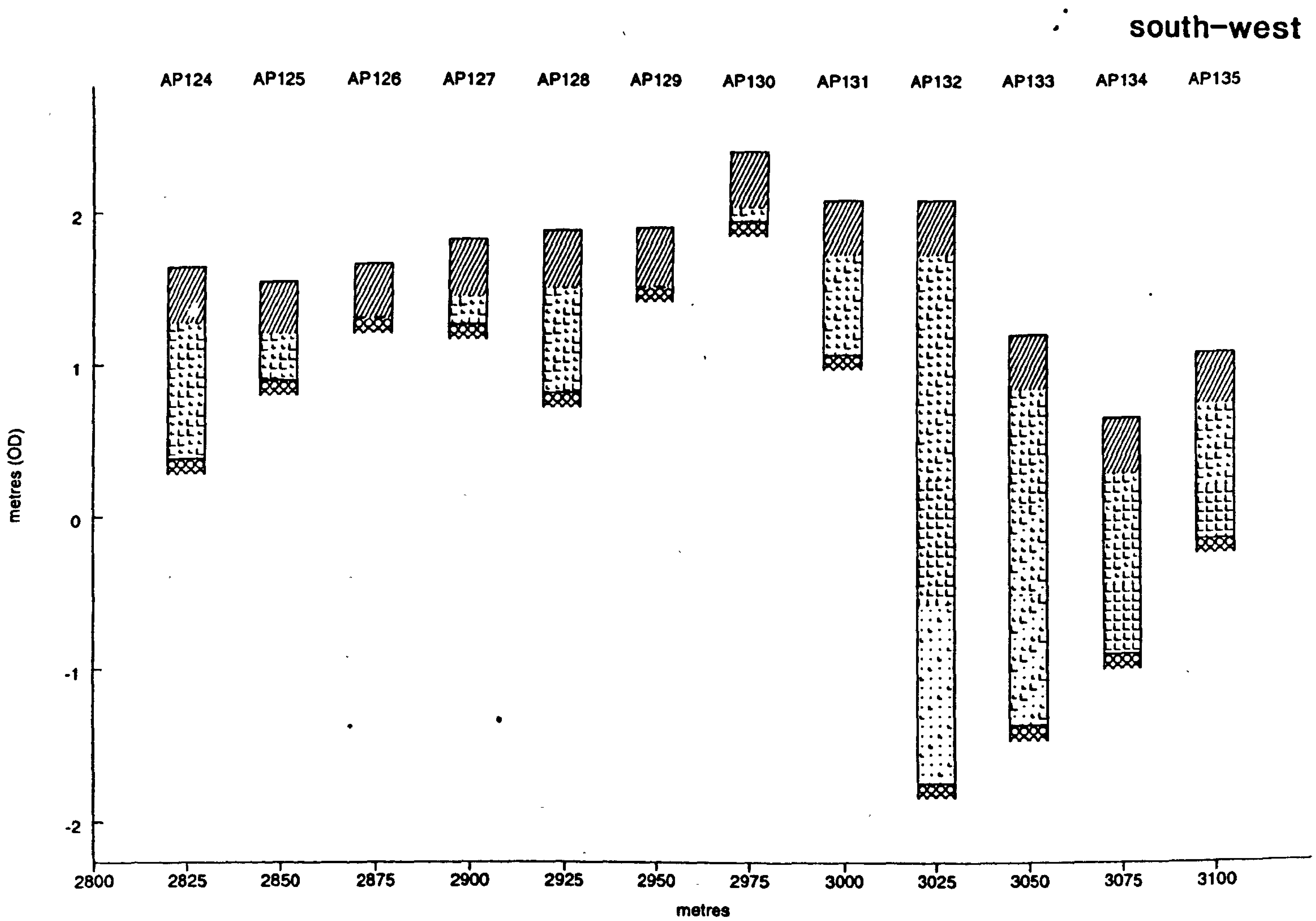


Figure 4.19j: Longitudinal stratigraphic transect through Scotney Marsh (legend figure 4.2).

In the eastern extreme of the transect, the sediments are typical of the main back-barrier environment, with most cores containing peats and peaty-clays, *i.e.* cores A0 - A48 (see figures 4.19a-d). In a number of the cores, *i.e.* A18 - A18-19 and A44 - 46, the gravel morphology is relatively elevated. This indicates that a number of smaller gravel ridges (offshoot limbs) extend at right angles from ridge 1 into the main back-barrier environment.

Between cores A48 and A63, no data are available due to the location of the Brett Gravel plant site. However, to the west of the plant site (figure 4.19e), cores A63 - AZ72 indicate that the gravel morphology is higher, with only one core extending below +1.50m OD. The suite of sediments above the gravel are oxidation mottled silts and topsoil only. Indeed, in some cores a gravelly topsoil occurs directly on the gravel surface. Here, the transect is situated along the strike of gravel ridge Gb. Further west, a significant change is recorded in the sediments as the gravel surface is sharply depressed at the location of cores AZ72-73 and AZ73. These cores are located in the Scotney Marsh trough, although they both contain organic sediments at c. +1.00m OD and are, therefore, typical of the stratigraphy found at gravel ridge margins in the study area.

Figure 4.19f illustrates that the gravel surface becomes further depressed, and cores AZ75 - AZ80 possess sedimentary sequences which are typical of the Scotney Marsh trough, *i.e.* blue-grey silts passing directly upward into oxidation mottled silts with no interdigitating organic sediments. Further towards the western end of the transect, a series of undulations on the gravel surface record the relatively small gravel ridges Gc and Ga in turn. Where the gravel surface is sufficiently depressed, sediments typical of the Scotney Marsh trough are present.

4.3 Palaeoenvironmental Reconstruction : Results.

In this section the results of the diatom, pollen and particle size analyses will be presented. For each of the typecores, the diatom analysis will be presented first, with the diatom diagram and a consideration of the autochthonous and allochthonous components of the assemblage of each sample. Secondly, the results of the pollen analysis will be presented, and then, thirdly, for two of the typecores (AY17 and G60), the results of particle size analysis are presented. Finally, the results of radiocarbon dating of a number of significant organic units from the typecores are presented.

4.3.1 Core AY17, Palaeoenvironmental Reconstruction : Results.

4.3.1.1 Core AY17 : Diatom analysis.

LDAZ AY17 - I : -2.25 - +0.02m OD

Predominant species : *Paralia sulcata, Thalassiosira excentrica, Actinoptychus senarius.*

Predominant ecological groups : *Melosira sulcata, Cymatosira belgica, Delphineis surirella.*

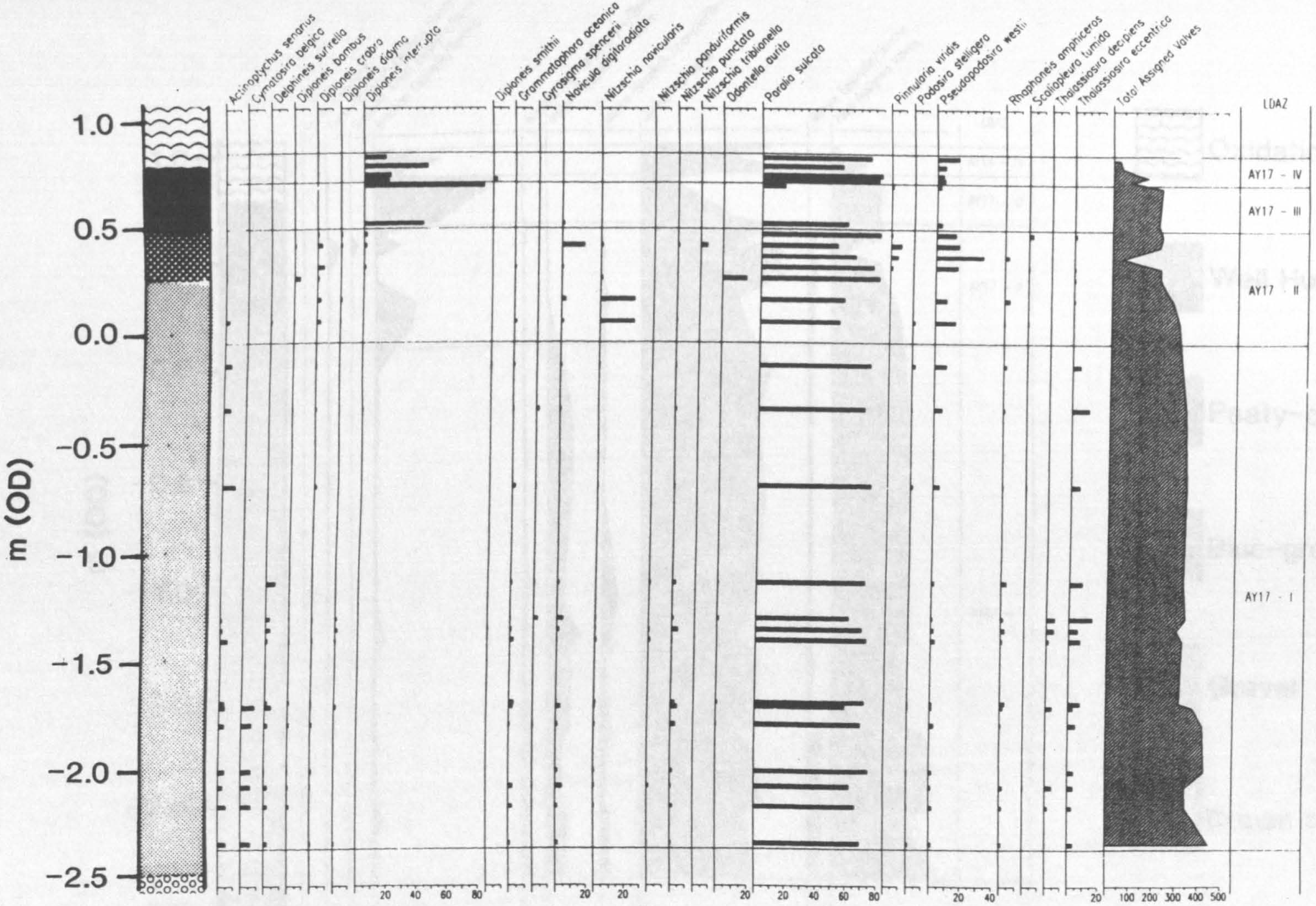


Figure 4.20a: Core AY17, Diatom Diagram : species (legend figure 4.20b).

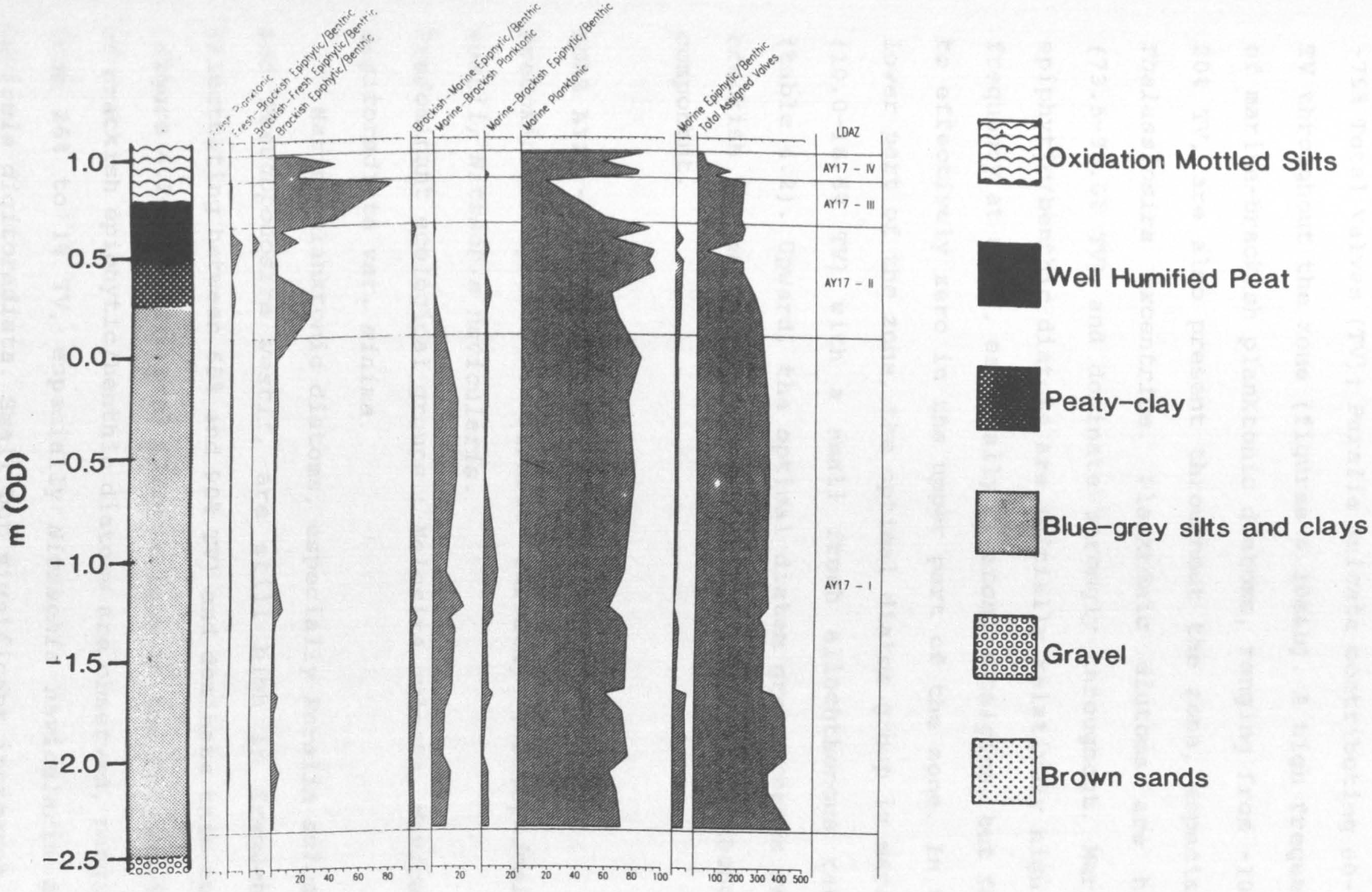


Figure 4.20b: Core AY17, Diatom Diagram : salinity tolerance groups.

Marine planktonic diatoms dominate in this zone at ~75% Total Valves (TV); *Paralia sulcata* contributing 60-75% TV throughout the zone (figures 4.20a&b). A high frequency of marine-brackish planktonic diatoms, ranging from ~10 to 20% TV, are also present throughout the zone, especially *Thalassiosira excentrica*. Planktonic diatoms are high (73.5-95.5% TV) and dominate strongly throughout. Marine epiphytic/benthic diatoms are initially relatively high in frequency at 8% TV, especially *Cymatosira belgica*, but fall to effectively zero in the upper part of the zone. In the lower part of the zone, the optimal diatom group is marine (10.0-16.6% TV) with a small fresh allochthonous ratio (table 4.2). Upward, the optimal diatom group ranges from brackish to marine with a variable fresh allochthonous component.

LDAZ AY17 - II : +0.02 - +0.55m OD

Predominant species : *Paralia sulcata, Pseudopodosira westii, Nitzschia navicularis*.

Predominant ecological groups : *Melosira sulcata, Navicula digitoradiata var. minima*.

Marine planktonic diatoms, especially *Paralia sulcata* and *Pseudopodosira westii*, are still high in frequency (fluctuating between 65% and 90% TV) and dominate this zone (figure 4.20a). Reciprocal fluctuations in the frequencies of brackish epiphytic/benthic diatoms are observed, ranging from 26% to 1% TV, especially *Nitzschia navicularis* and *Navicula digitoradiata*. Small but significant increases in the frequencies of fresh-brackish epiphytic/benthic taxa occur in this zone, particularly *Pinnularia viridis,* which reaches 8% TV. The percentage of planktonic diatoms dominate strongly (63.4-94.3% TV), but with increased percentages of epiphytic/benthic diatoms relative to the zone below. Initially the optimal diatom group is brackish (~25% TV) which becomes marine upwards and, eventually, fresh then brackish at the top of the zone (table 4.2).

LDAZ AY17 - III : +0.55 - +0.76m OD

Predominant species : *Diploneis interrupta, Paralia sulcata.*

Predominant ecological groups : *Diploneis interrupta, Melosira sulcata.*

The frequencies of marine planktonic diatoms are initially high, at 56% TV, but fall significantly to 19% TV with a reduction in the frequency of *Paralia sulcata* through the zone. A corresponding rise in the frequencies of brackish epiphytic/benthic diatoms from 43% to 81% TV occurs, especially *Diploneis interrupta* which dominates at the top of the zone (figures 4.20a&b). Significantly, the frequencies of marine epiphytic/benthic diatoms are reduced to zero in this zone. A considerable increase in the percentage of epiphytic/benthic diatoms into and through this zone, from 41.0% TV to 82.3% TV occurs upward in this zone, with the opposite trend for planktonic diatoms (59.0-17.7% TV). The optimal diatom group is brackish in this zone (table 4.2), which dominates with only a very low number of fresh allochthonous diatoms (0.011).

Sample Depth m (OD) LDAZ		Total diatoms % Epiphytic / Benthic	Total diatoms % Planktonic	Autochthonous diatoms (Optimal Epiphytic / Benthic group) *	Allochthonous diatoms (Fresh-water) *	Allochthonous diatoms (Marine) *
IV	+0.865	14.3	85.7	B - 14.3%	-	-
IV	+0.825	42.1	57.9	B - 42.1%	-	-
IV	+0.785	17.6	82.4	B - 17.6%	-	-
IV	+0.765	18.8	81.2	B - 16.5%	-	0.145
III	+0.735	82.3	17.7	B - 81.4%	0.011	-
III	+0.565	41.0	59.0	B - 41.0%	-	-
II	+0.515	8.3	91.7	B - 6.9%	0.072	0.145
II	+0.465	34.6	65.4	B - 22.0%	0.332	0.245
II	+0.415	7.1	92.9	F-B - 4.8%	-	0.500
II	+0.365	10.0	90.0	M - 5.7%	0.754	-
II	+0.315	5.7	94.3	M - 4.3%	0.326	-
II	+0.215	35.8	63.4	B - 25.7%	0.086	0.304
II	+0.115	29.0	71.0	B - 24.3%	0.012	0.181
I	-0.085	5.8	94.2	M-B - 3.8%	0.500	-
I	-0.285	8.7	91.3	M - 3.9%	1.231	-
I	-0.635	4.5	95.5	M - 2.1%	1.143	-
I	-1.075	17.6	82.1	M-B - 14.5%	0.207	-
I	-1.235	15.4	84.6	B - 8.9%	0.067	0.663
I	-1.285	16.4	83.6	M-B - 13.0%	0.254	-
I	-1.335	9.0	90.7	M - 5.6%	0.589	-
I	-1.615	11.7	88.3	M-B - 9.5%	0.232	-
I	-1.635	26.5	73.5	M - 16.6%	0.596	-
I	-1.715	18.5	81.5	M - 11.2%	0.652	-
I	-1.925	14.9	85.1	M - 10.0%	0.490	-
I	-1.995	23.0	77.0	M - 13.4%	0.716	-
I	-2.085	19.1	80.9	M - 13.0%	0.469	-
I	-2.255	17.7	82.1	M - 12.2%	0.443	-

Table 4.2 : Diatoms, Core AY17.

* Adapted from Beyens and Denys (1982).

LDAZ AY17 - IV : +0.76 - +0.87m OD

Predominant species : *Paralia sulcata, Diploneis interrupta, Pseudopodosira westii.*

Predominant ecological groups : *Melosira sulcata, Diploneis interrupta.*

Marine planktonic diatoms dominate, fluctuating between ~80% and 62% TV (figure 4.20b). This dominance is controlled by the very high frequencies of *Paralia sulcata.* Indeed, the planktonic diatoms dominate strongly at >80% TV at all but one level through the zone. Brackish epiphytic/benthic diatoms contribute significantly to the assemblage at ~25% TV, and reach a peak at 41% TV in the middle of the zone due to the high frequencies of *Diploneis interrupta.* The optimal diatom group is brackish throughout, with negligible contributions from either freshwater or marine allochthonous ratios (table 4.2).

4.3.1.2 Core AY17 : Pollen analysis.

LPAZ AY17 - I : +0.47 - +0.52m OD

Predominant species : Gramineae, Cyperaceae, Chenopodiaceae.

Initially Gramineae and Cyperaceae are high in frequency and dominate at 35% TLP and 26% TLP respectively (figure 4.21a). However, Cyperaceae rapidly decline through the zone to <5% TLP, whereas Gramineae increase to 71% TLP. A relatively high frequency of Chenopodiaceae (8% TLP) falls rapidly upward, whilst the frequencies of tree pollen generally increase upward, especially *Quercus* and *Alnus.*

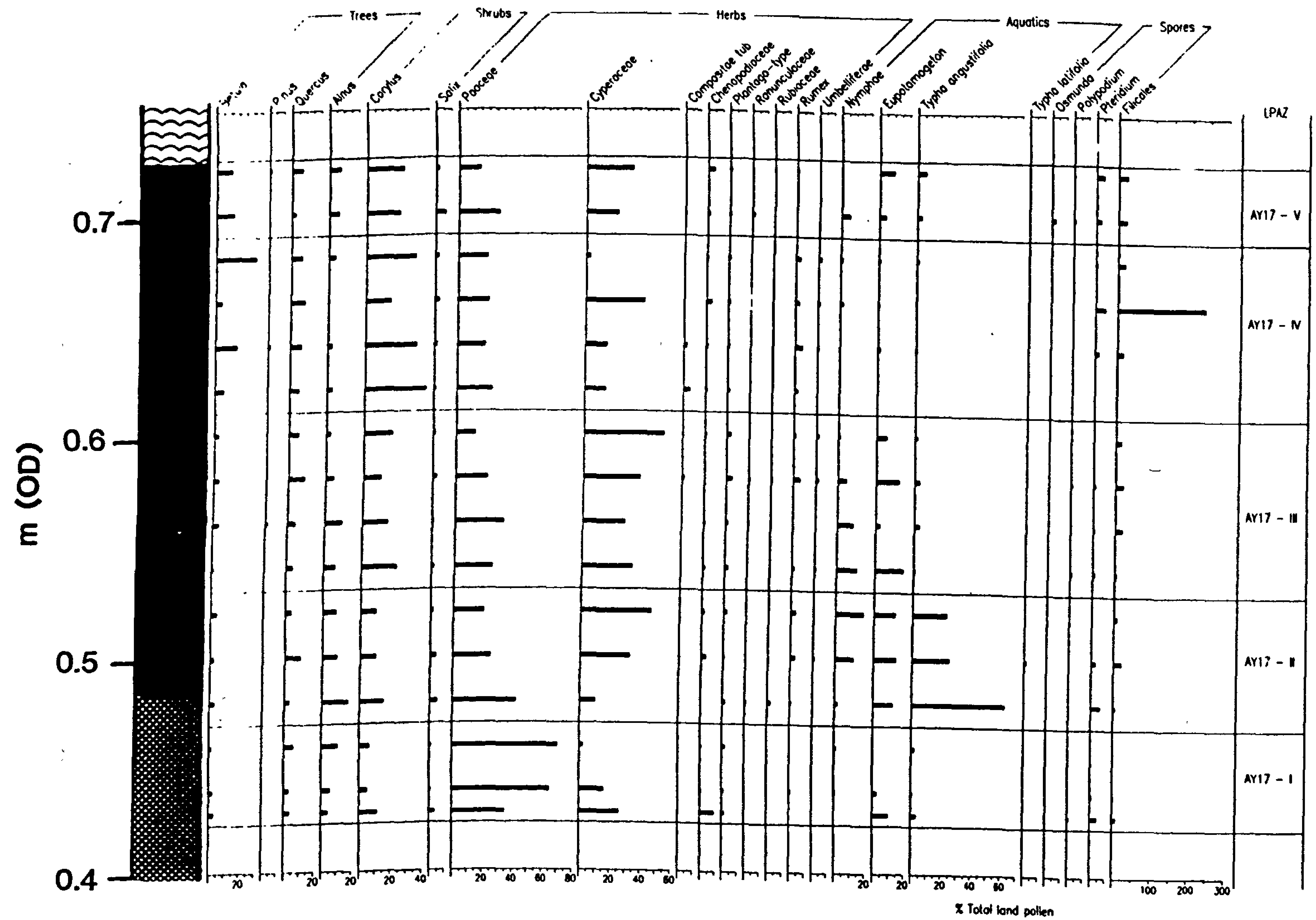


Figure 4.21a: Core AY17, Pollen Diagram : species (legend figure 4.20b).

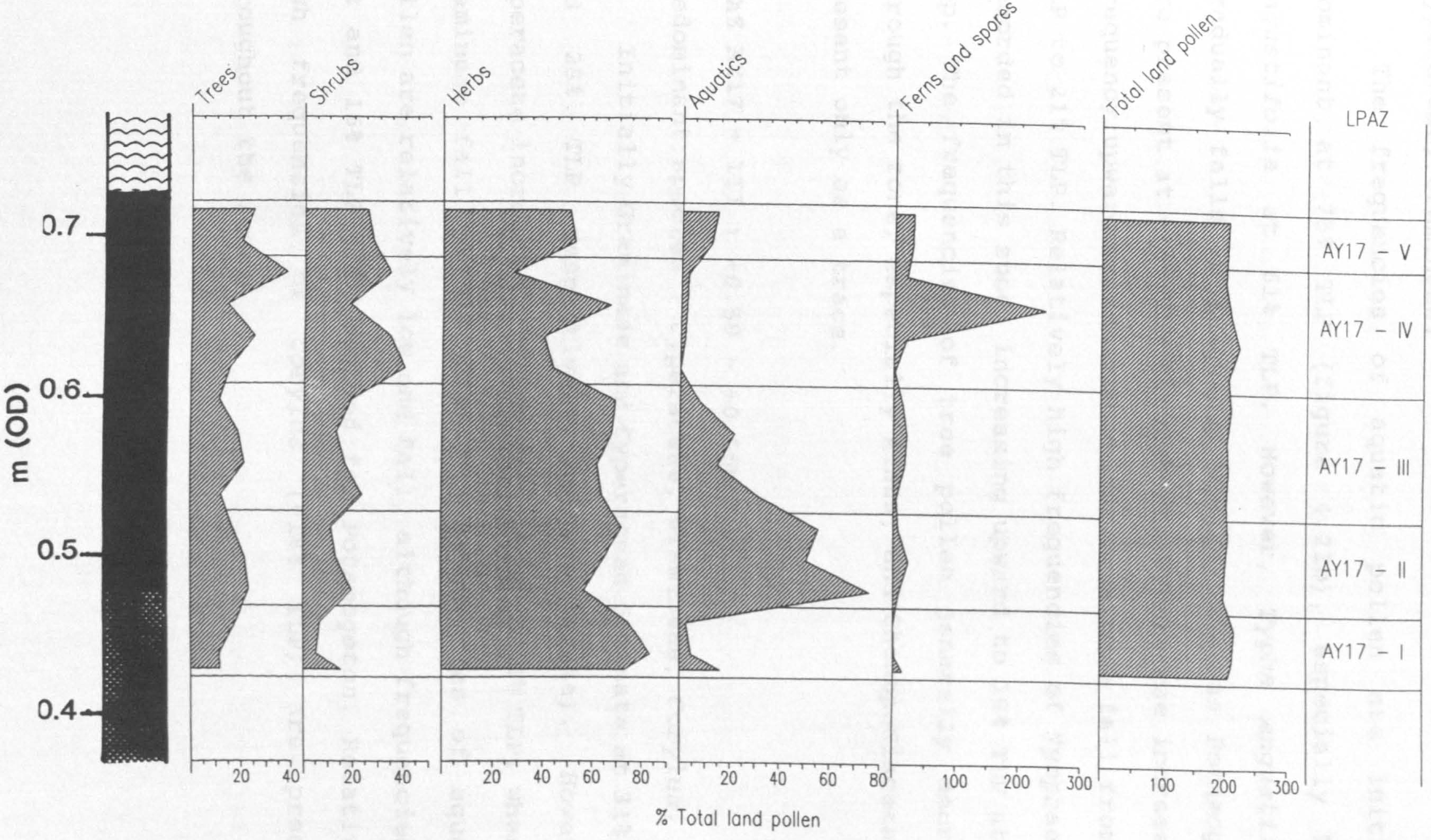


Figure 4.21b: Core AY17, Pollen Diagram : groups (legend figure 4.20b).

LPAZ AY17 - II : +0.52 - +0.59m OD

Predominant species : *Typha angustifolia, Potamogeton, Cyperaceae, Gramineae.*

The frequencies of aquatic pollen are initially dominant at 75% TLP (figure 4.21b), especially *Typha angustifolia* at 61% TLP. However, *Typha angustifolia* gradually falls in frequency upward, whereas *Potamogeton* are present at ~15% TLP throughout. Cyperaceae increase in frequency upward to 46% TLP whilst Gramineae fall from 44% TLP to 21% TLP. Relatively high frequencies of *Nymphae* are recorded in this zone, increasing upward to 19% TLP at the top. The frequencies of tree pollen generally decrease through the zone, especially *Alnus*, and Chenopodiaceae are present only as a trace.

LPAZ AY17 - III : +0.59 - +0.65m OD

Predominant species : Cyperaceae, Gramineae, *Corylus*.

Initially Gramineae and Cyperaceae dominate at 31% TLP and 28% TLP respectively (figure 4.21a). However, Cyperaceae increase in frequency upward to 52% TLP, whereas Gramineae fall to 11% TLP. The frequencies of aquatic pollen are relatively low and fall, although frequencies of 19% and 16% TLP are recorded for *Potamogeton*. Relatively high frequencies of *Corylus* (~18% TLP) are present throughout the zone.

LPAZ AY17 - IV : +0.65 - +0.73m OD

Predominant species : *Corylus*, *Betula*, Gramineae, Cyperaceae.

Initially *Corylus* are high in frequency and dominate at 41% TLP (figure 4.21a), but fall in frequency upward to 18% TLP before increasing again at the top of the zone. The frequencies of Gramineae are ~20% TLP throughout the zone, whereas Cyperaceae is initially low in frequency at 14% TLP, and increases to 40% TLP before falling to 2% TLP at the top of the zone. The frequencies of *Betula* are relatively high, reaching a peak value of 26% TLP at the top of the zone. A very high frequency of Filicales (250% TLP) is recorded.

LPAZ AY17 - V : +0.73 - +0.77m OD

Predominant species : Gramineae, Cyperaceae, *Corylus*, Chenopodiaceae.

The pollen of Gramineae, Cyperaceae and *Corylus* dominate the assemblage (figure 4.21a). Cyperaceae increase in frequency from 21% TLP to 30% TLP, whereas Gramineae decrease from 30% TLP to 16% TLP. The frequencies of *Corylus* are relatively stable at ~22% TLP. The frequencies of Chenopodiaceae are low but increase slightly upward reaching 5% TLP. Also the aquatic pollen *Potamogeton* and *Typha angustifolia* increase in frequency slightly upward.

4.3.1.3 Core AY17 : Particle Size Analysis.

The sediments at the base of core AY17, *i.e.* -2.10 - 1.905m OD, have a mean grain size of fine sand and are very well sorted (figure 4.22). These sediments exhibit a range of skewness values, from fine- to coarse-skewed, and are predominantly platykurtic.

The remainder of core AY17, *i.e.* -1.905 - +1.16m OD, is considerably different to the lowermost suite of sediments, with mean grain sizes ranging from medium to very fine silt (figure 4.22). Throughout the upper part of core AY17, the clay fraction fluctuates around a general proportion of ~20%. The sand fraction fluctuates between ~0% and ~34%, with increases in the percentage of sand being accommodated by reductions in the percentage of silt but with relatively little change in the clay fraction.

The sediments in this upper part of core AY17 are predominantly very poorly sorted (with a few poorly sorted examples), near symmetrical to fine-skewed, and predominantly platykurtic with some mesokurtic samples. Some notable exceptions to the general rule are present. At a depth of c. -1.765m OD, a sample of very fine sand is present which is made up >80% sand, ~15% clay and negligible silt. This sample is also poorly sorted, strongly fine-skewed and very leptokurtic.

Towards the top of core AY17, *i.e.* +0.99 - +1.135m OD, and also at a depth of +0.845m OD, the sediments exhibit subtly different particle size distributions. These samples are very poorly sorted, medium to fine silts and, most

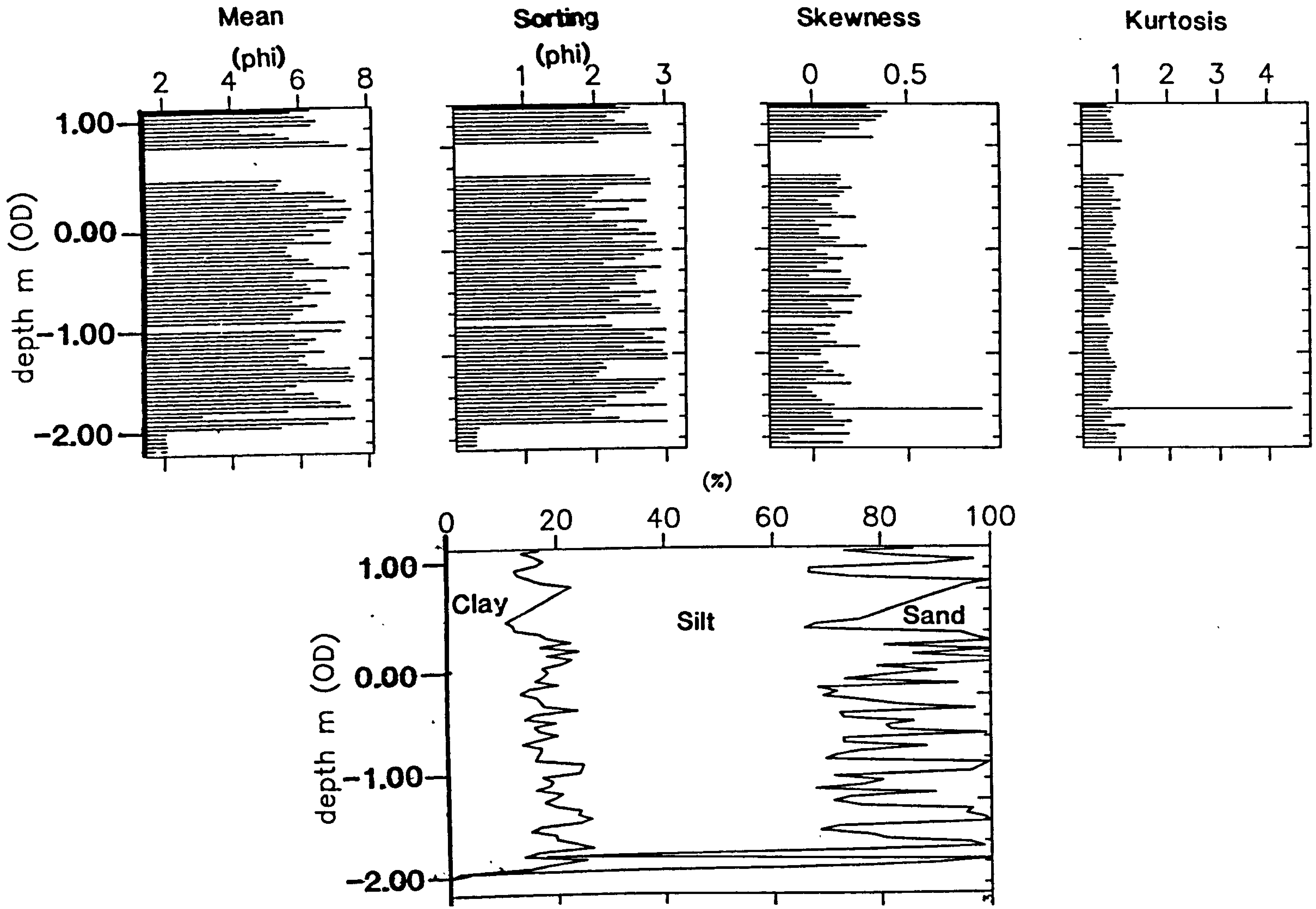


Figure 4.22: Core AY17, Particle Size Diagram.

significantly, they are strongly fine-skewed.

4.3.2 Core A-B27, Palaeoenvironmental Reconstruction : Results.

4.3.2.1 Core A-B27 : Diatom Analysis.

LDAZ A-B27 - I : +0.65 - +0.88m OD

Predominant species : *Navicula peregrina, Nitzschia navicularis, Nitzschia scalaris.*

Predominant ecological groups : *Navicula digitoradiata var. minima, Melosira sulcata.*

Brackish epiphytic/benthic diatoms dominate this zone at a high frequency (~55% TV); *Navicula peregrina* and *Nitzschia navicularis* strongly influence this (figures 4.23a&b). The taxon *Nitzschia scalaris* is high in frequency (~14% TV) and contributes significantly to the frequencies of brackish-fresh epiphytic/benthic diatoms. A relatively low frequency (<11% TV) of marine planktonic diatoms occurs throughout the zone, *i.e. Pseudopodosira westii.* The frequencies of brackish-marine epiphytic/benthic diatoms are relatively low (<5% TV) throughout the zone but increase upwards, especially *Diploneis didyma*. The opposite trend is exhibited by fresh-brackish epiphytic/benthic diatoms, especially *Epithemia zebra* which decreases upward at <5% TV. Epiphytic/benthic diatoms dominate strongly (table 4.3), with a brackish optimal diatom group accounting for >62% TV throughout, but there are also significant numbers of planktonic diatoms (7-14% TV).

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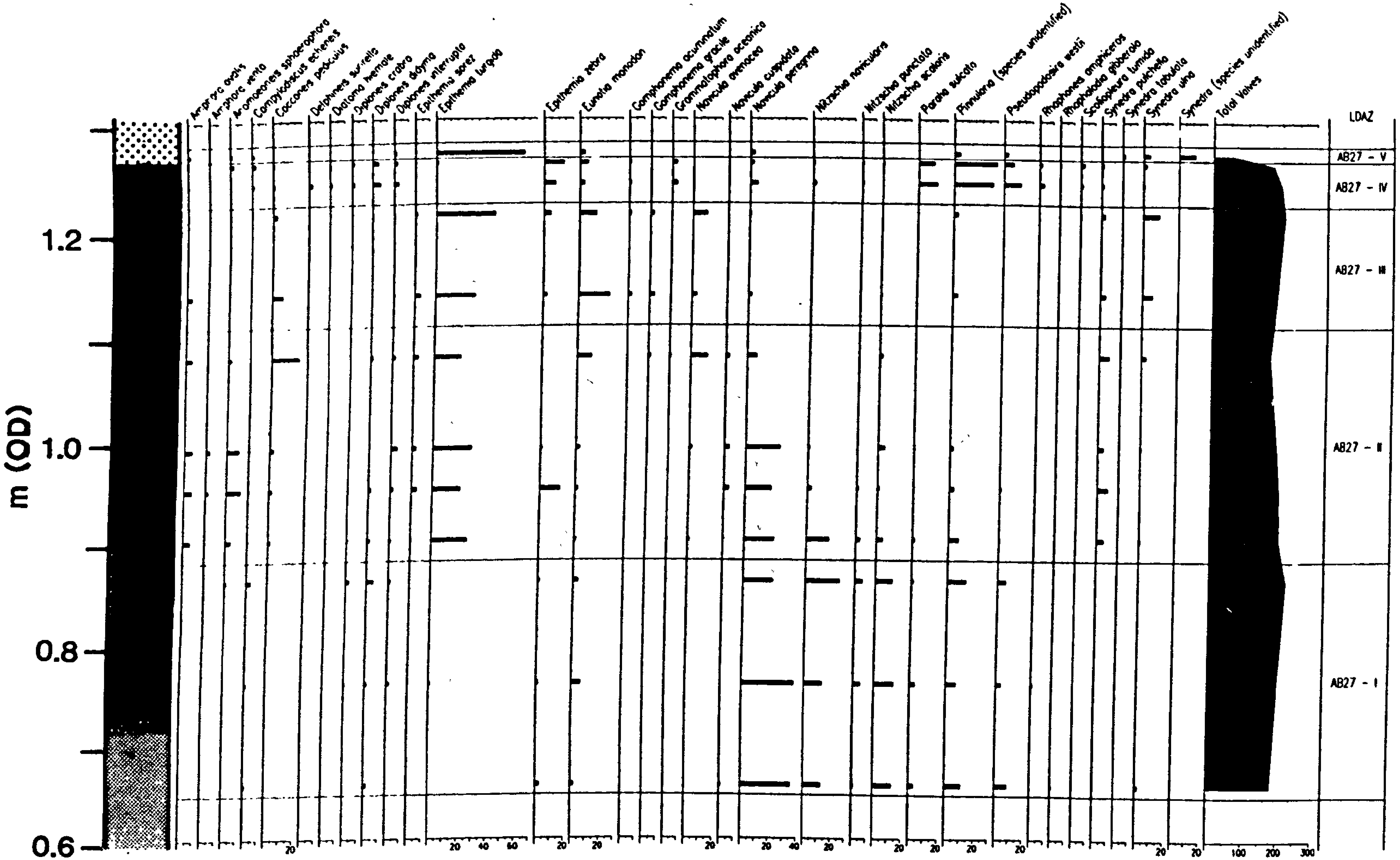

Figure 4.23a: Core A-B27, Diatom Diagram : species (legend figure 4.20b).

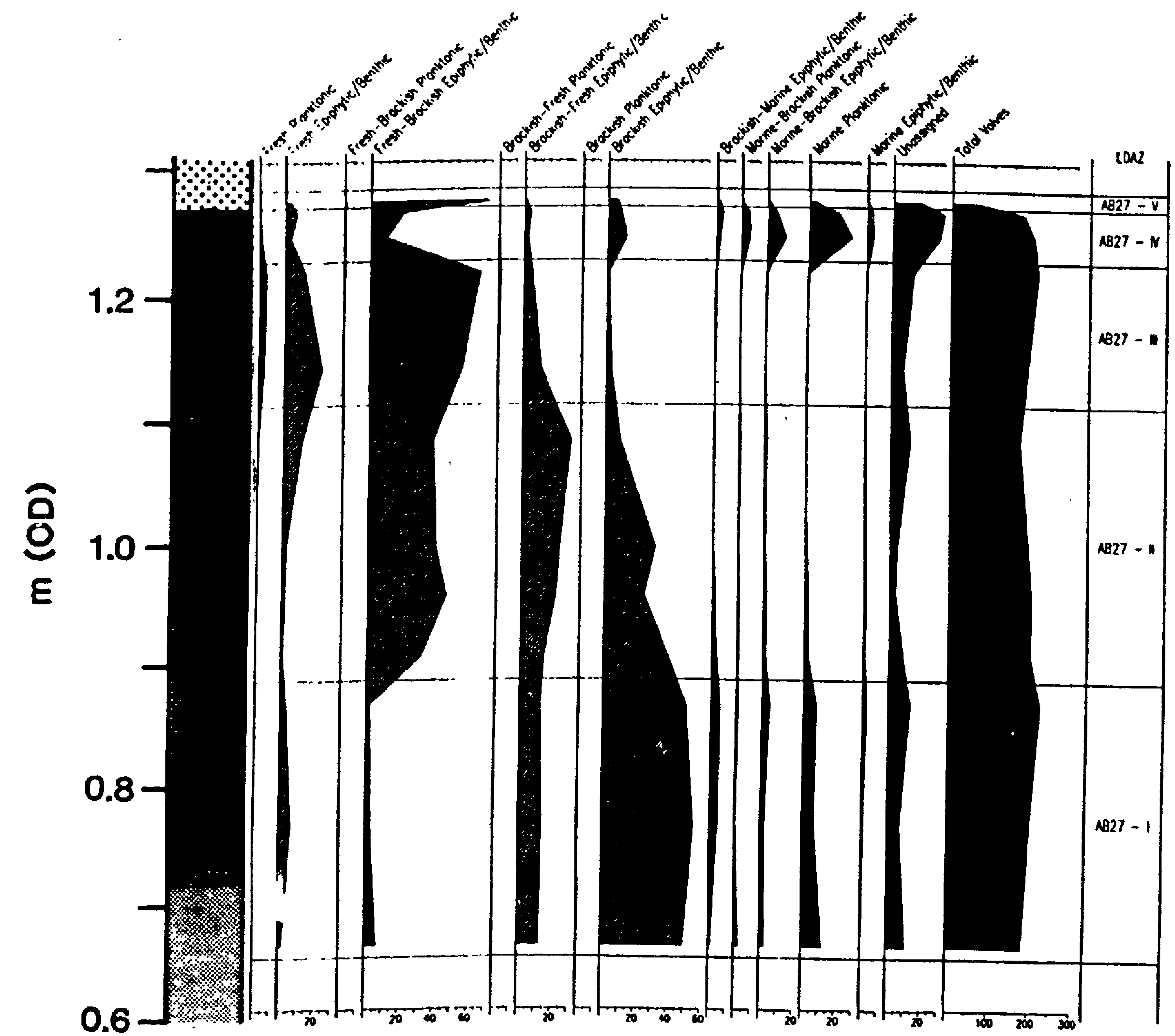

Figure 4.23b: Core A-B27, Diatom Diagram : salinity tolerance groups.

Sample Depth m (OD) LDAZ	Total diatoms % Epiphytic / Benthic	Total diatoms % Planktonic	Autochthonous diatoms (Optimal Epiphytic / Benthic group) *	Allochthonous diatoms (Fresh-water) *	Allochthonous diatoms (Marine) *
V +1.28	75.3	2.8	F-B - 68.4%	-	0.102
IV +1.27	40.2	21.9	F-B - 21.2%	-	0.896
IV +1.25	38.7	31.2	B - 15.0%	0.640	0.940
III +1.22	70.0	1.4	F-B - 69.3%	-	0.010
III +1.14	68.0	0.0	F-B - 65.1%	-	0.045
II +1.08	73.2	0.9	F-B - 63.4%	-	0.155
II +0.99	79.6	0.4	F-B - 48.4%	-	0.643
II +0.95	79.0	1.6	F-B - 52.0%	-	0.517
II +0.90	85.5	2.2	B - 51.8%	0.616	0.033
I +0.86	75.0	8.3	B - 67.0%	0.028	0.101
I +0.76	80.0	7.0	B - 73.4%	0.054	0.035
I +0.66	70.8	14.0	B - 62.9%	0.105	0.030

Table 4.3 : Diatoms, Core A-B27.

* Adapted from Beyens and Denys (1982).

LDAZ A-B27 - II : +0.88 - +1.11m OD

Predominant species : *Epithemia turgida, Navicula peregrina, Synedra pulchella.*

Predominant ecological groups : *Epithemia zebra, Navicula digitoradiata var. minima, Synedra tabulata.*

The frequencies of fresh-brackish epiphytic/benthic diatoms rise initially and dominate the zone at ~40% TV, especially *Epithemia turgida* (figures 23 a&b). Initially high frequencies (~40% TV) of brackish epiphytic/benthic diatoms fall upward to 8% TV, as illustrated by the fall of *Navicula peregrina* in particular. The opposite trend is demonstrated by brackish-fresh epiphytic/benthic diatoms, especially *Cocconeis pediculus* which increases upward, reaching 28% TV toward the top of the zone. The epiphytic/benthic diatoms dominate strongly throughout the zone (>73% TV), whereas very low numbers of planktonic diatoms are recorded (<2.5% TV). In the middle of this zone an increase in the marine allochthonous diatoms is present (table 4.3).

LDAZ A-B27 - III : +1.11 - +1.23m OD

Predominant species : *Epithemia turgida, Eunotia monodon, Synedra ulna*.

Predominant ecological group : *Epithemia zebra*.

The frequencies of fresh-brackish epiphytic/benthic diatoms especially *Epithemia turgida*, are high and increase upward from 50% to 64% TV (figure 23a&b). Fresh epiphytic/benthic diatoms are initially relatively high (21% TV) in frequency and fall upward to 12% TV, especially *Eunotia monodon*. The frequencies of brackish-fresh epiphytic/benthic and brackish epiphytic/benthic diatoms are low and fall upward. A fresh / brackish optimal diatom group accounts for 65-70% TV in this zone (table 4.3), and there are also very low values of marine allochthonous diatoms present throughout.

LDAZ A-B27 - IV : +1.23 - +1.275m OD

Predominant species : *Paralia sulcata, Pseudopodosira westii, Epithemia zebra*.

Predominant ecological groups : *Melosira sulcata, Epithemia* *zebra, Navicula digitoradiata var. minima.*

A dramatic change occurs in this zone as the frequencies of fresh-brackish epiphytic/benthic diatoms fall from dominance to 11% TV (figure 4.23b), and then rise to 18% TV at the top of the zone. This is reflected in the absence of *Epithemia turgida*. An increase in the marine influence occurs; marine planktonic diatoms initially dominating at 25% TV, especially *Paralia sulcata* and *Pseudopodosira westii*. Increases in frequencies are observed for all brackish and marine groups, which increase rapidly at the base of the zone and fall in frequency at the top. The optimal diatom group is initially brackish and becomes fresh-brackish upward, but at significantly lower percentages than the zones above or below (15-21.2% TV). A significantly higher ratio of marine allochthonous diatoms is also present (table 4.3). The percentage of epiphytic/benthic diatoms are considerably lower throughout the zone with a corresponding increase recorded in planktonic diatoms.

LDAZ A-B27 - V : +1.275 - +1.29m

Predominant species : *Epithemia turgida, Navicula peregrina, Eunotia monodon, Synedra ulna.*

Predominant ecological groups : *Epithemia zebra, Navicula digitoradiata var. minima.*

Again fresh-brackish epiphytic/benthic diatoms dominate this zone, reaching peak frequencies of 69% TV (figure 4.23b). This is strongly controlled by the very high frequencies of *Epithemia turgida*, which reach 60% TV. A frequency of 6% TV is recorded for brackish epiphytic/benthic taxa in this zone, which is mainly due to the contribution from *Navicula peregrina*. Low frequencies (<5% TV) of marine planktonic and fresh epiphytic/benthic diatoms are also present (table 4.3). The optimal diatom group, *i.e.* fresh-brackish, dominates strongly at 68.4% TV with a low ratio (0.102) of marine allochthonous diatoms.

4.3.2.2 Core A-B27 : Pollen analysis.

LPAZ A-B27 - I : +0.72 - +0.91m OD

Predominant species : Cyperaceae, Gramineae, Chenopodiaceae.

The frequencies of Gramineae (27% TLP) and Cyperaceae (20% to 43% TLP) are high and dominate the assemblage. Chenopodiaceae are present at a relatively high frequency, reaching 20% TLP, with a low but significant trace of *Plantago maritima* through the zone (figure 4.24a). Significant traces of the aquatic pollen *Typha angustifolia* and *Potamogeton* also occur.

LPAZ A-B27 - II : +0.91 - +1.07m OD

Predominant species : Cyperaceae, Gramineae, *Corylus*, *Alnus*.

The frequencies of Gramineae, at ~28% TLP throughout, and Cyperaceae, ranging from 22% to 46% TLP, are high and dominate this zone (figure 4.24a). *Corylus* and *Alnus* frequencies are relatively high; the former increasing up

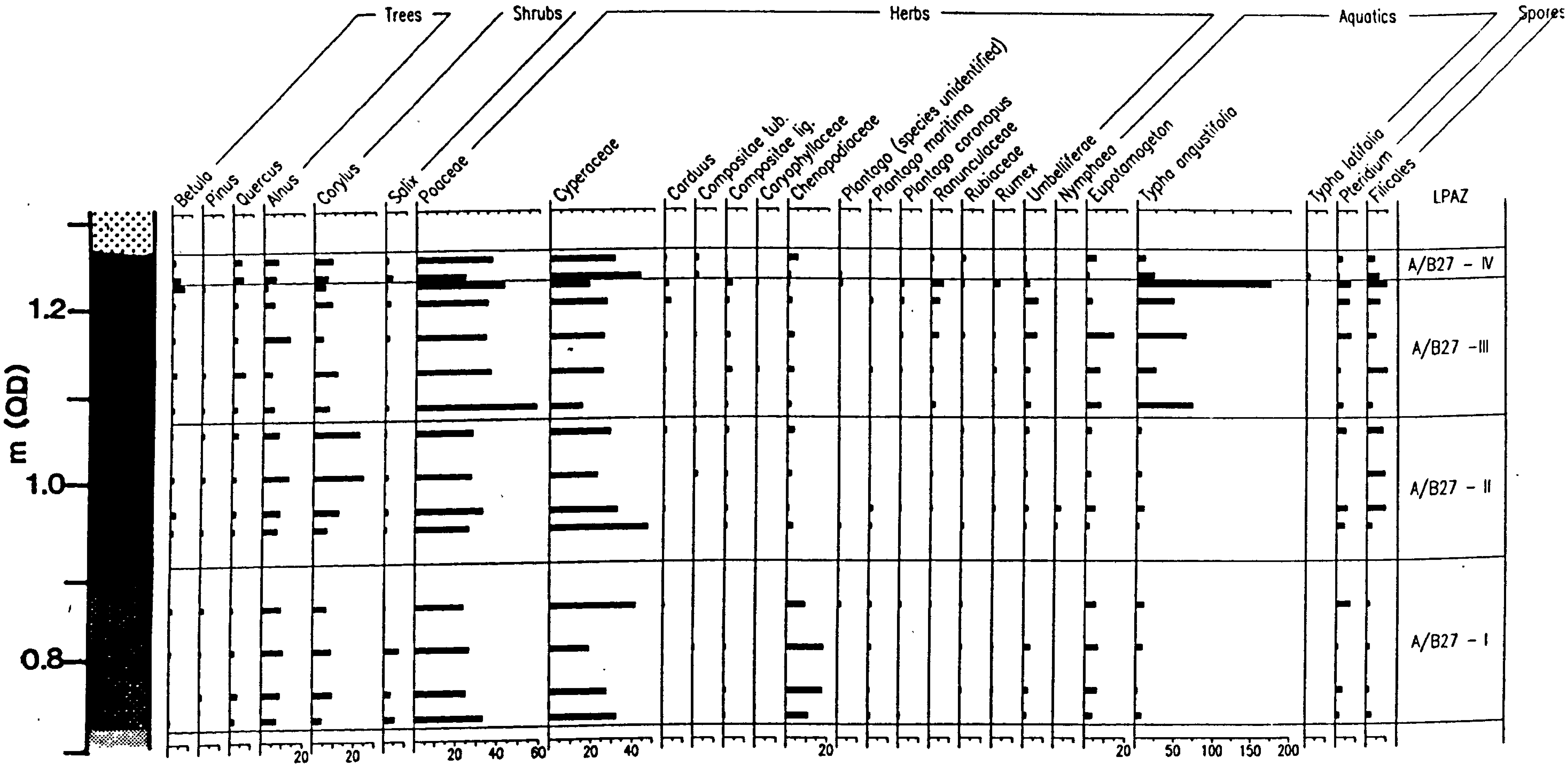

Figure 4.24a: Core A-B27, Pollen Diagram : species (legend figure 4.20b).

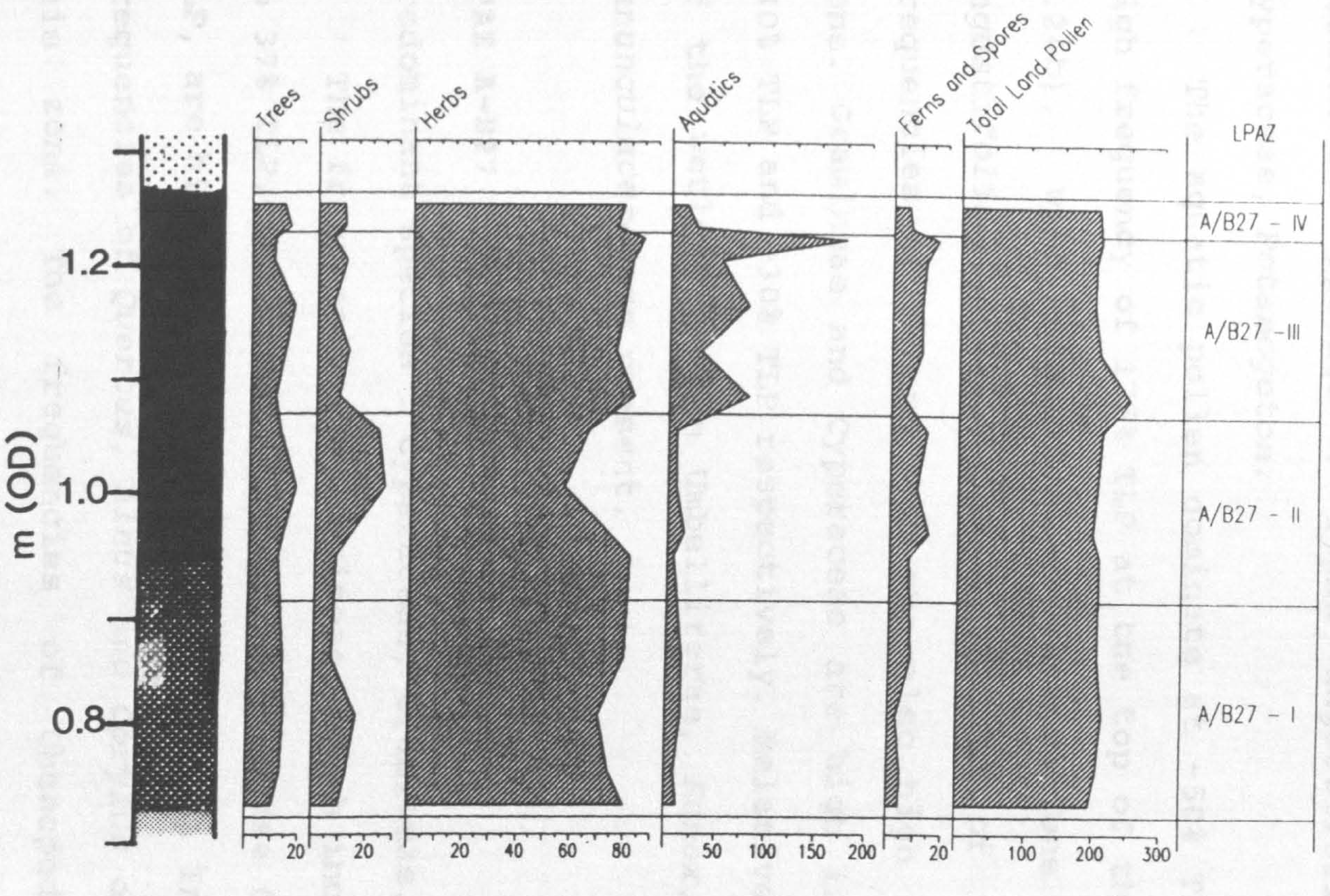

Figure 4.24b: Core A-B27, Pollen Diagram : groups (legend figure 4.20b).

to 30% TLP. Traces of aquatic pollen occur throughout the zone, and Filicales increases to 10% TLP in the upper part. Traces of the wetland pollen types Umbelliferae, *Rumex* and *Ranunculaceae* occur throughout, as do Chenopodiaceae.

LPAZ A-B27 - III : +1.07 - +1.24m OD

Predominant species : *Typha angustifolia*, Gramineae, Cyperaceae, *Potamogeton*.

The aquatic pollen dominate at ~50% TLP and reach a high frequency of 175% TLP at the top of the zone (figure 4.24b), with important contributions from *Typha angustifolia*, which reach a maximum of 170% TLP. The frequencies of *Potamogeton* are also high throughout the zone. Gramineae and Cyperaceae are high in frequency at ~40% TLP and ~30% TLP respectively. Relatively high traces of the wetland pollen Umbelliferae, *Rumex*, *Rubiacae* and *Ranunculaceae* are present.

LPAZ A-B27 - IV : +1.24 - +1.27m OD

Predominant species : Cyperaceae, Gramineae.

The frequencies of Gramineae, which increase from 24% to 37% TLP, and Cyperaceae, which decrease from 43% to 31% TLP, are high and dominate (figure 4.24a). Increases in the frequencies of *Quercus*, *Alnus* and *Corylus* occur upward in this zone. The frequencies of Chenopodiaceae pollen increase from initially low values to a high of 6% TLP at the top of the zone. Aquatic pollen is at a frequency of ~20% TLP, with *Typha angustifolia* decreasing upward as

Potamogeton increases.

4.3.3 Core G60, Palaeoenvironmental Reconstruction : Results.

4.3.3.1 Core G60 : Diatom analysis.

LDAZ G60 - I : -2.34 - -2.21m OD

Predominant species : *Pseudopodosira westii, Paralia sulcata, Diploneis crabro, Pinnularia gibba.*

Predominant ecological groups : *Melosira sulcata, Pinnularia maior.*

Most significant in this zone is the sparsity of diatoms at the base (figure 4.25a). Marine planktonic diatoms, especially *Pseudopodosira westii*, dominate the assemblage here at >70% TV. A significant (~10% TV) presence of fresh epiphytic/benthic diatoms, especially *Pinnularia gibba* and *Pinnularia viridis*, and also brackish epiphytic/benthic diatoms, especially *Navicula peregrina*, are recorded in this zone. Planktonic diatoms dominate, although to a lesser extent than at the base of the zone. The optimal diatom group is initially marine (table 4.4), then brackish and finally fresh, with significant allochthonous components in the upper part of the zone.

LDAZ G60 - II : -2.21 - -2.13m OD

Predominant species : *Paralia sulcata, Navicula peregrina, Pseudopodosira westii, Nitzschia navicularis.*

Predominant ecological groups : *Melosira sulcata, Navicula digitoradiata var. minima.*

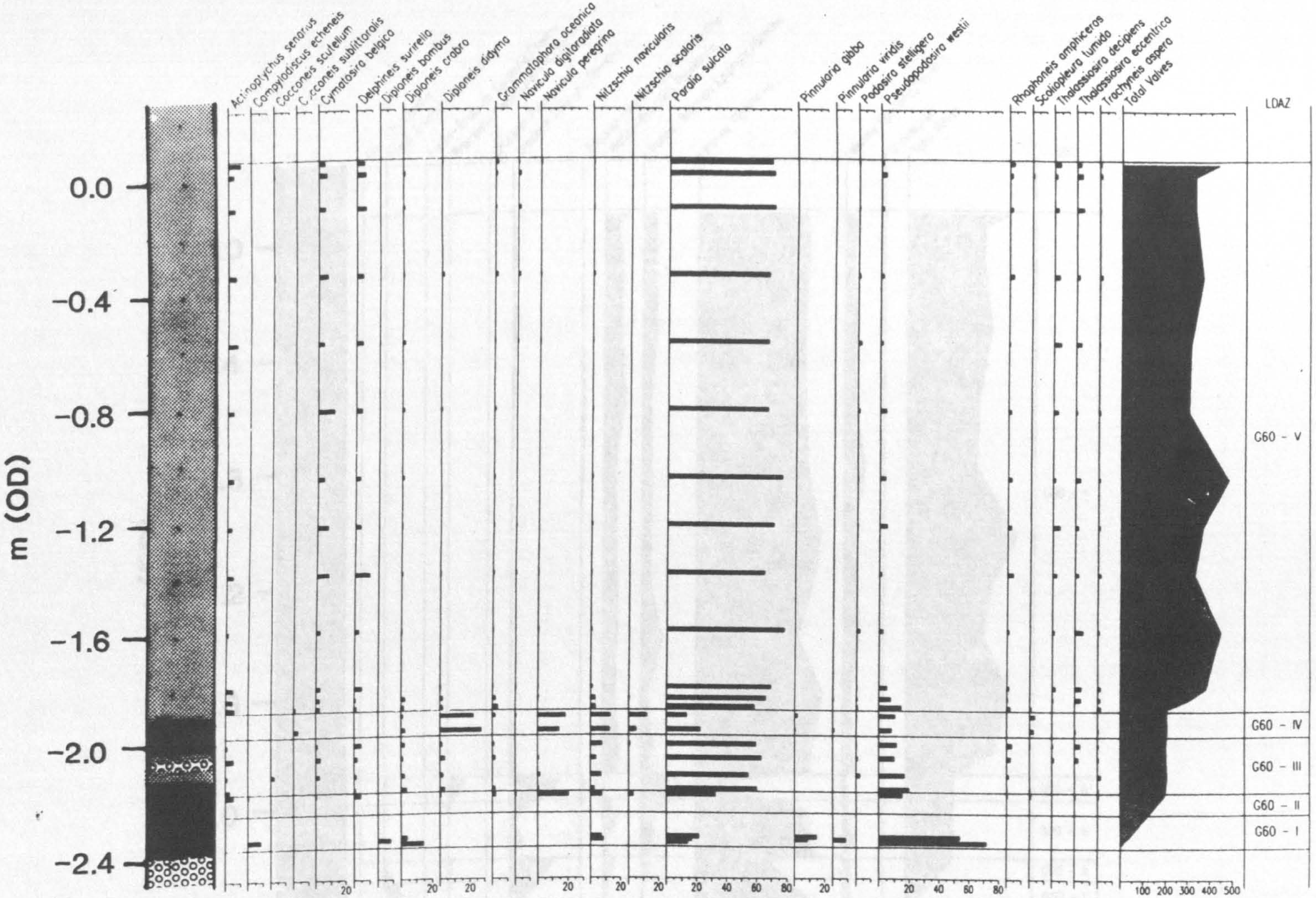

Figure 4.25a: Core G60, Diatom Diagram : species (legend figure 4.20b).

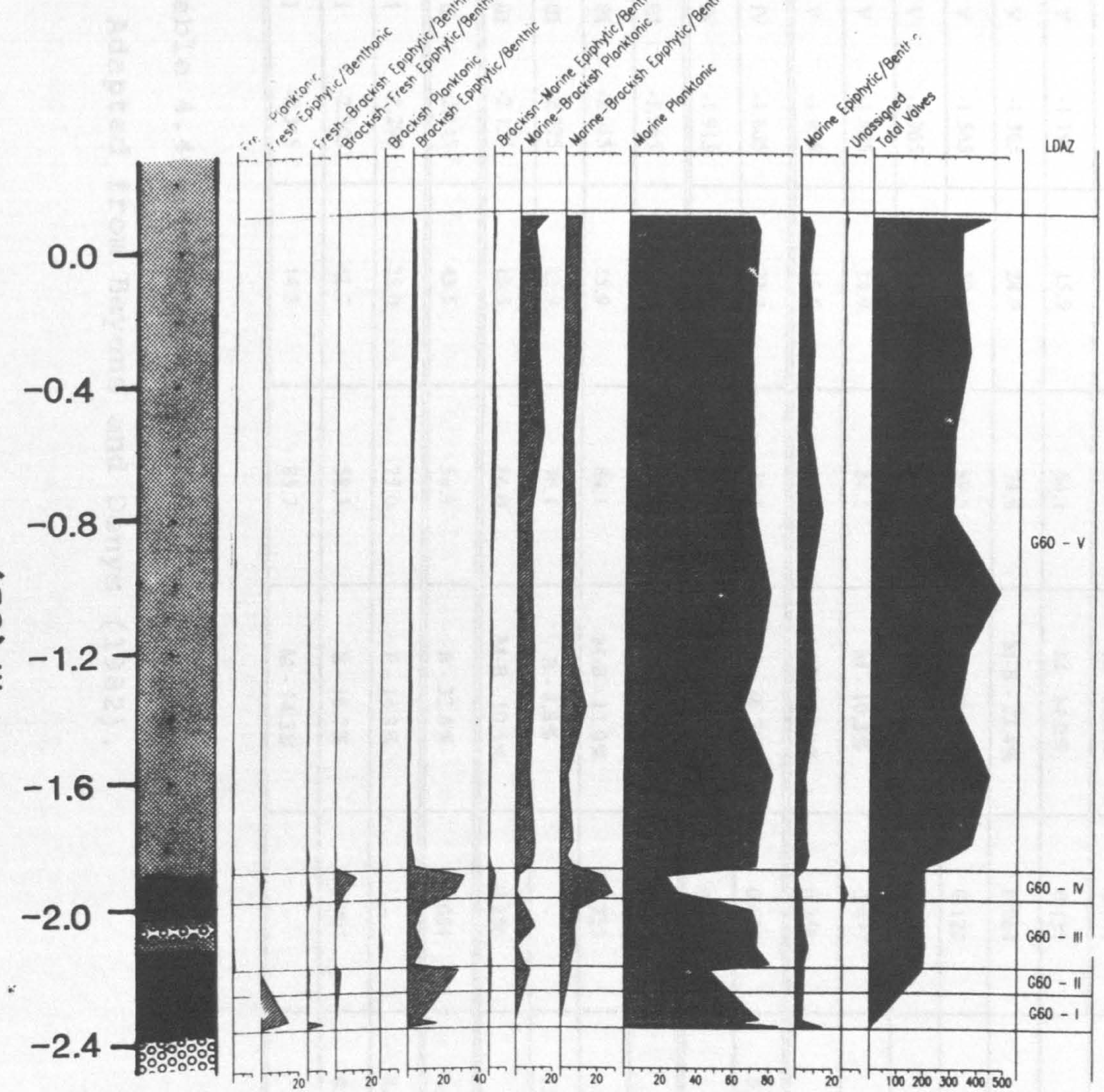

Figure 4.25b: Core G60, Diatom Diagram : salinity tolerance groups.

Sample Depth m (OD) LDAZ		Total diatoms % Epiphytic / Benthic	Total diatoms % Planktonic	Autochthonous diatoms (Optimal Epiphytic / Benthic group) *	Allochthonous diatoms (Fresh-water) *	Allochthonous diatoms (Marine) *
V	+0.115	15.0	84.5	M-B - 13.6%	0.103	-
V	+0.055	17.7	80.8	M-B - 16.1%	0.093	-
V	-0.065	15.0	82.6	M-B - 13.2%	0.136	-
V	-0.305	19.2	80.0	M-B - 16.7%	0.150	-
V	-0.545	15.0	84.4	M - 11.4%	0.579	-
V	-0.785	20.7	79.0	M - 19.0%	0.084	-
V	-1.025	12.1	87.9	M - 11.3%	0.071	-
V	-1.195	15.9	84.1	M - 14.0%	0.136	-
V	-1.365	24.9	74.8	M-B - 23.4%	0.064	-
V	-1.565	10.1	89.5	M - 9.0%	0.122	-
V	-1.765	14.5	85.0	M-B - 13.0%	0.123	-
V	-1.805	14.9	84.1	M - 10.3%	0.447	-
V	-1.835	16.6	82.9	M-B - 13.2%	0.258	-
IV	-1.865	73.4	26.1	B - 46.7%	0.054	0.518
IV	-1.915	68.0	31.0	M-B - 34.1%	0.997	-
III	-1.965	22.0	77.0	M-B - 13.5%	0.630	-
III	-2.015	15.9	84.1	M-B - 12.0%	0.333	-
III	-2.075	22.9	76.1	B - 8.8%	-	1.602
III	-2.125	13.5	86.0	M-B - 10.5%	0.286	-
II	-2.145	43.5	54.9	B - 32.6%	0.031	0.304
I	-2.295	25.0	75.0	F - 14.3%	-	0.748
I	-2.305	41.7	58.3	B - 16.7%	0.994	0.497
I	-2.315	14.3	85.7	M - 14.3%	-	-

Table 4.4 : Diatoms, Core G60.

* Adapted from Beyens and Denys (1982).

This zone has a reduced marine influence compared with the zone below. Marine planktonic diatoms are still dominant, although they fall to 48% TV, especially *Paralia sulcata* and *Pseudopodosira westii* (figures 4.25a&b).

Conversely, brackish epiphytic/benthic diatoms increase in frequency in this zone to 28% TV, especially *Navicula peregrina*. The planktonic diatoms dominate (54.9% TV) with high but subordinate epiphytic/benthic diatoms (43.5% TV). The optimal diatom group is brackish with a relatively low ratio of marine allochthonous diatoms (table 4.4).

LDAZ G60 - III : -2.13 - -1.95m OD

Predominant species : Paralia sulcata, Pseudopodosira westii.

Predominant ecological groups : *Melosira sulcata, Navicula digitoradiata var. minima.*

Marine planktonic diatoms dominate this zone at a high frequency (~75% TV) throughout; the taxa *Paralia sulcata* contributing ~60% TV and *Pseudopodosira westii* contributing between 10% and 20% TV (figures 4.25a&b). Significantly low frequencies of brackish epiphytic/benthic, marine-brackish planktonic and marine-brackish epiphytic/benthic occur, all at <10% TV. The planktonic diatoms are dominant (>76% TV). The optimal diatom group is brackish to marine-brackish throughout, although at a rather low (~10% TV) frequency (table 4.4).

LDAZ G60 - IV : -1.95 - -1.85m OD

Predominant species : Diploneis didyma, Paralia sulcata, Navicula peregrina, Nitzschia navicularis.

Predominant ecological groups : *Navicula digitoradiata var. minima, Melosira sulcata.*

Marine planktonic diatoms, especially *Paralia sulcata*, fall significantly in frequency from the high value of the previous zone to ~28% TV (figures 4.25a&b). Also marine epiphytic/benthic diatoms decrease slightly upward through the zone, particularly *Diploneis crabro*. Significant increases in brackish epiphytic/benthic species, especially *Navicula peregrina*, and marine-brackish epiphytic/benthic species, especially *Diploneis didyma*, occur into this zone; reaching ~16% TV and ~26% TV respectively. The frequencies of brackish-fresh epiphytic/benthic diatoms, especially *Nitzschia scalaris* are relatively low but rise upward through the zone, reaching 11% TV. The epiphytic/benthic diatoms dominate (68.0-73.4% TV) with subordinate planktonic diatoms (31.0-26.1% TV). The optimal diatom group is initially marine-brackish (34.1% TV) with a high ratio of fresh allochthonous diatoms (0.997) (table 4.4). At the top of the zone, the optimal diatom group becomes strongly brackish (46.7% TV).

LDAZ G60 - V : -1.85 - +0.12m OD

Predominant species : Paralia sulcata, Cymatosira belgica, Delphineis surirella.

Predominant ecological groups : *Melosira sulcata, Cymatosira belgica, Delphineis surirella*.

The dominance of marine planktonic diatoms returns in this zone, contributing between 65% and 80% TV to the assemblage throughout (figure 4.25a&b). The dominant taxon is *Paralia sulcata* at ~70% TV. Significantly low frequencies of marine-brackish planktonic species (~10% TV), especially *Actinoptychus senarius,* marine-brackish epiphytic/benthic taxa (~8% TV), especially *Delphineis surirella*, and marine epiphytic/benthic diatoms (~7% TV), especially *Cymatosira belgica*, are also present. The optimal diatom groups are marine-brackish to marine, at a relatively low level throughout (9.0-23.4% TV), with planktonic diatoms dominating at >80% TV at all but one level (table 4.4).

4.3.3.2 Core G60 : Pollen analysis.

LPAZ G60 - Ia : -2.31 - -2.30m OD

Predominant species : Gramineae, Cyperaceae, *Typha angustifolia*.

High frequencies of Gramineae (41% TLP) and Cyperaceae (36% TLP) dominate this sub-zone (figure 4.26a). Also, Chenopodiaceae are present at 7% TLP, as are the aquatic pollen *Potamogeton* and especially *Typha angustifolia* (14% TLP). Low but significant traces of *Hydrocotyle*, *Myriophylum* and Umbelliferae pollen are also present in this sub-zone.

LPAZ G60 - Ib : -2.30 - -2.29m OD

Predominant species : Chenopodiaceae, Cyperaceae.

Chenopodiaceae are the dominant taxa in this sub-zone, reaching 51% TLP, with the frequencies of Cyperaceae at 26% TLP (figure 4.26a). The frequencies of Gramineae drop considerably from LPAZ G60-Ia. Low but significant frequencies of Umbelliferae, *Myriophylum*, *Potamogeton* and *Typha angustifolia* are also present.

LPAZ G60 - II : -2.29 - -2.25m OD

Predominant species : Cyperaceae, Gramineae.

The frequencies of Cyperaceae pollen are initially high and dominate the zone at 76% TLP, but fall in frequency upward to 51% TLP (figure 4.26a). Conversely, the frequencies of Gramineae pollen increase upward from 16% to 38% TLP. Chenopodiaceae are low in frequency, although they do occur as a trace throughout, as does *Typha angustifolia*.

LPAZ G60 - III : -2.25 - -2.18m OD

Predominant species : Cyperaceae, Gramineae, *Typha angustifolia*, Umbelliferae.

The pollen of Gramineae and Cyperaceae again dominate the assemblage, with the former decreasing in frequency upward from 41% to 26% TLP, whereas the latter increases in frequency upward from 36% to 50% TLP. Aquatic pollen, especially *Potamogeton* and *Typha angustifolia*, are high at the base of the zone at 32% TLP and fall in frequency upward (figure 4.26b). The frequencies of Umbelliferae are also relatively high in this zone at ~7% TLP. Chenopodiaceae are present as a trace throughout, with a significant trace of *Plantago maritima* pollen in the middle of the zone.

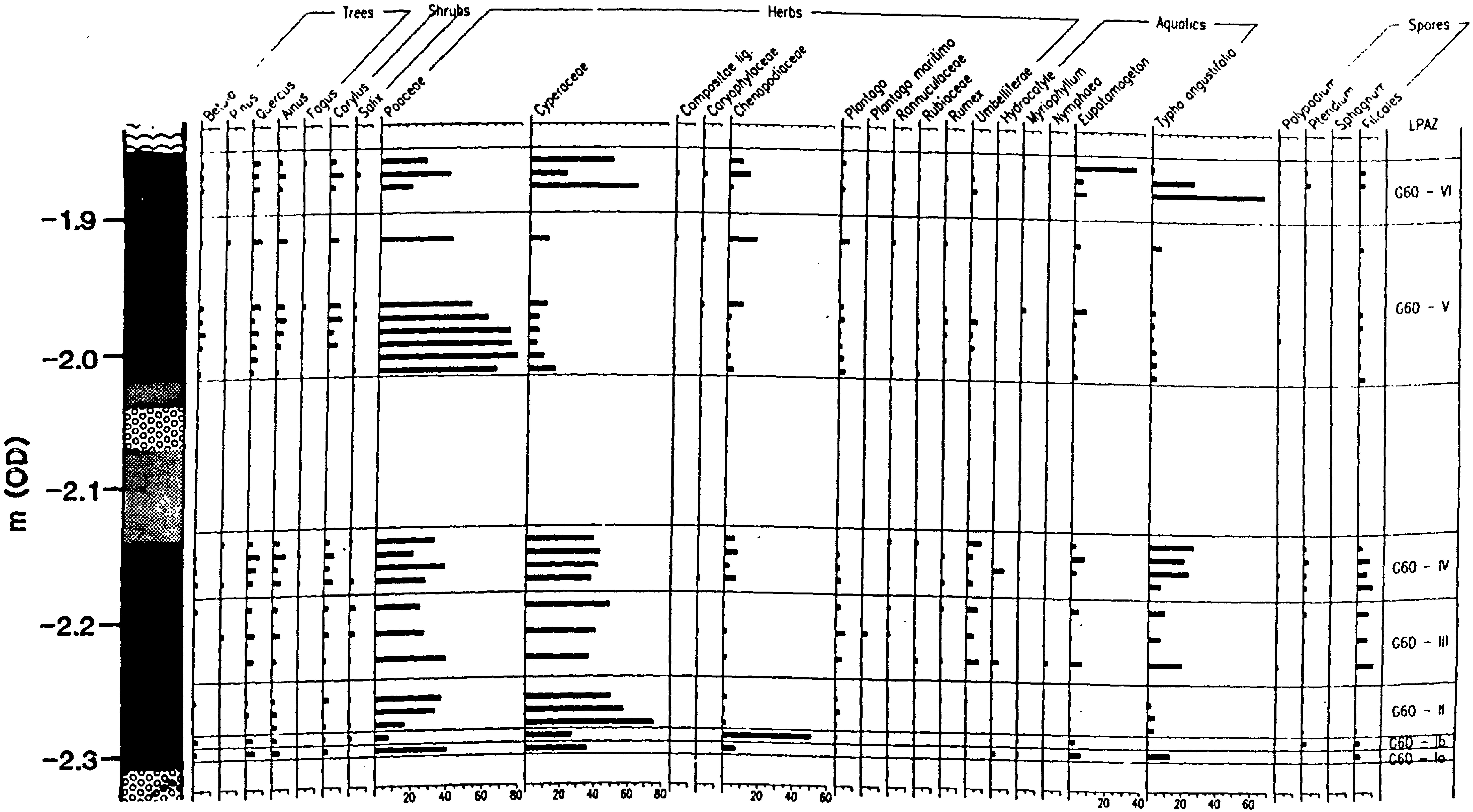

Figure 4.26a: Core G60, Pollen Diagram : species (legend figure 4.20b).

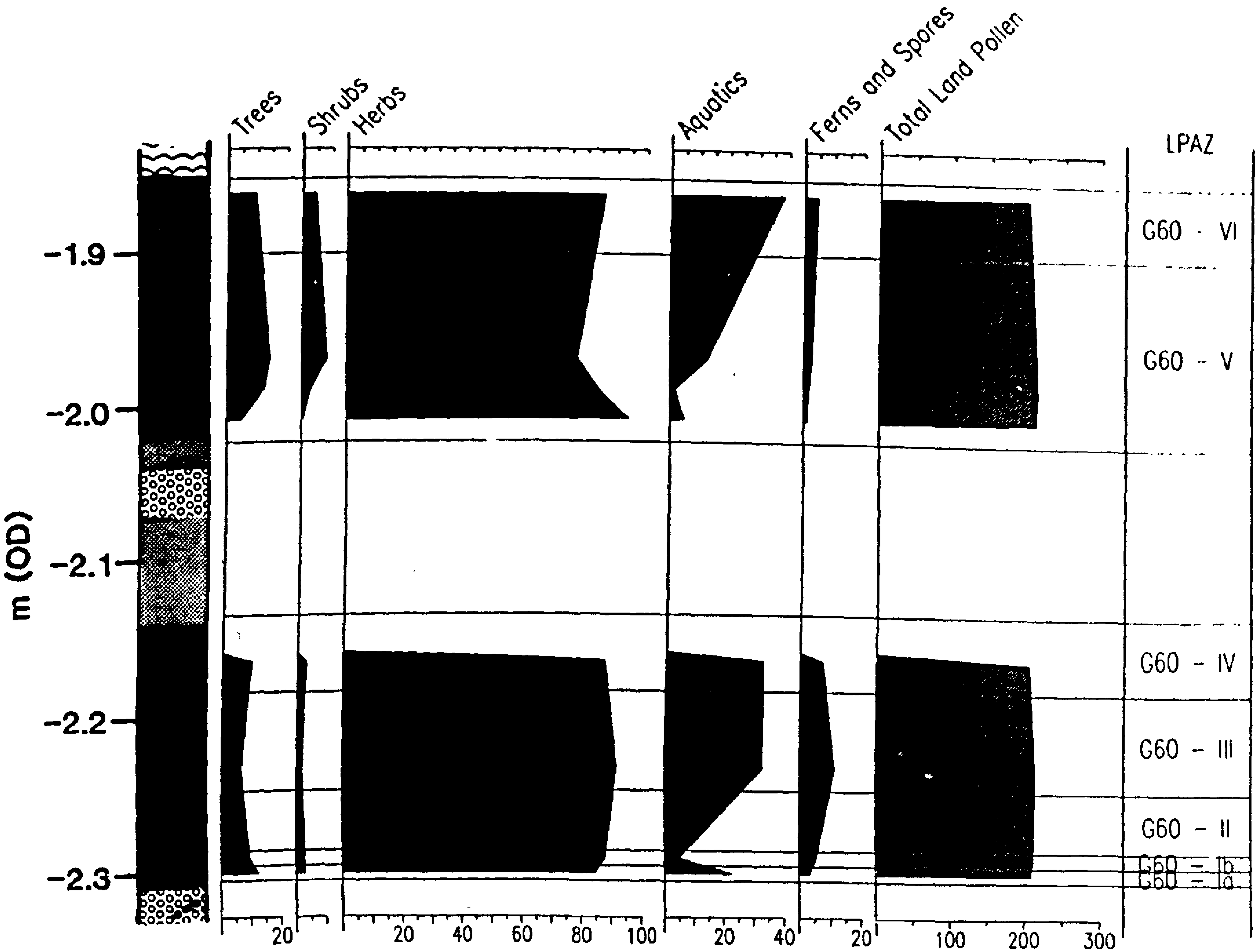

Figure 4.26b: Core G60, Pollen Diagram : groups (legend figure 4.20b).

LPAZ G60 - IV : -2.18 - -2.14m OD

Predominant species : Gramineae, Cyperaceae, *Typha angustifolia*, Chenopodiaceae.

In this zone Cyperaceae pollen are the dominant type at a frequency of ~40% TLP (figure 4.26a), whereas the frequencies of Gramineae pollen fluctuate from 21% to 39% TLP. Initially low frequencies of *Typha angustifolia* (6% TLP) rise rapidly to ~23% TLP in the upper part of the zone. The frequencies of Chenopodiaceae pollen are relatively high at ~9% TLP. Also, relatively high frequencies (<10% TLP) of *Quercus*, *Alnus* and *Corylus* pollen are recorded.

LPAZ G60 - V : -2.02 - -1.90m OD

Predominant species : Gramineae, Cyperaceae, Chenopodiaceae.

High frequencies of Gramineae (79% TLP) generally fall upward through the zone to 42% TLP, but they are the dominant pollen type (figure 4.26a). Subordinate frequencies (~8% TLP) are recorded for Cyperaceae. Initially low frequencies of Chenopodiaceae pollen gradually increase upward from ~2% to 18% TLP. The frequencies of *Quercus*, *Alnus* and *Corylus* generally increase upward to relatively high frequencies of ~8% TLP.

LPAZ G60 - VI : -1.90 - -1.85m OD

Predominant species : Cyperaceae, *Typha angustifolia*, *Potamogeton*, Chenopodiaceae, Gramineae.

Very high frequencies of *Typha angustifolia* dominate the base of the zone (66% TLP) but fall upward to <5% TLP (figure 4.26a). In contrast, *Potamogeton* pollen are initially low in frequency (7% TLP) and increase upward to 36% TLP at the top of the zone. The frequencies of Cyperaceae pollen are also high in this zone, fluctuating from 20% to 62% TLP, and Gramineae also fluctuates but at the lower frequencies of between 18% and 43% TLP. The frequencies of Chenopodiaceae pollen increase from 2% to ~11% TLP at the top of the zone.

4.3.3.3 Core G60 : Particle Size Analysis.

Throughout core G60 the sediments are very poorly sorted and tend to have a mean grain size of fine to very fine silt (figure 4.27), although some coarser samples are present. The clay fraction of most of the samples is ~20%, with a sand fraction generally <15%. Silt, therefore, forms the predominant particle size component of the samples. The sediments are also predominantly fine-skewed, although a number of near symmetrical and very fine skewed samples are present between the depths of -1.83 and -0.73m OD. Also, the sediment distributions are predominantly platykurtic with a number of mesokurtic samples at various levels.

Some very gradual changes in the particle size characteristics of core G60 can be identified in figure 4.27. For example, the mean grain size increases slightly upward from predominantly very fine silt between -1.97m OD and -0.96m OD to predominantly fine silt from -0.91m OD to

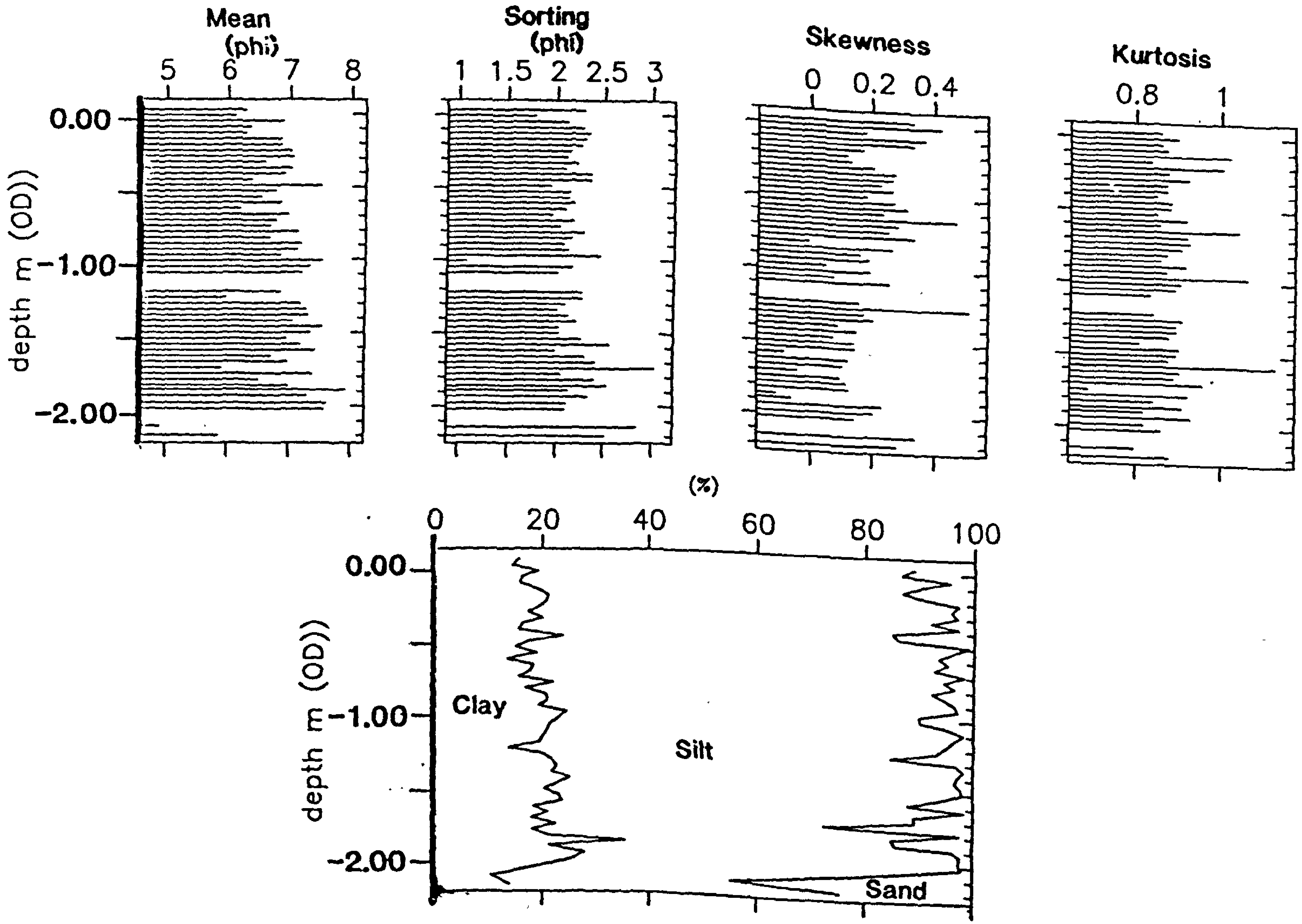

Figure 4.27: Core G60, Particle Size Diagram.

+0.10m OD. Equally, a subtle change in the skewness of the sediments occurs upwards from predominantly near symmetrical distributions in the lower part of core G60, *i.e.* -2.135 to -1.155m OD, to predominantly fine-skewed and occasionally fine-skewed samples in the upper part, *i.e.* -1.025 - +0.095m OD.

Significantly, two samples in core G60 contain gravel. These are not included in figure 4.27. Sample 53 at a depth of -2.135m OD contains one pebble of -4 phi, and sample 51 at a depth of -2.03m OD comprises pebbles of -4 phi exclusively.

4.3.4 Core AW63, Palaeoenvironmental Reconstruction : Results

4.3.4.1 Core AW63 : Diatom analysis.

LDAZ AW63 - I : -0.13 - +0.04m OD

Predominant species : *Diploneis didyma, Nitzschia navicularis, Paralia sulcata, Rhophalodia gibberula.*

Predominant ecological groups : *Navicula digitoradiata var. minima, Melosira sulcata, Synedra tabulata.*

Initially, marine-brackish epiphytic/benthic diatoms dominate, reaching 58% TV (figure 4.28b). These fall through the zone before increasing slightly at the top. Variation in the frequency *Diploneis didyma* is the primary reason of this observed trend. The frequencies of brackish epiphytic/benthic diatoms increase upward from 24% to 44% TV, and dominate at the top of the zone. This is especially true of *Rhophalodia gibberula* which reaches a peak of 26%

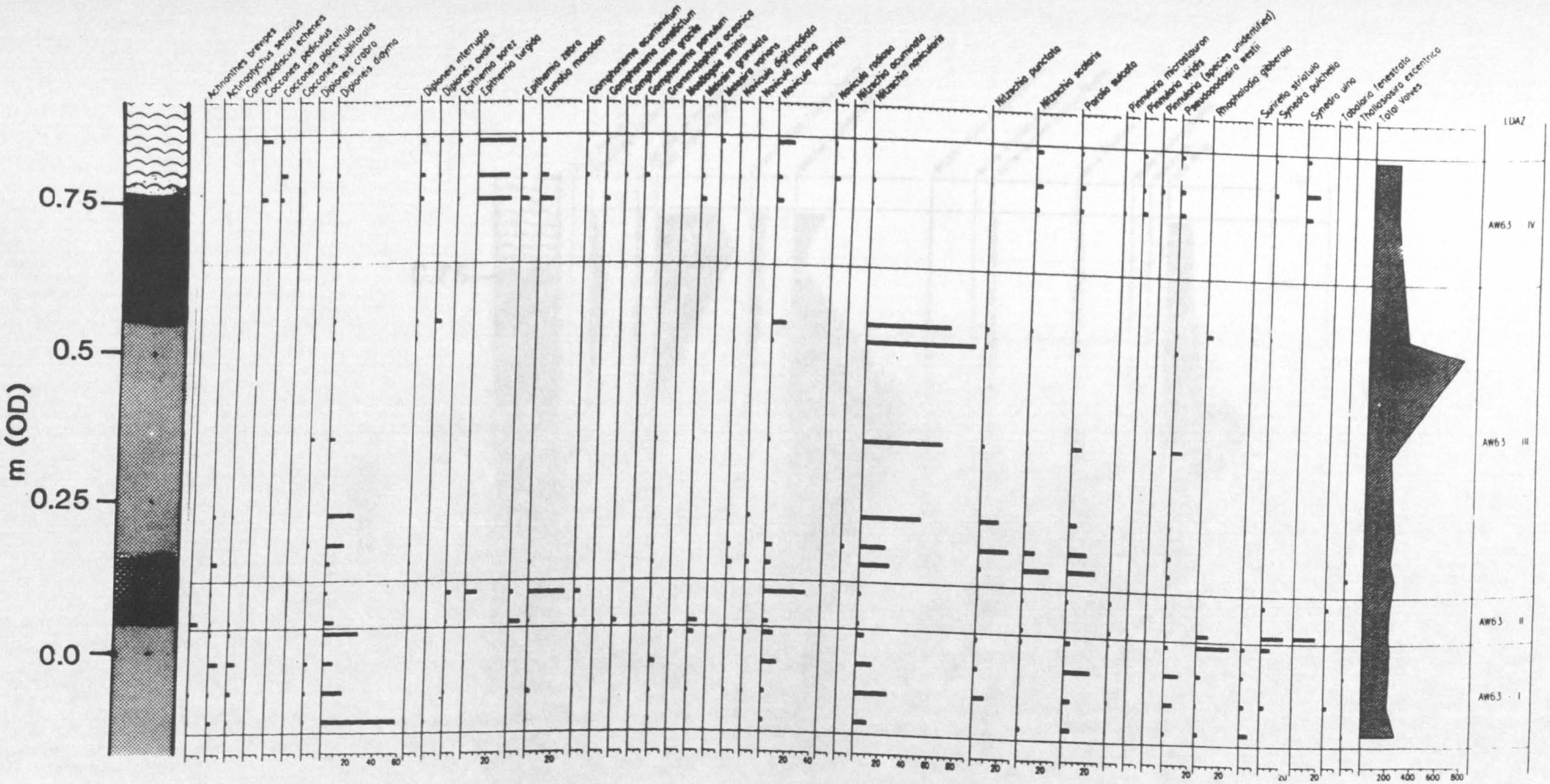

Figure 4.28a: Core AW63, Diatom Diagram : species (legend figure 4.20b).

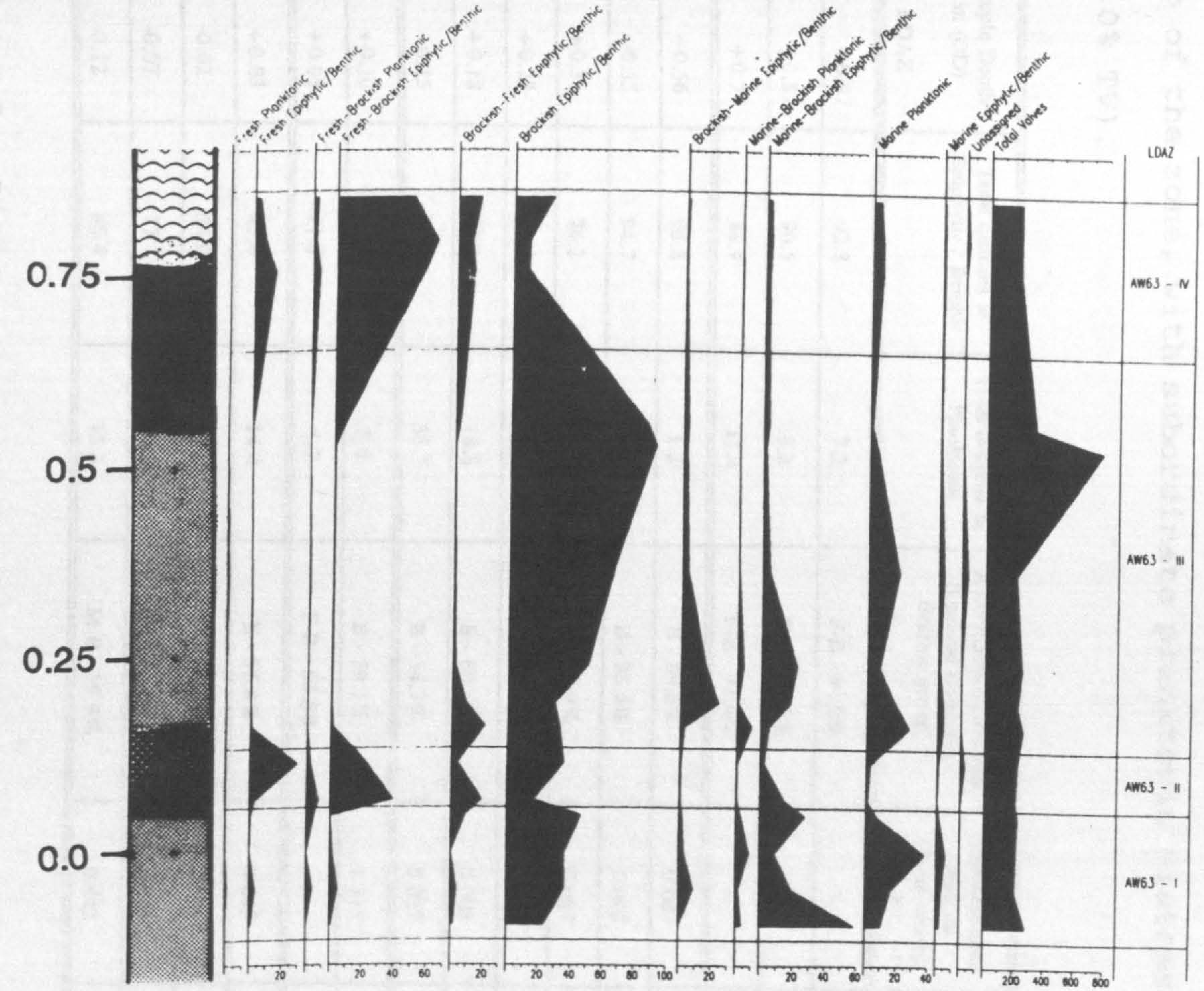

Figure 4.28b: Core AW63, Diatom Diagram : salinity tolerance groups.

TV. A strong presence of marine planktonic diatoms is evident, which reaches a peak of 36% TV due to a significant increase in the frequencies of *Paralia sulcata* and *Pseudopodosira westii*. The epiphytic/benthic diatoms dominate throughout, although at a lower level toward the top of the zone, with subordinate planktonic diatoms (13.6-46.0% TV).

Sample Depth m (OD) LDAZ		Total diatoms % Epiphytic / Benthic	Total diatoms % Planktonic	Autochthonous diatoms (Optimal Epiphytic / Benthic group) *	Allochthonous diatoms (Fresh-water) *	Allochthonous diatoms (Marine) *
IV	+0.87	92.3	7.2	F-B - 64.2%	-	0.438
IV	+0.81	90.5	8.2	F-B - 76.7%	-	0.181
IV	+0.77	88.9	11.1	F-B - 72.4%	-	0.173
III	+0.56	98.8	1.2	B - 89.6%	0.069	0.035
III	+0.53	94.7	5.3	B - 95.6%	0.005	0.018
III	+0.36	76.2	20.8	B - 68.9%	0.007	0.102
III	+0.23	88.8	11.2	B - 66.2%	0.012	0.331
III	+0.18	81.2	18.8	B - 60.6%	0.008	0.333
III	+0.15	64.3	35.7	B - 54.7%	0.037	0.139
II	+0.10	92.7	5.0	B - 39.1%	1.315	0.059
II	+0.05	91.3	8.7	F-B - 60.6%	-	0.507
I	+0.03	85.6	14.4	B - 56.8%	0.009	0.500
I	-0.02	54.0	46.0	B - 38.6%	0.036	0.363
I	-0.07	77.7	22.3	B - 48.0%	0.029	0.448
I	-0.12	86.4	13.6	M-B - 59.9%	0.442	-

Table 4.5 : Diatoms Core AW63.

* Adapted from Beyens and Denys (1982).

LDAZ AW63 - II : +0.04 - +0.13m OD

Predominant species : *Eunotia monodon, Navicula peregrina, Synedra pulchella, Synedra ulna.*

Predominant ecological groups : *Navicula digitoradiata var. minima, Synedra tabulata, Epithemia zebra.*

Low frequencies of brackish epiphytic/benthic diatoms, especially *Navicula peregrina*, increase significantly from 16% to 35% TV to dominate the upper part of the zone (figures 4.28a&b). Fresh-brackish epiphytic/benthic diatoms are high in frequency at 39% TV at the base of the zone, but fall upward through the zone to 21% TV due to a fall in the frequencies of *Synedra ulna*. The frequencies of fresh epiphytic/benthic diatoms, especially *Eunotia monodon*, rise from effectively zero at the base of the zone up to 28% TV at the top. Epiphytic/benthic diatoms dominate strongly at >90% TV throughout the zone. A very high ratio of fresh allochthonous diatoms is present at the top of the zone, coinciding with a relatively low value for the brackish optimal diatom group (table 4.5).

LDAZ AW63 - III : +0.13 - +0.66m OD

Predominant species : *Nitzschia navicularis, Paralia sulcata, Diploneis didyma, Nitzschia punctata, Nitzschia scalaris.*

Predominant ecological groups : *Navicula digitoradiata var. minima, Melosira sulcata.*

The frequencies of brackish epiphytic/benthic diatoms are initially low (31% TV) (figure 4.28b). However, this group rises through the zone to dominate the assemblage at 88% TV. The frequencies of *Nitzschia navicularis* are a clear example of this trend. Significant but relatively low frequencies of brackish-marine epiphytic/benthic (*e.g.* *Nitzschia punctata*) marine-brackish epiphytic/benthic (*e.g.* *Diploneis didyma*), marine planktonic (*e.g.* *Paralia sulcata*) and marine epiphytic/benthic diatoms (*e.g.* *Nitzschia scalaris*) occur in the lower part of this zone and decrease upward. A general increase in epiphytic/benthic diatoms (64.3-98.8% TV) is recorded upward through the zone, clearly dominating throughout. The opposite trend is recorded for planktonic diatoms. The optimal diatom group is brackish, ranging from 64.7 to 95.6% TV throughout the zone (table 4.5).

LDAZ AW63 - IV : +0.66 - +0.88m OD

Predominant species : *Epithemia turgida, Eunotia monodon, Synedra ulna, Navicula peregrina.*

Predominant ecological groups : *Epithemia zebra, Navicula digitoradiata var. minima.*

Fresh-brackish epiphytic/benthic diatoms especially *Epithemia turgida* and *Synedra ulna*, dominate the assemblage (figures 4.28a&b), reaching frequencies of 60% TV. Brackish epiphytic/benthic diatoms, especially *Navicula peregrina*, are initially low in frequency at 10% TV, and increase to 24% TV at the top of the zone. The frequencies of fresh epiphytic/benthic diatoms (*e.g.* *Eunotia monodon*) are relatively low (11% TV) and fall upward. Also low but significant frequencies of marine planktonic, fresh-brackish planktonic and brackish-fresh epiphytic/benthic diatoms are present. Epiphytic/benthic diatoms dominate strongly (~90% TV), with the optimal fresh-brackish group ranging from 64.2-76.7% TV (table 4.5). The ratio of marine allochthonous diatoms increases upward.

4.3.4.2 Core AW63 : Pollen analysis.

LPAZ AW63 - I : +0.04 - +0.08m OD

Predominant species : Gramineae.

Gramineae pollen dominate at a high frequency (75% TLP), and low but significant frequencies (<10% TLP) for *Alnus*, *Corylus* and Cyperaceae occur (figure 4.29a). The frequencies of the aquatic pollen are very low. Low but significant traces of Chenopodiaceae, *Plantago maritima*, *Plantago coronopus*, *Ranunculaceae* and *Rumex* also occur.

LPAZ AW63 - II : +0.08 - +0.18m OD

Predominant species : Gramineae, Cyperaceae, *Typha angustifolia*.

The frequencies of Gramineae pollen are dominant, although they decrease to ~45% TLP from the relatively high values of the previous zone. Cyperaceae are subordinate at a frequency of ~15% TLP. The aquatic pollen, especially *Typha angustifolia* (17% TLP), increase significantly in frequency in this zone, ranging from 10% to 18% TLP, figure 4.29b. The frequencies of Chenopodiaceae pollen increase upward to 11% TLP. Significant traces of the wetland herb pollen types, *i.e.* *Ranunculaceae*, *Rubiacae* and *Rumex*, occur. The tree pollen generally increases upwards, whilst the frequencies of shrub pollen, in particular *Corylus* (which reaches 13% TLP), tend to increase slightly to the middle of the zone and then fall at the top.

LPAZ AW63 - III : +0.56 - +0.61m OD

Predominant species : Gramineae, Cyperaceae, Chenopodiaceae.

The frequencies of Gramineae pollen increase from 25% TLP to 52% TLP (figure 4.29a), whereas Cyperaceae pollen fall in frequency from 46% TLP to 21% TLP upward. These herb types dominate the assemblage. Chenopodiaceae pollen are initially high in frequency at 13% TLP, but this decreases upward. The frequencies of aquatic pollen types in this zone are low.

LPAZ AW63 - IV : +0.61 - +0.71m OD

Predominant species : Gramineae, *Typha angustifolia*, Cyperaceae.

The frequencies of Gramineae pollen are high and dominate at 80% TLP, but fall to 62% TLP at the top of the zone (figure 4.29a). High frequencies of *Typha angustifolia* occur in this zone, rising upward from 26% to 50% TLP. Frequencies of tree and shrub pollen are relatively low. Chenopodiaceae pollen occur as a trace, whereas *Plantago maritima* increases in frequency to a relatively low peak in this zone.

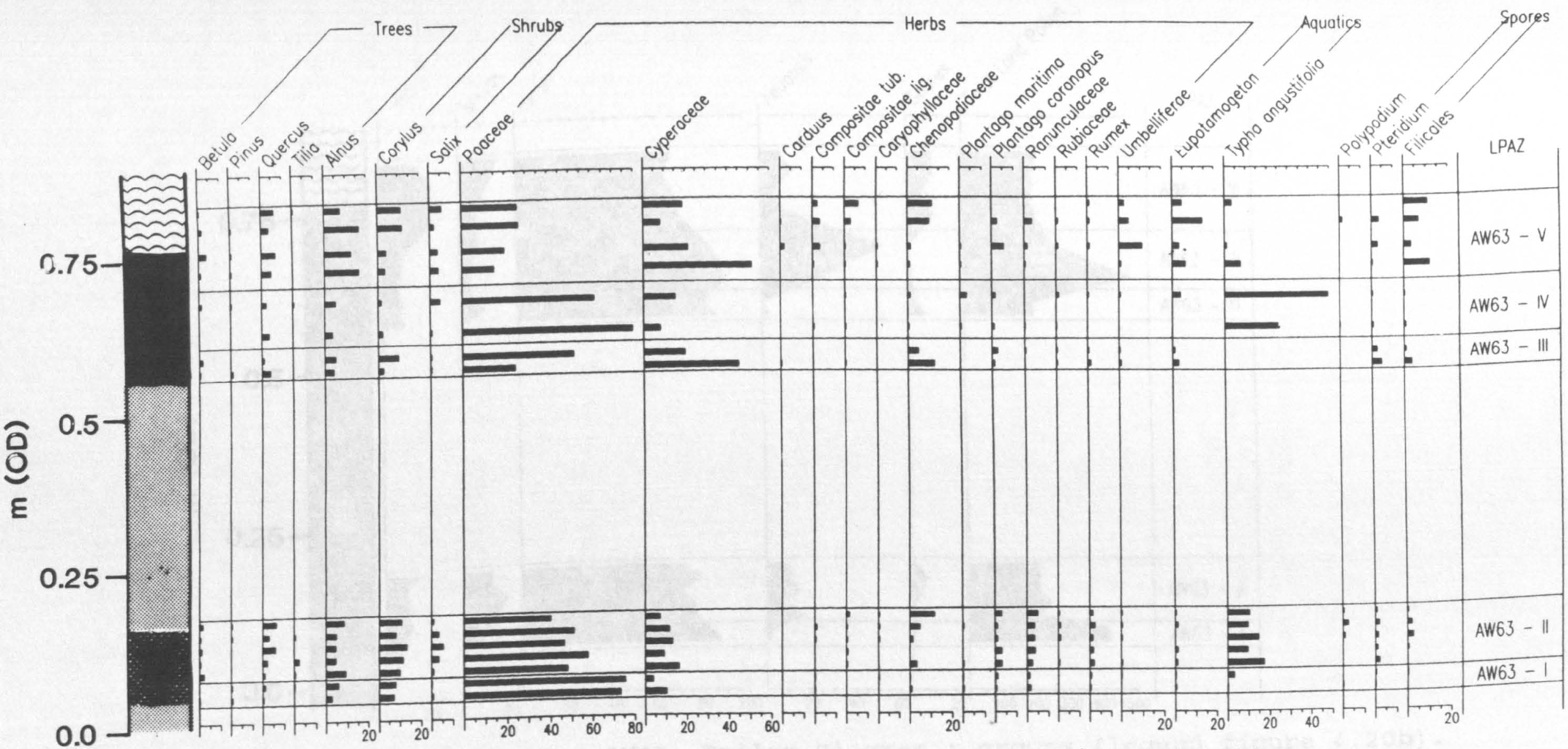

Figure 4.29a: Core AW63, Pollen Diagram : species (legend figure 4.20b).

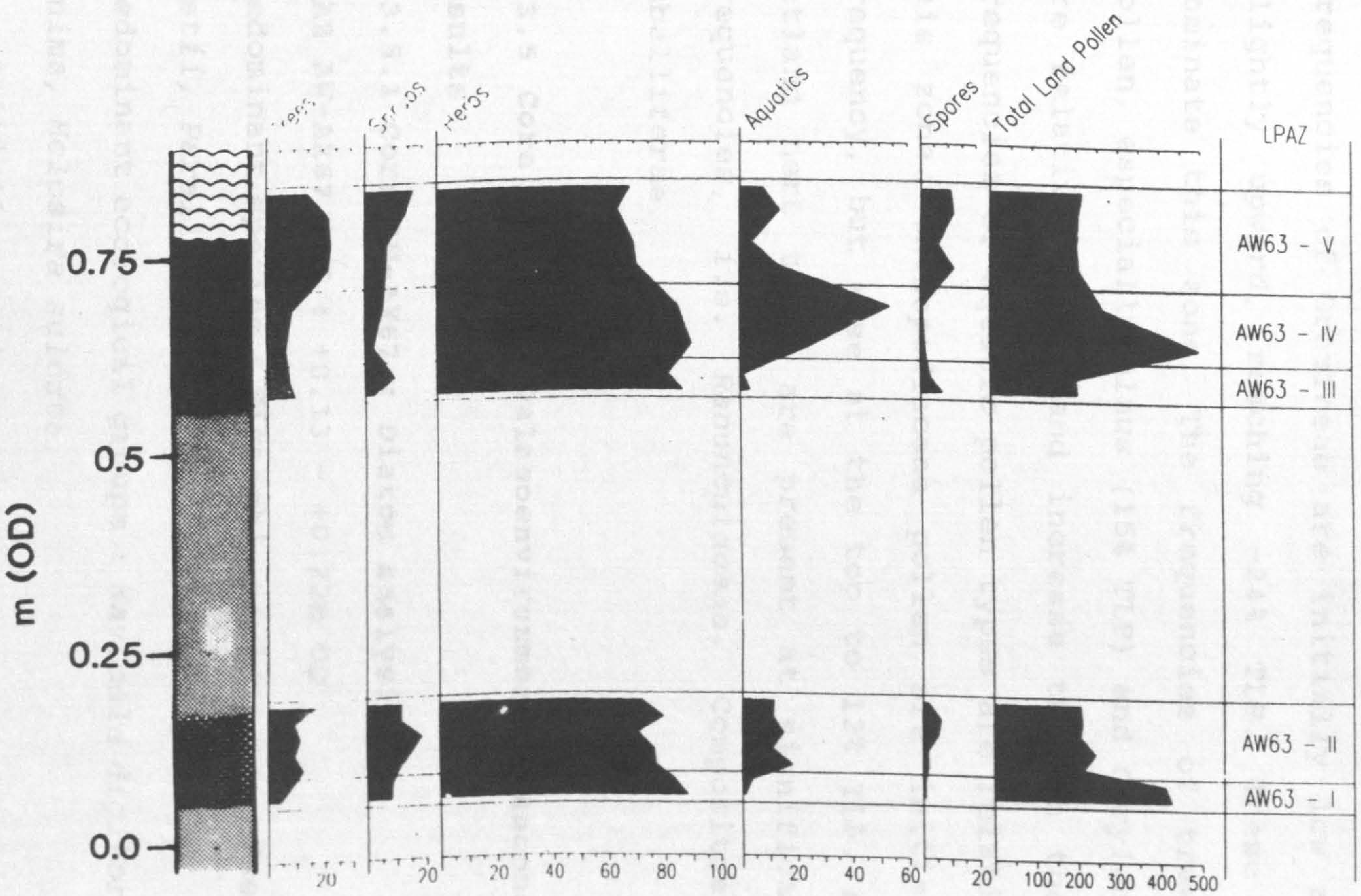

Figure 4.29b: Core AW63, Pollen Diagram : groups (legend figure 4.20b).

LPAZ AW63 - V : +0.71 - +0.84m OD

Predominant species : Cyperaceae, Gramineae, Chenopodiaceae.

Initially, Cyperaceae pollen are high in frequency and dominate at 52% TLP (figure 4.29a). However, Cyperaceae fall to ~15% TLP in the upper part of the zone. The frequencies of Gramineae are initially low and increase slightly upward, reaching ~24% TLP. These herb types dominate this zone. The frequencies of tree and shrub pollen, especially *Alnus* (15% TLP) and *Corylus* (13% TLP), are relatively high and increase through the zone. The frequencies of aquatic pollen types are relatively low in this zone. Chenopodiaceae pollen are initially low in frequency, but rise at the top to 12% TLP. A number of wetland herb types are present at significant but low frequencies, *i.e. Ranunculaceae,* Compositae tub. and Umbelliferae.

4.3.5 Core AW-AX67, Palaeoenvironmental Reconstruction : Results

4.3.5.1 Core AW-AX67 : Diatom analysis.

LDAZ AW-AX67 - I : +0.13 - +0.22m OD

Predominant species : *Nitzschia navicularis, Pseudopodosira westii, Paralia sulcata, Navicula peregrina.*

Predominant ecological groups : *Navicula digitoradiata* var. *minima, Melosira sulcata.*

Initially, marine planktonic diatoms are dominant at 36% TV (figure 4.30b). However, this is reduced to 18% TV,

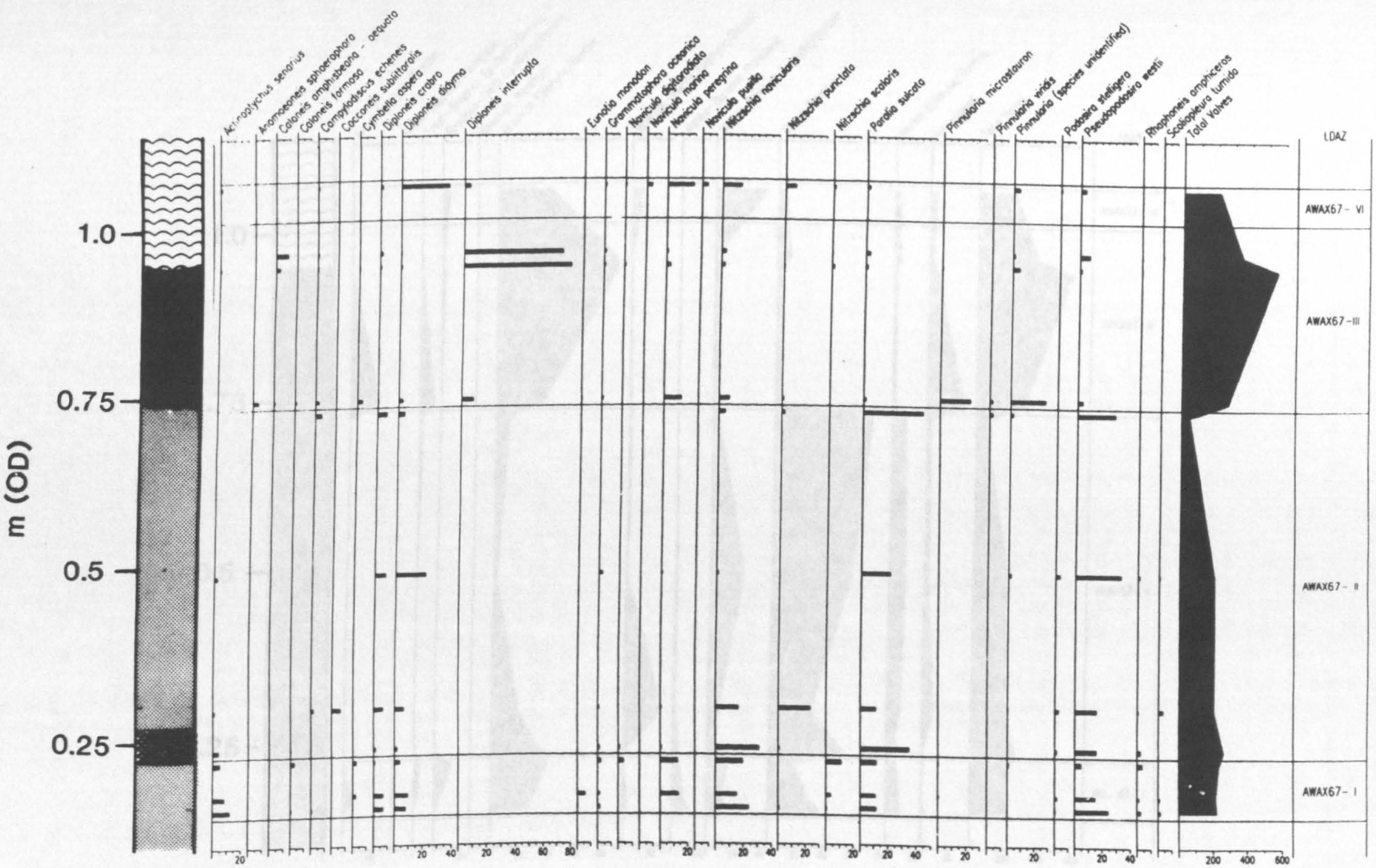

Figure 4.30a: Core AW-AX67, Diatom Diagram : species (legend figure 4.20b).

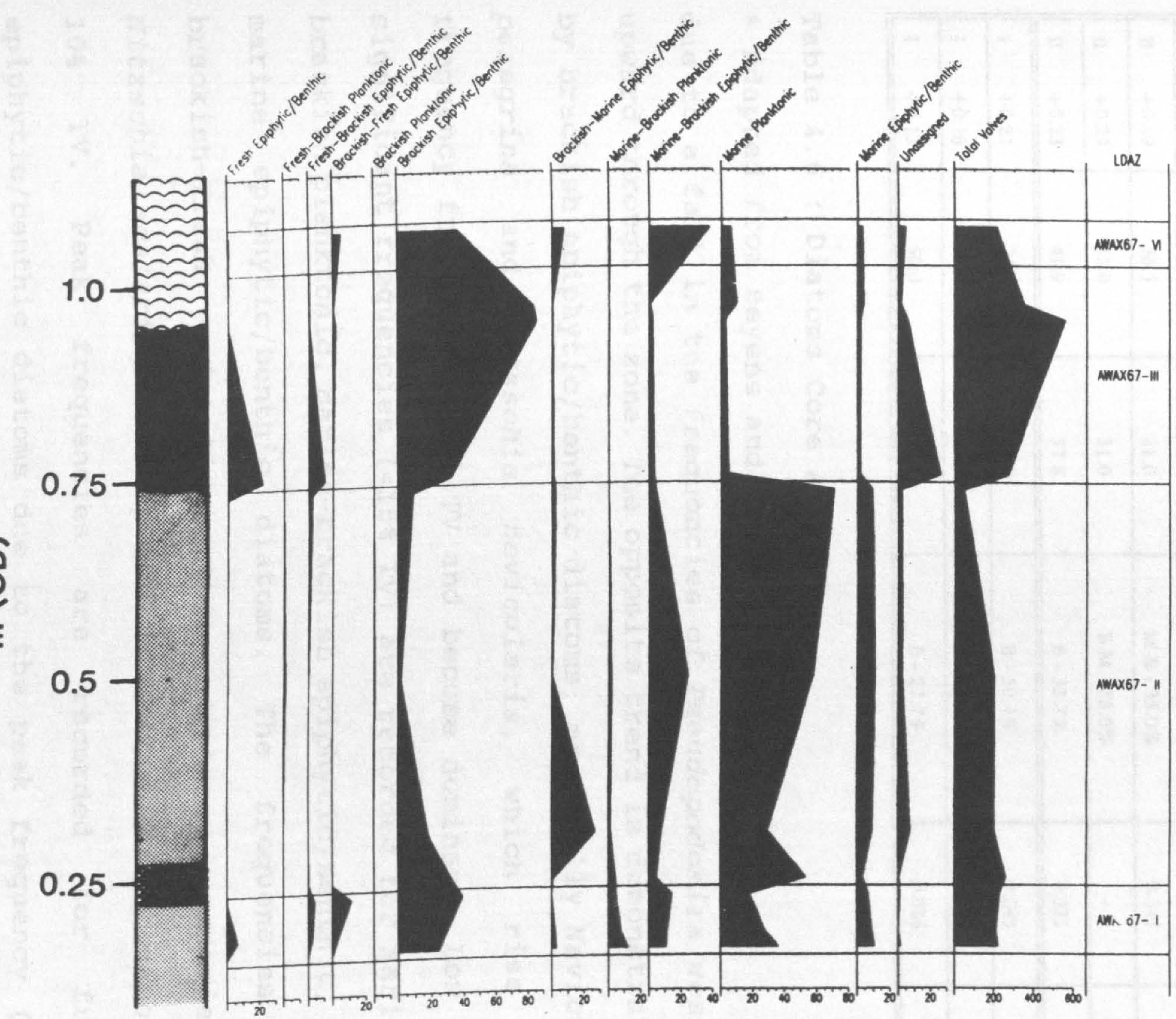

Figure 4.30b: Core AW-AX67, Diatom Diagram : salinity tolerance groups.

Sample Depth m (OD) LDAZ		Total diatoms % Epiphytic / Benthic	Total diatoms % Planktonic	Autochthonous diatoms (Optimal Epiphytic / Benthic group) *	Allochthonous diatoms (Fresh-water) *	Allochthonous diatoms (Marine) *
IV	+1.07	89.0	6.7	M-B - 47.6%	0.872	-
III	+0.97	88.5	9.8	B - 83.6%	0.004	0.055
III	+0.95	91.5	2.9	B - 86.8%	0.007	0.048
III	+0.75	69.3	4.8	B - 33.0%	1.048	0.145
II	+0.73	27.3	70.5	M - 13.6%	1.007	-
II	+0.49	36.5	61.0	M-B - 35.0%	0.043	-
II	+0.29	62.0	31.0	B-M - 53.0%	-	0.170
II	+0.23	41.9	57.8	B - 33.7%	0.231	0.012
I	+0.21	76.1	23.0	B - 50.4%	0.093	0.413
I	+0.16	64.0	36.0	B - 34.1%	0.276	0.598
I	+0.13	50.0	48.6	B - 27.7%	0.036	0.773

Table 4.6 : Diatoms Core AW-AX67.

* Adapted from Beyens and Denys (1982).

due to a fall in the frequencies of *Pseudopodosira westii* upward through the zone. The opposite trend is demonstrated by brackish epiphytic/benthic diatoms, especially *Navicula peregrina* and *Nitzschia navicularis*, which rise in frequency from 24% to 38% TV and become dominant. Low but significant frequencies (<15% TV) are recorded for marine-brackish planktonic, marine-brackish epiphytic/benthic, and marine epiphytic/benthic diatoms. The frequencies of brackish-fresh epiphytic/benthic diatoms, especially *Nitzschia scalaris*, rise upward from effectively zero to 10% TV. Peak frequencies are recorded for fresh epiphytic/benthic diatoms due to the peak frequency (11% TV) of *Eunotia monodon*. Epiphytic/benthic diatoms dominate at the base of the zone and increase in frequency to the top at 76.1% TV, with planktonic valves at 23.0% TV. The optimal diatom group is brackish throughout (table 4.6); its dominance increasing upwards (27.7-50.4% TV) with a coincident reduction in the ratio of marine allochthonous diatoms.

LDAZ AW-AX67 - II : +0.22 - +0.74m OD

Predominant species : *Paralia sulcata, Pseudopodosira westii, Nitzschia navicularis, Diploneis didyma.*

Predominant ecological groups : *Melosira sulcata, Navicula digitoradiata var. minima.*

Marine planktonic diatoms, particularly *Pseudopodosira westii* and *Paralia sulcata*, are dominant in this zone (figure 4.30a&b), initially at 52% TV and falling slightly before increasing to 70% TV at the top. Marine-brackish epiphytic/benthic diatoms, especially *Diploneis didyma*, increase from 5% to 23% TV in the middle of the zone, but then fall to 5% TV at the top. A peak in the frequencies of *Nitzschia punctata* produces a peak in the frequencies of brackish-marine epiphytic/benthic diatoms; reaching a value of 26% TV. Initial relatively high frequencies (30% TV) for brackish epiphytic/benthic diatoms, especially *Nitzschia navicularis*, fall upward and remain low at ~10% TV. Initially the optimal diatom group is brackish, becoming brackish-marine, marine-brackish and eventually marine at the top of the zone. High percentages of planktonic diatoms (31.0-70.5% TV) are present throughout the zone with correspondingly low percentages of epiphytic/benthic valves (table 4.6).

LDAZ AW-AX67 - III : +0.74 - +1.02m OD

Predominant species : *Diploneis interrupta, Pinnularia microstauron, Navicula peregrina.*

Predominant ecological groups : *Diploneis interrupta, Navicula digitoradiata var. minima.*

The frequencies of brackish epiphytic/benthic diatoms increase upward from 32% TV to ~84% TV and dominate the zone (figure 4.30b). The opposite trend is exhibited by fresh epiphytic/benthic diatoms, especially *Pinnularia microstauron*, which fall from 20% to 0% TV, and also brackish-fresh epiphytic/benthic diatoms, especially *Cymbella aspera* and *Anomoeoneis sphaeaphora*, which fall upward from 9% to 0% TV. Low frequencies (<5% TV) of marine-brackish epiphytic/benthic and marine planktonic diatoms are recorded throughout the zone, although the latter increase at the top. The optimal diatom group is brackish throughout (table 4.6); initially at low frequencies (33% TV) but increasing in the upper part of the zone (~85% TV). Epiphytic/benthic diatoms dominate, ranging from 69.3% TV to 91.5% TV, with low percentages of planktonic valves. At the base of the zone, high ratios of allochthonous diatoms are also present.

LDAZ AW-AX67 - IV : +1.02 - +1.08m OD

Predominant species : *Diploneis didyma, Navicula peregrina, Nitzschia navicularis.*

Predominant ecological groups : *Navicula digitoradiata var. minima*

The frequencies of brackish epiphytic/benthic diatoms and marine-brackish epiphytic/benthic diatoms dominate the zone at ~37% TV (figures 4.30a&b), with *Navicula peregrina* and *Nitzschia navicularis* contributing to the former group and *Diploneis didyma* the latter. Significantly low frequencies (<10% TV) are recorded for brackish-fresh epiphytic/benthic, brackish-marine epiphytic/benthic and marine planktonic diatoms. Epiphytic/benthic diatoms dominate (89% TV) with subordinate planktonic diatoms (6.7% TV). The optimal diatom group is marine-brackish at 47.6% TV (table 4.6).

4.3.5.2 Core AW-AX67 : Pollen analysis.

LPAZ AW-AX67 - I : +0.15 - +0.17m OD

Predominant species : Gramineae, Cyperaceae, Chenopodiaceae.

Gramineae pollen are high in frequency (66% TLP) and dominate the assemblage (figure 4.31a), with Cyperaceae at a relatively low frequency of 10% TLP. Low but significant frequencies of *Quercus*, *Alnus* and *Corylus* also occur. The frequencies of Chenopodiaceae are relatively low at 6% TLP.

LPAZ AW-AX67 - II : +0.17 - +0.20m OD

Predominant species : Gramineae, Cyperaceae, *Potamogeton*, *Typha angustifolia*.

The frequencies of *Typha angustifolia* pollen are high

Figure 4.31a: Core AW-AX67, Pollen Diagram : species (legend figure 4.20b).

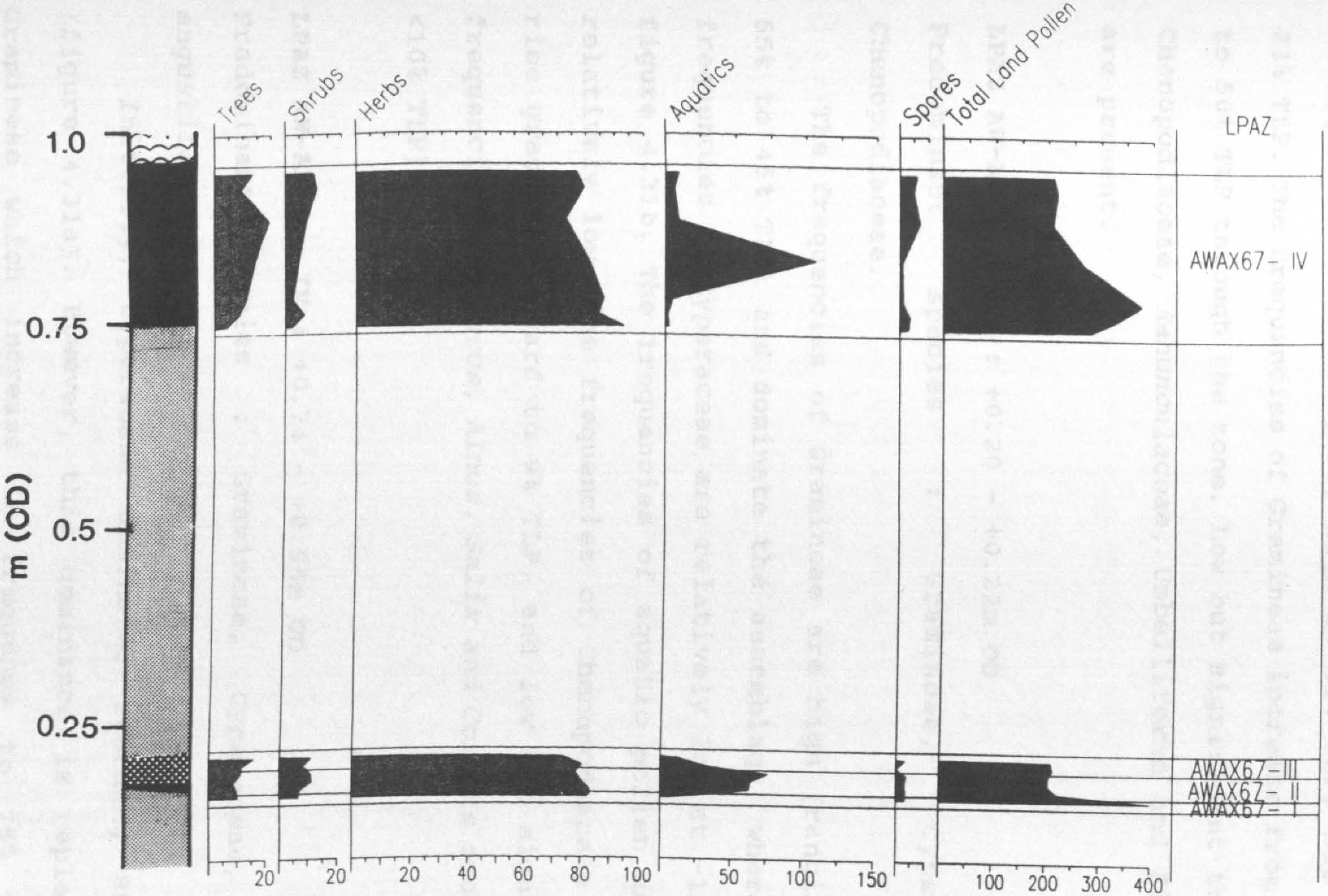

Figure 4.31b: Core AW-AX67, Pollen Diagram : groups (legend figure 4.20b).

and dominate the assemblage at ~50% TLP throughout the zone (figure 4.31a). *Potamogeton* are also relatively high in frequency and increase upward to 16% TLP. The frequencies of Cyperaceae are initially high at 33% TLP, but fall to 23% TLP. The frequencies of Gramineae increase from 39% TLP to 50% TLP through the zone. Low but significant traces of Chenopodiaceae, *Ranunculaceae*, Umbelliferae and Filicales are present.

LPAZ AW-AX67 - III : +0.20 - +0.22m OD

Predominant species : Gramineae, Cyperaceae, Chenopodiaceae.

The frequencies of Gramineae are high (ranging from 55% to 45% TLP) and dominate the assemblage, whereas the frequencies of Cyperaceae are relatively low at ~15% TLP, figure 4.31b. The frequencies of aquatic pollen are also relatively low. The frequencies of Chenopodiaceae pollen rise gradually upward to 9% TLP, and low but significant frequencies of *Quercus*, *Alnus*, *Salix* and *Corylus* occur (all <10% TLP).

LPAZ AW-AX67 - IV : +0.74 - +0.96m OD

Predominant species : Gramineae, Cyperaceae, *Typha angustifolia*.

Initially, Cyperaceae dominate, reaching 85% TLP (figure 4.31a). However, this dominance is replaced by Gramineae which increase in frequency to 75% TLP as Cyperaceae fall to 12% TLP. At the top of the zone, both pollen types are at ~35% TLP. Chenopodiaceae pollen increase gradually upward, although at a low frequency, reaching 6% TLP at the top of the zone. *Plantago maritima* is present as a trace. The frequencies of tree and shrub pollen increase gradually upward, although they fall in frequency at the top of the zone. This is especially true of *Alnus*, which reaches 15% TLP, and *Corylus*. A significant peak of *Typha angustifolia* occurs in this zone; reaching >100% TLP.

4.4 Chronostratigraphy of Scotney Marsh : Results.

The locations of the samples selected for radiocarbon analysis, along with the reason for their selection and the ages obtained are summarised in table 4.7. The dating of these sedimentary units was completed in order to allow the environmental changes in Scotney Marsh, such as possible changes in sea-level and the emplacement and / or removal of protective gravel barriers, to be dated.

Sample Name : Core, Altitude, *Laboratory Number* Location.	Environment of deposition and the significance relative to sea-level. *(Troels-Smith, 1955)*	^{14}C Age (Calibrated years BP, with 2 sigma analytical uncertainty)	Uncalibrated radiocarbon age +/- 1 sigma
Sample 1 : G60 -2.28 to -2.235m OD *Beta-81363* (TR026203)	Minimum age of the gravel barrier emplacement. Peat development traces the removal of marine conditions. Regressive contact. *(Sh3,Dh1,Th²+,As+)*	3370 to 2970	3020 +/- 70 BP
Sample 2 : G60 -2.235 to -2.185m OD *Beta-81364* (TR026203)	Sea-Level index point (SLIP): Upper contact of terrestrial peat becoming saltmarsh peaty-clay under marine conditions. Transgressive contact. *(Sh3,Dh1,Th³+,As+)*	3375 to 3070	3050 +/- 60 BP
Sample 3 : AY17 +0.52 to +0.57m OD *Beta-81365* (TR034210)	SLIP: Lower contact of peat, as mudflat became saltmarsh, tracing the removal of marine conditions. Regressive contact. *(Sh2,Dh2,As+,Th²+)*	3355 to 2980	3010 +/- 60 BP
Sample 4 : AY17 +0.72 to +0.77m OD *Beta-81366* (TR034210)	SLIP: Upper contact of peat which is overlain by marine deposited sediments. Transgressive contact. *(Sh4,Dh+,Th³+)*	2710 to 2585 and 2510 to 2320	2380 +/- 60 BP
Sample 5 : A-B27 +0.96 to +1.01m OD *Beta-31367* (TR031208)	SLIP: Upper contact between saltmarsh and freshwater dominated peat, tracing the complete oremoval of marine conditions. Regressive contact. *(Sh4,Dh+,Th³+)*	2795 to 2710 and 2620 to 2500	2610 +/- 60 BP
Sample 6 : AW63 +0.12 to +0.05m OD *Beta-81368* (TR023204)	SLIP: Lower contact between mudflat becoming saltmarsh and eventually semi-terrestrial peat. Minimum date for gravel emplacement. Regressive contact. *(Sh3,Th²1,Dh+,As+)*	3715 to 3565	3410 +/- 40 BP
Sample 7 : AW63 +0.74 to +0.59m OD *Beta-81369* (TR023204)	SLIP: Entire peat unit sampled to give an age for the period of organic sedimentation caused by a change in sea-level. *(Sh3,Dh1,Th³+)*	3265 to 2930	2950 +/- 60 BP
Sample 8 : AW-AX67 +0.16 to +0.21m OD *Beta-81370* (TR022204)	SLIP: Entire peat unit sampled to give an age for the period of organic sedimentation caused by a change in sea-level. *(Sh3,Dh1,Th¹+)*	4060 to 4040 and 3990 to 3700	3580 +/- 60 BP
Sample 9 : AW-AX67 +0.76 to +0.85m OD *Beta-81371* (TR022204)	SLIP: Lower contact of the upper peat unit, saltmarsh peat-clay becomes terrestrial peat, indicating the removal of marine conditions. Regressive contact. *(Sh3,Th³1,Dh+,As+)*	3220 to 2795	2850 +/- 60 BP
Sample 10: AW-AX67 +0.85 to +0.94m OD *Beta-81372* (TR022204)	SLIP: Upper contact of the upper peat unit, terrestrial peat becomes saltmarsh peaty-clay, indicating the return of marine conditions. Transgressive contact. *(Sh4,Th³+,Dh+,As+)*	2950 to 2725	2690 +/- 80 BP

Table 4.7 : Radiocarbon Dating of Scotney Marsh.

5 Interpretation of the Palaeoenvironmental Analysis of the Typecores.

5.1 Introduction

The results of the palaeoenvironmental analysis will be interpreted here to provide a palaeoenvironmental reconstruction for each of the typecore sites. These reconstructions will be considered together with palaeoenvironmental studies of other sites both on Romney Marsh and also, where relevant, other barrier and back-barrier environments elsewhere.

5.2.1 Core AY17 : Lithostratigraphy

Table 5.1

Unit	Description	Altitude (OD)
5	Orange/grey oxidation mottled clay with silt.	+1.52 - +0.77m
4	Dark brown well humified peat.	+0.77 - +0.48m
3	Dark grey-browny clay with silt and substantial organic content at the top of the unit. (transitional)	+0.48 - +0.28m
2	Blue-grey clay with silt and rare laminations.	+0.28 - -2.48m
1	Gravel (not recovered).	Below -2.48m

5.2.2 Core AY17 : Palaeoenvironmental Interpretation

The environment of deposition following gravel emplacement is recorded within unit 2. No palaeoenvironmental indicators were recovered from the basal gravel (unit 1). However, during the deposition of unit 2 marine conditions prevailed, as illustrated by the predominance of the *Melosira sulcata* group throughout LDAZ

AY17 - I, summarised in table 5.2. This diatom group is indicative of a tidal inlet, whilst the geomorphological context of this site points to it being part of the main back-barrier semi-enclosed lagoon that formed behind the initial protective gravel barrier extending across Rye Bay. At the base of unit 2 (below -1.905m OD) relatively coarse grained, well sorted, fine sands are indicative of deposition under wave action. However, the absence of a consistent negative skew is somewhat contradictory to this.

The environment of deposition is also indicated to have been a tidal inlet from the employment of the statistical particle size data (mean grain size and standard deviation) in an interpretive model developed by Tanner (1991). Figure 5.1 illustrates that the sediments of core AY17 cluster on the closed basin / river transition. The same trend is recorded by Lario *et al.* (1996) from the Guadalquivir Marsh, south-west Spain, where sediments were being deposited in an infilling back-barrier estuary. The notable exception to this overall trend is that a number of the sediments from the base of the core are coarser, and are more indicative of a beach environment in which wave action dominated. The samples from the lower part of core AY17 appear well within the closed basin envelope and migrate progressively toward the boundary with the river environment envelope upward through the core (as indicated by the arrow in figure 5.1).

It would appear, therefore, that the back-barrier environment was becoming more enclosed, and possibly

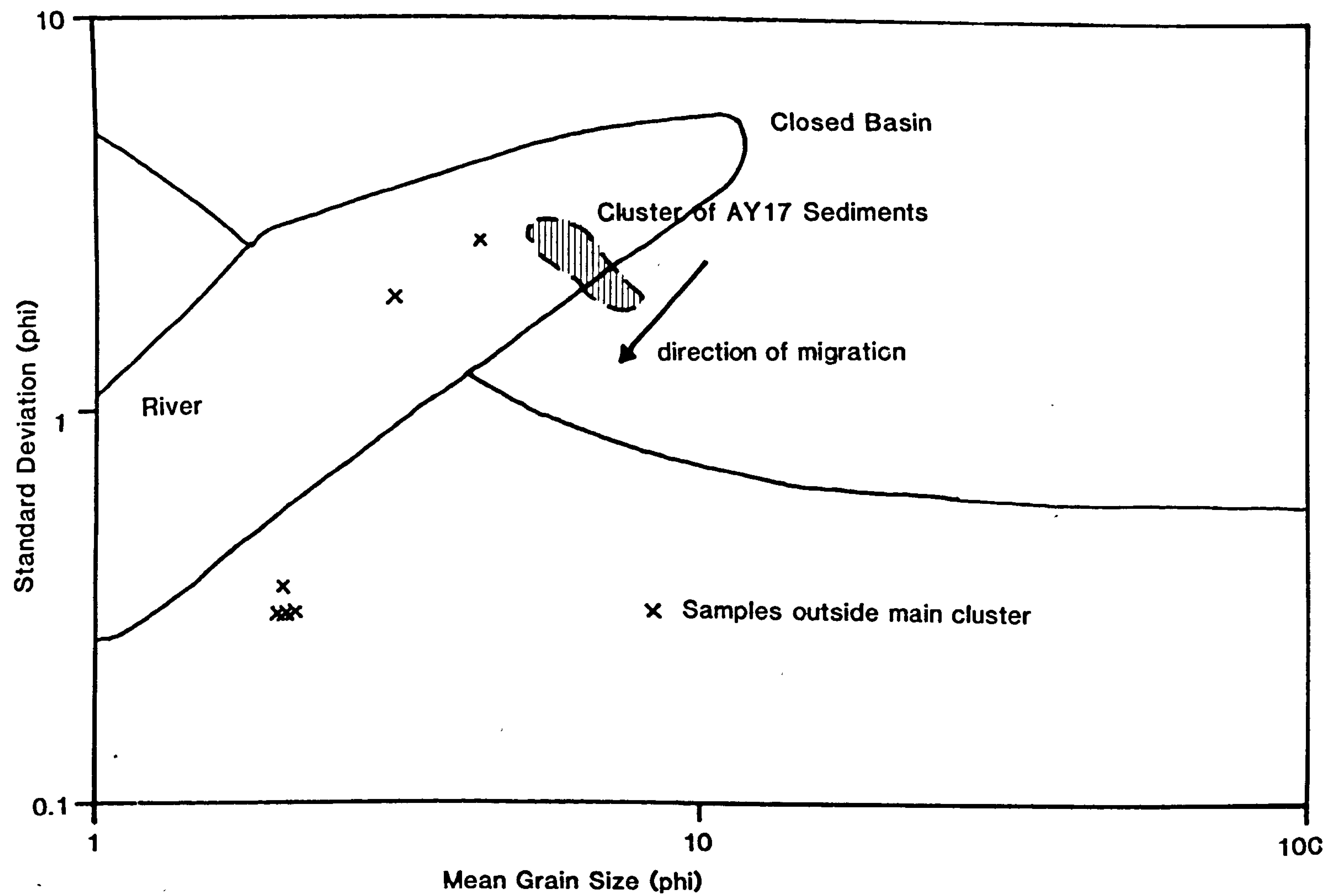

Figure 5.1: Environment of deposition, core AY17, utilising mean grain size and standard deviation, developed by Tanner (1991).

shallower, as a reduction in the linkage with the open sea is suggested by the decrease in the frequencies of marine epiphytic/benthic diatoms, especially *Cymatosira belgica*, above -1.63m OD. The diatom assemblage is initially indicative of a lower mudflat, but becomes more indicative of the higher mudflat of Denys (1994) upward through unit 2. Also, on occasions during the deposition of unit 2, the water became less marine, as illustrated by decreases in the frequencies of marine planktonic diatoms. At these same levels, corresponding rises in the frequencies of brackish, brackish-marine and marine-brackish epiphytic/benthic diatoms, especially *Navicula digitoradiata* and *Nitzschia punctata*, are observed at altitudes of -1.99m OD, -1.62m OD and -1.23m OD. This might also indicate a shallowing of the back-barrier environment, with an increased representation of the *Navicula digitoradiata var. minima* group at these altitudes. This diatom group is indicative of intertidal to subtidal mudflats or subtidal basins and lagoons (Vos and de Wolf, 1988).

The particle size distributions of the sediments of units 2 and the lower part of unit 3 indicate that the environment was probably a tidal lagoon or inlet. The sediments are very poorly sorted medium to very-fine silts, with changes in the depth of the water and tidal velocity being represented by subtle variations in the sorting and mean grain size. Sediments from Denge Marsh with similar sedimentary characteristics have been described by Plater (1992) as tidal muds. Postma (1967) illustrated that a lag existed at the turn of the tide when current velocity was reduced to zero. Thus settling out of the fine sediments occurs either side of high water. It is under conditions such as these that the sediments of the upper part of core AY17 were probably deposited.

The subtle variations in the particle size parameters of core AY17 (figure 4.22) record slight variations in the depositional environment, although it is apparent to have remained one of tidal mud sedimentation throughout. Some samples exhibited a more near symmetrical trend. In such cases it is suggested that deposition occurred near to the margin of tidal influence, as proposed by Long *et. al.* (1996).

The variations in the depositional environment may be explained by phases of gravel movement, possibly after a storm, when, aperiodically, the back-barrier may have become somewhat cut off from the open sea. The reduction in the frequencies of marine epiphytic/benthic diatoms would also indicate occasional closure. The effect of removing or reducing the exposure to open sea would have allowed the site to become more brackish, with marine planktonic input, until such a barrier was removed or became degraded.

The dominance of the planktonic diatoms make palaeoenvironmental interpretation somewhat problematic, as they are, by definition, allochthonous (Vos and de Wolf, 1993). However, they demonstrate clear tidal access to the site as the marine and marine-brackish diatoms would be brought in at high water with the tide. Towards the top of unit 2, in the lower part of LDAZ AY17 - II, a significant increase in the brackish influence occurred, as indicated by a significant increase in the frequencies of *Nitzschia navicularis* with a corresponding reduction in the marine influence. The diatom assemblage at this point becomes more representative of an upper mudflat (Denys, 1994). However, in unit 3, the marine influence is again dominant, but punctuated by significant contributions from the brackish to fresh-brackish epiphytic/benthic diatoms. Importantly, the frequencies of *Pinnularia viridis*, of the *Hantzschia amphioxys* group (Vos and de Wolf, 1993), increase. This is indicative of deposition on a an upper mudflat to saltmarsh (Denys, 1994) and corresponds with the increased organic material in the sediments at the top of unit 3.

The relatively high frequencies of the *Melosira sulcata*, *Navicula digitoradiata var. minima* and *Hantzschia amphioxys* groups suggest that the environment during the deposition of the upper part of unit 3 was intertidal, with a significant marine input but also with periodic sub-aerial exposure allowing the development of saltmarsh (Vos and de Wolf, 1988). This is also illustrated by the variations in the optimal diatom group. In addition, the increases in the percentage of epiphytic/benthic diatoms makes the interpretation less problematic than for the unit below (Vos and de Wolf, 1993).

The factors that probably allowed peat to accumulate may have been a reduction in the rate of relative sea-level rise or a rapid increase in the sedimentation rate allowing vegetation to colonise a sub-aerial substrate. The emplacement of a protective gravel barrier might provide an alternative explanation. As discussed above, gravel barrier formation would have caused the environment to become more brackish to freshwater, but would also have offered protection from the erosive wave and tidal current influences, and encouraged marginal colonisation by peat communities.

The accumulation of peat (unit 4) in core AY17 corresponds with a removal of marine conditions, although initially the diatom assemblage is characteristic of an upper mudflat to saltmarsh (Denys, 1994). This was followed by the dominance of brackish epiphytic/benthic diatoms, especially *Diploneis interrupta* (LDAZ AY17 - III) which is characteristic of a *Phragmites* saltmarsh environment (Denys, 1994). Pollen evidence at the base of unit 4 (LPAZ AY17 - I) also records the removal of marine conditions as, initially, an upper saltmarsh environment, as indicated by the relatively high frequencies of Chenopodiaceae, was replaced by the dominance of Gramineae. This is similar to the ecological trends associated with initiation of peat at Brookland (Long and Innes, 1995a).

Following the removal of the marine influence, the environment eventually became a freshwater reedswamp; with the aquatic pollen *Potamogeton* and *Typha angustifolia* dominating in LPAZ AY17 - II, in addition to continued high frequencies of Gramineae. Long and Innes (1995b) noted that *Typha angustifolia* often formed the dominant vegetation component at the transition from upper saltmarsh to freshwater reedswamp. This dominance falls upward in core AY17, as do the frequencies of Gramineae; being replaced by Cyperaceae as the dominant taxa. This is indicative of a gradual drying out of the environment as the local fresh-water table fell (Long and Innes, 1993). The regressive contact between +0.52 and +0.55m OD, as illustrated by the diatom assemblage and the vegetation succession, has been radiocarbon dated to 3355-2980 cal. yrs. BP.

Further evidence of a drying out of the environment of deposition is present through LPAZ AY17 - III, as Cyperaceae continues to dominate in combination with wetland herb types such as *Rumex* and Umbelliferae. This suggests that a sedge fen environment may have developed. Peaks in the frequencies of *Potamogeton* indicate that freshwater pools may have been present on the peat surface nearby.

The frequencies of Cyperaceae become reduced to a level similar to that of Gramineae in LPAZ AY17 - IV as the frequencies of *Corylus* and *Betula* increase along with *Rumex*. This is indicative of a continued drying trend as a fen carr was established. The very low frequencies of aquatic pollen also support this. However, gradual increases in the low frequencies of Chenopodiaceae suggest that saltmarsh conditions were established nearby. This is further suggested in LPAZ AY17 - V where an increase in the frequencies of aquatic pollen indicate a wetter environment. Further increases in the frequencies of

Chenopodiaceae indicate that adjacent saltmarsh conditions continued to develop. Furthermore, high frequencies of Cyperaceae and Gramineae pollen were still present. This transgressive contact is similar to that described by Tooley and Switsur (1988) from the upper contacts of the biogenic sediments at Tishy's Sewer, Broomhill. The presence of a *Phragmites* saltmarsh at the site is further illustrated by the predominance of *Diploneis interrupta* at the top of LDAZ AY17 - III. The transgressive contact at the top of this peat unit has been dated between altitudes of +0.72 and +0.77m OD, and an age of either 2710-2585 or 2510-2320 cal. yrs. BP has been obtained.

Into LDAZ AY17 - IV, across the contact between units 4 and 5, a sudden rise in the frequencies of marine planktonic species, *Paralia sulcata* in particular, dominate the assemblage, and fluctuate with *Diploneis interrupta* throughout the zone. This indicates a return to marine conditions, although the high percentage of planktonic diatoms indicates a significant allochthonous component with the optimal brackish diatom group. At the base of unit 5, poorly to very poorly sorted fine to very fine silts were deposited. Plater (1992) suggested that such sediments were indicative of tidal mud deposition during slack water. However, a slight increase in the energy of the environment is indicated by the sediments in the upper part of unit 5, at altitudes of +0.845m OD and between +0.99 - +1.135m OD. These sediments exhibit a slightly higher mean grain size than those underlying and, more significantly, are strongly fine-skewed; the cause of which is probably a higher tidal flow velocity.

Importantly, no marine epiphytic/benthic diatoms are present in unit 5 and, thus, the environment of deposition must have become relatively cut off from the open sea. Thus, the marine contribution is likely to have originated from overwashing of the protective gravel barrier at high tide or from gradual sea-level rise leading to the flooding of a tidal channel thought to have existed to the north of Scotney Marsh (Long and Innes, 1995a). The marine water input would have been flooding of the site at high tide, as indicated by the high frequencies of marine planktonic diatoms and lack of marine epiphytic/benthic diatoms. Deposition was probably taking place on a high mudflat at this time (Denys, 1994).

The cause of the return of the marine dominance may have been due to the overtopping or the removal of a protective gravel barrier, allowing marine dominance to return to the back-barrier environment. However, it is more likely that a continued gradual sea-level rise led to the inundation of the peat and deposition of unit 5. This would have resulted, first, in a rise in the water table at the site, followed by the development of saltmarsh conditions and, finally, the inundation of the plant communities. This continued rise in sea-level would have enabled marine conditions to return to the site and yet still offered some protection, thus preventing erosion of the peat. Importantly, no indication of wave action is evident from the particle size distributions of unit 5. This would have been required if the peat were to be eroded. The absence of wave action is to be expected of deposition in a partially enclosed back-barrier environment.

Table 5.2 : Core AY17 Summary

Altitude m (OD)	Sediment Type / Particle Size Analysis	Predominant Diatoms	Predominant Pollen	Environment of Deposition	^{14}C Age (Cal. yrs. BP, 2 sigma)
+1.52 to +0.79	Unit 5: Orange-grey oxidation mottled clay with silt. (Initially tidal mud deposition, however, mean grain size increased upward and sediments became more fine skewed).	Autochthonous *Diploneis interupta* dominates with allochthonous *Paralia sulcata* and subordinate *Pseudopodasira westii*.	N/S	Up to +0.87m OD a brackish tidal mudflat exists into which significant inputs of marine water occur. The energy of the environment appears to increase upward.	N/S
+0.79 to +0.48	Unit 4: Dark brown well-humified peat.	Brackish epiphytic/benthic diatoms dominate at the top. Brackish epiphytic/benthic diatoms dominate at the base.	Initially high frequencies of Chenopodiaceae are followed by high frequencies of Gramineae and aquatic pollen. Upward, there is dominance of Cyperaceae and wetland herb pollen. Toward the top there is an increase in Gramineae, Chenopodiaceae and aquatic pollen types.	A brackish *Phragmites* saltmarsh becoming a freshwater pool / reedswamp. Progressive drying out recorded by sedge fen with pools and then fen carr. Towards the top the environment becomes wetter again and, eventually, a saltmarsh with marine dominance returns.	2710 to 2585 or 2510 to 2320 3355 to 2980
+0.48 to +0.08	Unit 3: Grey-brown silty-clay to peaty-clay transition.	Marine dominance with fluctuating brackish and fresh inputs. Presence of the *Hantzschia amphyioxys* group.	Gramineae and Cyperaceae dominate, with significant frequencies of Chenopodiaceae and Potamogeton.	Progressive change from tidal flats to intertidal saltmarsh.	N/S
+0.08 to -2.48	Unit 2: Blue-grey sand with silt and some clay. (Very poorly sorted, medium to very fine silts).	Marine dominance, especially the *Melosira sulcata* group. A gradual reduction in Marine epiphytic/benthic diatoms. Increased Brackish influence at the top, *Nitzschia navicularis*.	N/S	Tidal mudflats in a back-barrier environment, gradually shallowing with periodic variations in salinity.	N/S
Below -2.48	Unit 1: Gravel.	N/S	N/S	N/S	N/S

5.3.1 Core A-B27, Lithostratigraphy.

Table 5.3

Unit	Description	Altitude (OD)
7	Orange, oxidation mottled sandy-silt.	+1.31 - +1.27m
6	Brown sand with silt and clay.	+1.27 - +1.24m
5	Dark brown well humified peat with some clay (transitional top).	+1.24 - +1.12m
4	Brown well humified peat with clay and silt.	+1.12 - +0.95m
3	Light brown clay with organics and silt (transitional).	+0.95 - +0.72m
2	Blue-grey silt and clay.	+0.72 - +0.41m
1	Gravel (not recovered).	Below +0.41m

5.3.2 Core A-B27 : Palaeoenvironmental Interpretation

Following the deposition of unit 1; the basal gravel in which no palaeoenvironmental indicators were discovered, unit 2 was laid down. The upper part of unit 2 (sampled above +0.66m OD) was deposited in a predominantly brackish environment, as shown by the base of LDAZ A-B27 - I in which there are significant contributions from fresh-brackish and brackish-fresh epiphytic/benthic and marine planktonic diatoms. The *Navicula digitoradiata var. minima* and *Melosira sulcata* groups are dominant in this zone. An assemblage characteristic of an upper mudflat is present throughout (Denys, 1994). This illustrates that deposition occurred on an intertidal mudflat, and the absence of marine epiphytic/benthic diatoms indicates that marine planktonic diatoms were supplied to the mudflat as each high tide covered the mudflats. The environment of deposition of unit 2 is, therefore, similar to that recorded in unit 2 in core AY17, summarised in table 5.4.

This intertidal environment persisted into the base of unit 3 where the organic content of the sediment increases. Pollen from these sediments reveals that an environment of deposition characteristic of a saltmarsh, with high frequencies of Chenopodiaceae, Gramineae and Cyperaceae present (LPAZ A-B27 - I). The saltmarsh which developed at the site appears to have colonised the intertidal flats gradually. This was probably caused by continued sedimentation in the back-barrier environment, leading to vegetation colonising the emerging sub-aerial substrate.

Towards the top of unit 3, the frequencies of Chenopodiaceae become reduced with a corresponding rise in the frequencies of Cyperaceae. This suggests that exposure to marine conditions was reduced further, allowing sedges to become established. This continued into the base of LPAZ A-B27 - II. This trend is also supported by diatom evidence, which indicates that the environment became more fresh-brackish as brackish to marine conditions attenuated. The environment of deposition became dominated by the *Epithemia zebra* group, with a significant contribution by the *Navicula digitoradiata var. minima* group at the base of LDAZ A-B27 - II. The optimal diatom group is fresh-brackish with a relatively low level of allochthonous marine diatoms throughout. These assemblages are indicative of a freshwater reedswamp with saltmarsh and sedge fen at the margins (Vos and de Wolf, 1988). The vegetation assemblage records a regressive contact, which is verified by the diatom evidence, *i.e.* initially marine to marine brackish becoming fresh-brackish upward. The regressive contact at the base of the peat (unit 4) has been dated to between 2795-2710 and 2620-2500 cal. yrs. BP.

The fresh to brackish water environment continued through the deposition of unit 4, with fresh-brackish to brackish epiphytic/benthic conditions dominating and fresh epiphytic/benthic diatoms increasing towards the top of the unit. This coincides with a significant increase in the frequencies of the aquatic taxon *Typha angustifolia* into unit 5. The environment became more waterlogged, as illustrated by the increase of aquatic pollen. A pollen assemblage similar to that of LPAZ A-B27 - III has been described by Long and Innes (1995b) as a freshwater reedswamp.

The environment of deposition continued to become more freshwater upward into unit 5, with further increases in fresh and fresh-brackish epiphytic/benthic diatoms, especially *Eunotia monodon* and *Epithemia turgida*. Corresponding decreases in the frequencies of brackish-fresh to marine diatoms are also recorded. The dominance of the *Epithemia zebra* group illustrates that the environment was still one of a freshwater dominated back-barrier environment (Vos and de Wolf, 1988). This is also suggested by the high frequencies of *Typha angustifolia* and *Potamogeton* pollen, which indicate that freshwater reedswamp conditions prevailed at the margins of which sedge fen may have existed. This is recorded in the upper part of LPAZ A-B27 - III.

A significant change in environment occurred into unit

6 with a dramatic increase in the frequencies of brackish to marine diatoms, particularly the marine planktonic diatom *Paralia sulcata*. Dominance of the *Melosira sulcata* and *Navicula digitoradiata var. minima* groups occur in LDAZ A-B27 - IV. These indicate that a sudden increase in the marine influence took place. However, as the optimal diatom groups are still brackish to fresh-brackish, a significant marine to marine-brackish allochthonous component is clearly present in the diatom assemblage. This is also represented by the significant increases in the number of planktonic relative to epiphytic/benthic diatoms. This change may have been due to overtopping or removal of a gravel barrier, or more extensive marine flooding of the back-barrier environment. However, a gradual sea-level rise is unlikely to have been the cause as the change appears to have been relatively rapid from the sudden change in vegetation. The diatom assemblage is characteristic of a saltmarsh to upper mudflat (Denys, 1994). A rapid reduction of the frequencies of aquatic pollen also occurs into unit 6 in LPAZ A-B27 - IV, with Cyperaceae and Gramineae becoming the dominant taxa. In addition, a gradual increase in the frequencies of Chenopodiaceae upward is indicative of the vegetation responding to the increased marine influence.

The environment of deposition appears to have reverted rapidly to an intertidal to supratidal saltmarsh from the freshwater reedswamp with sedge fen. This is suggested because taxa indicative of a number of palaeoenvironments are present; saltmarsh indicators, aquatic pollen types, and relatively high frequencies of Cyperaceae and Gramineae. A coastal reedswamp may well have been present at the transition from marine to freshwater dominance as described by Waller (1994b). Drier conditions may have prevailed nearby, allowing sedge fen to exist.

A return to autochthonous fresh-brackish epiphytic/benthic diatom dominance occurs at the base of unit 7. Again *Epithemia turgida* is the dominant taxon, indicating that a return to the predominantly fresh-brackish conditions occurred. It would appear that the cause of deposition for unit 6 was a short-lived period of marine deposition, as the conditions at the site again became predominantly fresh to brackish again into unit 7. This would not be expected if permanent gravel barrier removal were responsible for the marine-dominated deposition of unit 6. Therefore, it appears that the inundation of this area of the main back-barrier environment was facilitated by the flooding of a tidal channel known to exist to the north of Scotney Marsh. This flooding would have been driven by the continued rising sea-level eventually leading to the periodic flooding of the tidal channels. After such periods of inundation the back-barrier environments returned to fresh-brackish water dominated deposition, as recorded in unit 7. However, the periods of increased and decreased marine influence recorded in units 6 and 7 may suggest that short-term degradation or overtopping of a gravel barrier may provide a more credible interpretation for the sudden marine dominance of unit 6.

Table 5.4 : Core A-B27 Summary

Altitude m (OD)	Sediment Type	Predominant Diatoms	Predominant Pollen	Environment of Deposition	^{14}C Age (Cal. yrs. BP, 2 sigma)
+1.31 to +1.27	Unit 7: Orange, oxidation mottled sandy-silt.	Fresh-brackish diatoms dominate, especially *Epithemia turgida*.	N/S	Fresh-brackish tidal lagoon.	N/S
+1.27 to +1.24	Unit 6: Brown sand with silt and clay.	Brackish to marine diatoms dominate, especially *Paralia sulcata* and the *Navicula digitoradiata var. minima* group.	Cyperaceae, Gramineae and some Chenopodiaceae at the upper contact.	Coastal reedswamp to intertidal saltmarsh indicating a relatively sudden return to marine dominance.	N/S
+1.24 to +1.12	Unit 5: Dark brown well humified peat with some clay, transitional upper contact.	Fresh to fresh-brackish diatoms dominated especially *Epithemia turgida* and *Eunotia monodon*.	Gramineae dominate along with the aquatic pollen types. Some wetland herb pollen types also present.	Increased waterlogging leading to a freshwater reedswamp.	N/S
+1.12 to +0.95	Unit 4:Brown well humified peat with clay and silt.	Fresh-brackish diatoms dominated with increased freshwater diatoms upward.	Cyperaceae and Gramineae dominate, with increasing aquatic pollen toward the top. Significant inputs of Corylus and Alnus.	Freshwater lagoon with drier sedge fen conditions and alder carr nearby.	2795 to 2710 and 2620 to 2500
+0.95 to +0.72	Unit 3: Light brown lithologically transitional clay with organics and some silt.	At the base the same as unit 2, toward the top fresh-brackish diatoms dominated, especially the *Epithemia zebra* group with subordinate *Navicula digitoradiata var. minima* group.	Cyperaceae, Gramineae and Chenopodiaceae with some aquatic pollen types. At the upper contact the frequency of Chenopodiaceae reduce as Cyperaceae rises.	Initially intertidal mudflats becoming gradually colonised by saltmarsh and a freshwater sedge fen.	N/S
+0.72 to +0.41	Unit 2: Blue-grey silt and clay.	At the upper contact, brackish diatoms dominate, especially the *Navicula digitoradiata var. minima* group with subordinate *Melosira sulcata* group.	N/S	Intertidal to supratidal conditions in a tidal inlet with input from the open sea.	N/S
Below +0.41	Unit 1: Gravel.	N/S	N/S	N/S	N/S

5.4.1 Core G60, Lithostratigraphy.

Table 5.5

Unit	Description	Altitude (OD)
8	Grey silty-clay, some sandy laminations, organic rich at base.	+0.62 - -1.84m
7	Well humified brown peat.	-1.84 - -1.86m
6	Browny-grey, organic rich peaty silty-clay.	-1.86 - -1.98m
5	Well humified brown peat.	-1.98 - -2.01m
4d	Blue-grey silt matrix with gravel.	-2.01 - -2.03m
4c	Gravel.	-2.03 - -2.07m
4b	Blue-grey silty-clay matrix with some gravel.	-2.07 - -2.11m
4a	Blue-grey clayey-silt.	-2.11 - -2.15m
3	Brown well humified peat.	-2.15 - -2.29m
2	Browny-grey organic peaty-silt.	-2.29 - -2.32m
1	Gravel (not recovered).	Below -2.32m

5.4.2 Core G60 : Palaeoenvironmental Interpretation

At this site, the mixed minerogenic / biogenic sediments of unit 2 were deposited directly on top of the impenetrable gravel (unit 1). The deposition of unit 2 occurred in a predominantly marine environment with significant fresh and brackish water inputs. Very few diatoms occur in LDAZ G60 - I, but those which do occur indicate an environment with a significant marine input, probably a lower mudflat (Denys, 1994). In addition, a saltmarsh environment is suggested as being nearby from relatively high frequencies of *Pinnularia viridis* (Vos and de Wolf, 1988). However, the high percentages of planktonic diatoms make detailed palaeoenvironmental reconstruction somewhat problematic. This site is situated in a hollow in the gravel within the Scotney Marsh trough (as discussed in section 4.2.1.4). Consequently, there is a possibility that an isolated pond may have developed in the gravel hollow.

This is also supported by the pollen in LPAZ G60 - Ia, which is characteristic of a brackish water pond with relatively high frequencies of aquatic pollen and also significant numbers of Chenopodiaceae. High frequencies of Gramineae and Cyperaceae pollen are also present. The frequencies of Chenopodiaceae become dominant (>50% TLP) at the top of unit 2, corresponding with LPAZ G60 - Ib. Thus, the formation of a saltmarsh environment, as described by Long and Innes (1995a) appears to have taken place, see table 5.6. This regressive contact has been dated between altitudes of -2.235 and -2.28m OD at 3370-2970 cal. yrs. BP.

In unit 3, Cyperaceae dominate the pollen assemblage with Gramineae, suggesting that a reduction of the marine influence enabled a sedge fen to develop at the site. Some saltmarsh indicators are still present at a low frequency in LPAZ G60 - II. Upward through unit 3, Gramineae and Cyperaceae continue to dominate, with significant increases in aquatic pollen, especially *Typha angustifolia*, in both LPAZ G60 - III and IV. The environment of deposition would appear to have been one of a sedge fen, with high frequencies of Cyperaceae and other wetland herbs such as Umbelliferae, *Rumex* and *Hydrocotyle*. The increase in the aquatic taxa in the upper part of unit 3 illustrates that a shallow predominantly freshwater pond developed at the site, similar to that described by Long and Hughes (1995) at Wickmaryholme Pit. Saltmarsh to reedswamp conditions prevailed at the margins of the pond, as indicated by the presence of Chenopodiaceae, *Plantago maritima* and Gramineae.

At the top of unit 3, a return to marine dominance is suggested from the increase in the frequency of aquatic taxa and, more specifically, the increase in the frequencies of Chenopodiaceae in LPAZ G60 - IV. The environment at the top unit 3 was predominantly marine but with a significant brackish water element, as illustrated by the presence of *Melosira sulcata* and *Navicula digitoradiata var. minima* groups dominating LDAZ G60 - II. The optimal diatom group at this point is brackish with some allochthonous marine input, and the diatom assemblage is characteristic of an upper mudflat to saltmarsh (Denys, 1994). This succession illustrates a transgressive contact as marine conditions returned to the site. This contact has been dated to 3375-3070 cal. yrs. BP between altitudes of -2.235 and -2.185m OD.

Organic sedimentation ceased at the top of unit 3 as the minerogenic sediments of units 4a-d were deposited. This coincides with a significant increase in marine dominance in LDAZ G60 - III, as indicated, in particular, by the high frequencies of *Paralia sulcata*. Again, the high frequency (>76% TV) of planktonic diatoms make detailed palaeoenvironmental reconstruction more difficult. Significant but low (<10% TV) frequencies of brackish epiphytic/benthic and marine epiphytic/benthic diatoms are also present in this zone. Despite the significant changes in the grain size of the sediments between units 4a-d, little change in the diatom assemblage was observed.

The fact that silts and gravel are deposited on top of the basal organic deposits of core G60 is important as it demonstrates that gravel was still mobile, and that possibly the Forelands gravel complex was still accumulating at this time. Therefore, as sea-level continued to rise, storms were eventually able to overtop the gravel barrier, that until this time, had been protecting the site. However, the overall control of sea-level rise on the nature of sedimentation is indicated by the reversal of the vegetation succession prior to inundation. This led to the deposition of unit 4a, a very well-sorted coarse to very fine silt on a lower to upper mudflat (Denys, 1994). Upwards, this relatively low energy environment was punctuated by periods of energy strong enough to deposit the silts and gravels of units 4b-d. Importantly, this illustrates that at this time gravel was mobile in the environment.

During the deposition of unit 5, marine dominance still prevailed, even though further organic sediments were being deposited. This upper organic facies is probably a detrital peat, due to the lack of *turfa herbacea* and predominance of *detritus herbosus*. In addition, the very strong planktonic diatom component and the pollen assemblage being dominated by Gramineae with low frequencies of Cyperaceae, Chenopodiaceae, wetland herbs and aquatics, indicate that this peat unit may have been eroded and then transported to this site. Alternatively, the pollen assemblage may simply be that of an upper saltmarsh. Marine dominance persisted into the base of unit 6 with the deposition of a peaty clayey-silt. However, the marine influence declined significantly in the upper part of unit 6 into LDAZ G60 - IV, in which the environment became dominated by fresh-brackish to brackish-marine conditions characterised by the *Navicula digitoradiata var. minima* group. Here the optimal diatom group is strongly brackish water (46.7% TV). The micro-palaeoenvironmental data, therefore, record a regressive contact.

Throughout unit 6, a gradual increase in Chenopodiaceae is observed. A saltmarsh probably developed nearby, with Gramineae the dominant taxon, similar to the environment described for unit 2 below. The diatom assemblage indicates that deposition occurred on an upper mudflat to saltmarsh in unit 6 (Denys, 1994). However, into LPAZ G60 - VI at the top of this unit, a reduction in the frequencies of Chenopodiaceae occurs in combination with a significant increase in *Typha angustifolia*, which is characteristic of the transition from freshwater reedswamp to upper saltmarsh (Long and Innes, 1995b), and also Cyperaceae. The fresh-brackish to brackish-marine influence increases upward as unit 7 was deposited. Here too, high frequencies of Chenopodiaceae are recorded along with Gramineae, Cyperaceae and aquatic pollen, suggesting that a shallow pond again developed similar to that described for the biogenic sediments of unit 3, with saltmarsh and sedge fen conditions present at the margin. The vegetational succession from saltmarsh and reedswamp to a shallow freshwater pond is again characteristic of that described by Long and Hughes (1995) from Wickmaryholme Pit. Importantly, the geomorphological situations of core G60 and Wickmaryholme Pit are similar. The return of the conditions that allowed peat to develop at the site was probably caused by the emplacement of a gravel barrier across the north-eastern end of the Scotney Marsh trough, probably recorded in the stratigraphy of the site by the gravels in units 4b-d. The emplacing of this gravel barrier led to the Scotney Marsh trough, in which core G60 is situated, becoming semi-enclosed, and enabled vegetation to colonise the gravel hollow once again.

Peat deposition came to an end with the return of marine dominance and the deposition of unit 8, with the contact of units 7 and 8 being a transgressive contact. This was a predominantly marine environment with *Paralia sulcata* dominating throughout. The return to marine dominance was probably facilitated by a continued gradual rise in sea-level leading to the eventual over-topping of the protective gravel barrier. This led to the deposition of unit 8, in which planktonic diatoms are present at >80% TV. This is common in tidal channels and inlets (Vos and de Wolf, 1993), similar to the trough in which core G60 is situated. The assemblage indicates an environment of deposition was exposed to the open sea, although deposition tended to be close to the tidal limit as very poorly-sorted fine to very fine silts. settled out of suspension during

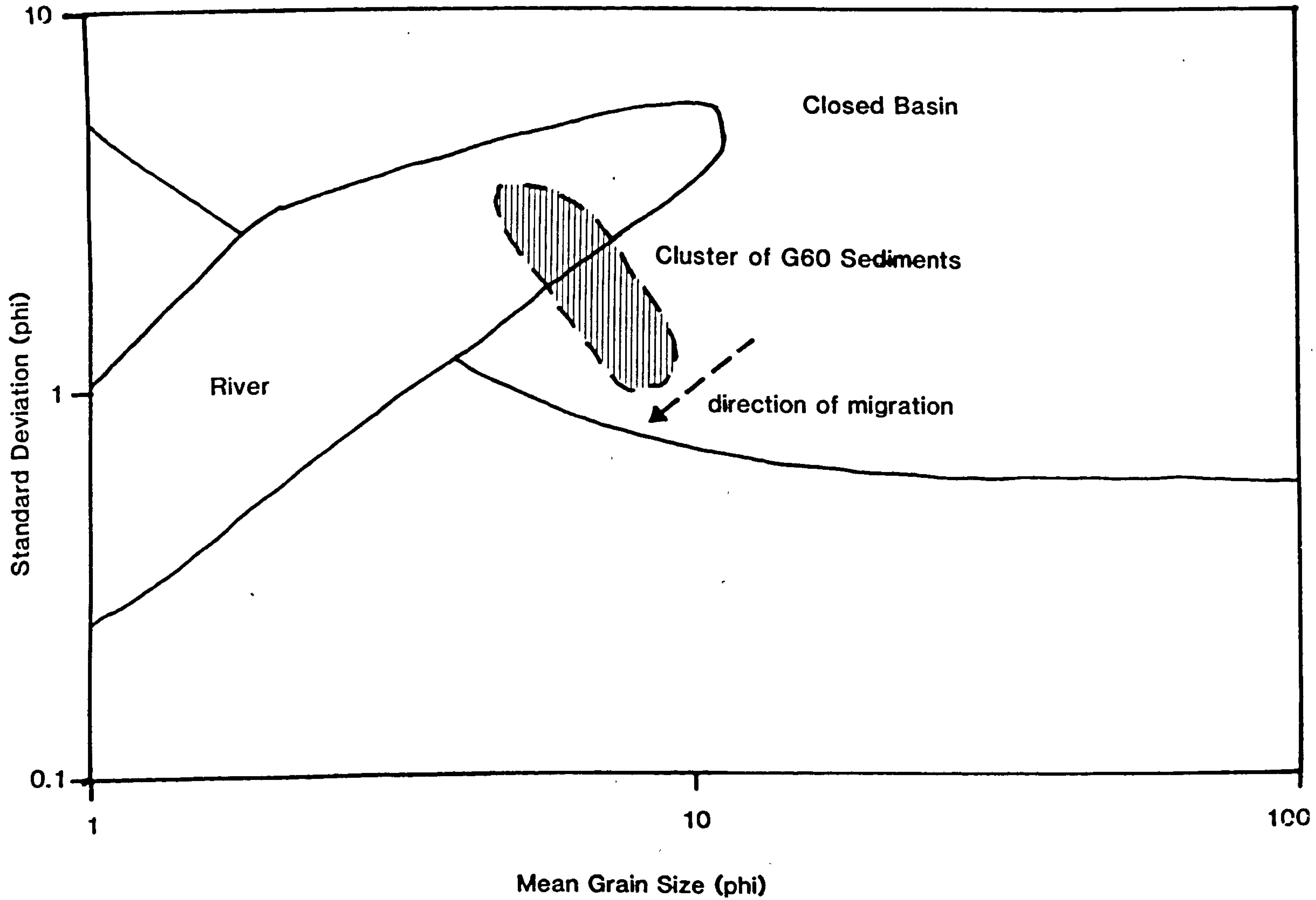

Figure 5.2: Environment of deposition, core G60, utilising mean grain size and standard deviation, developed by Tanner (1991).

slack water (Postma, 1967) within the tidal trough.

Throughout the deposition of unit 8, subtle variations in the particle distributions of the sediments indicate that slight variation in the tidal flow velocity and depth of the water occurred in the Scotney Marsh trough, although no clear evidence of direct wave action or reworking has been identified. Upward, a gradual increase in the mean grain size is recorded and the sediments also change from near-symmetrical particle size distributions to predominantly fine-skewed. This indicates that the sediments were deposited in an environment of progressively higher energy, but with the continued deposition of fines at high water. Using the model of Tanner (1991) to further interpret the particle size data, figure 5.2 illustrates that the sediments exhibit a similar trend to those of core AY17, in that they cluster across the boundary between the closed basin and river envelopes.

Table 5.6 : Core G60 Summary

Altitude m (OD)	Sediment Type / Particle Size Analysis	Predominant Diatoms	Predominant Pollen	Environment of Deposition	^{14}C Age (Cal. yrs. BP, 2 sigma)
+0.62 to -1.84	Unit 8: Grey silty clay with sandy laminations, organic rich at base. (Very poorly sorted fine to very fine silts).	Marine diatoms dominate, especially *Paralia sulcata*.	N/S	Tidal inlet with deposition of very poorly sorted fine to very fine silts under tidal conditions.	N/S
-1.84 to -1.86	Unit 7: Brown well humified peat.	Fresh-brackish to brackish-marine diatoms dominate.	Gramineae, Chenopodiaceae, Cyperaceae, and aquatic pollen.	Coastal reedswamp.	N/S
-1.86 to -1.98	Unit 6: Browny-grey, organic rich peaty silty-clay.	Initially, marine diatoms dominate, fresh-brackish to brackish marine conditions prevail upward.	Gramineae, Chenopodiaceae increase upward. At the upper contact, a reduction in Chenopodiaceae and rise in aquatic pollen.	Intertidal saltmarsh becoming a freshwater-dominated reedswamp.	
-1.98 to -2.01	Unit 5: Brown well humified peat.	Marine diatoms dominate.	Gramineae with Cyperaceae, Chenopodiaceae and aquatic pollen types.	Probably a detrital peat deposited under marine conditions.	N/S
-2.01 to -2.03 -2.03 to -2.07 -2.07 to -2.11 -2.11 to -2.15	Unit 4d: Blue-grey silt matrix with gravel. (Very well sorted coarse to very fine silts, with inputs of gravel). Unit 4c: Gravel. Unit 4b: Grey silty-clay with gravel. Unit 4a: Blue-grey clayey-silt.	High frequencies of marine diatoms, especially *Paralia sulcata*.	N/S	Tidal inlet with sediments deposited during slack water and periods of high energy depositing gravels.	N/S
-2.15 to -2.29	Unit 3: Brown well humified peat.	At the upper contact, marine and brackish diatoms dominate especially the *Melosira sulcata* and *Navicula digitoradiata var. minima* groups.	Cyperaceae, Gramineae and wetland herb pollen types dominate. An increase in aquatic pollen occurs upward and Chenopodiaceae are high at the top.	Initially, a wet sedge fen with standing pools prevailed. This became a saltmarsh / coastal reedswamp as the marine influence increased toward the top.	3375 to 3070 3370 to 2970
-2.29 to -2.32	Unit 2: Browny-grey organic peaty-silt.	Initially sparse, marine dominant with some brackish and fresh, presence of *Pinnularia viridis*.	Gramineae, Cyperaceae, Chenopodiaceae and aquatic pollen types dominate.	Transition upward from tidal mudflat to an intertidal saltmarsh.	N/S
Below -2.32	Unit 1: Gravel.	N/S	N/S	N/S	N/S

5.5.1 Core AW63 : Lithostratigraphy.

Table 5.7

Unit	Description	Altitude (OD)
7	Orange, oxidation mottled silty-sand.	+1.28 - +0.94m
6	Grey-brown peaty-clay.	+0.94 - +0.77m
5	Dark brown well humified peat.	+0.77 - +0.57m
4	Blue-grey silty-clay.	+0.57 - +0.17m
3	Brown clayey-peat, clay rich at top.	+0.17 - +0.05m
2	Grey silty-clay, organic rich at top.	+0.05 - -0.32m
1	Gravel (not recovered).	Below -0.32m

5.5.2 Core AW63: Palaeoenvironmental Interpretation

Following the deposition of the basal gravel (unit 1), a grey silty-clay (unit 2) was laid down. The depositional environment preserved in the upper part of unit 2 (at the base of LDAZ AW63-I) is initially dominated by marine-brackish epiphytic/benthic diatoms, especially *Diploneis didyma*, with significant frequencies of brackish epiphytic/benthic and marine planktonic diatoms. Thus, the assemblage is made up of the *Melosira sulcata* and *Navicula digitoradiata var. minima* groups. The environment is, therefore, suggested to have been an intertidal mudflat open to the sea, due to the high ratios of marine allochthonous diatoms along with relatively high percentages of planktonic diatoms (Vos and de Wolf, 1988), see table 5.8 for summary. The assemblage is characteristic of an upper mudflat (Denys, 1994).

At the contact between units 2 and 3, the pollen assemblage indicates that a reedswamp, dominated by Gramineae, existed, with the presence of Chenopodiaceae, *Plantago maritima* and *Plantago coronopus* indicative of nearby saltmarsh. A considerable reduction in the marine influence is represented by the transition into unit 3 where fresh to brackish conditions dominate the assemblage in LDAZ AW63 - II. Here, *Synedra pulchella* and *Synedra ulna* dominate initially, becoming replaced by *Eunotia monodon* and *Navicula peregrina*. Thus, it would appear that the environment became cut off from the direct influence of the sea, as illustrated by the significant reduction in the ratios of allochthonous marine diatoms above +0.05m OD. This regressive contact has been dated between altitudes of +0.05 and +0.12m OD to 3715-3565 cal. yrs. BP.

The regressive event may have been the result of gradual gravel barrier emplacement which would have prevented the sea from gaining access to the gravel inlet, allowing the inlet to have become dominated by fresh to brackish water. Equally, regression may have been due to sedimentation leading to the infilling of the gravel inlet and eventual colonisation of vegetation on the subaerial mudflat and, thus, the accumulation of the organic sediments. Both hypotheses are supported by the pollen data from LPAZ AW63 - II, which are dominated by Gramineae with significant frequencies of *Typha angustifolia*. These are indicative of deposition at the transition from freshwater reedswamp to upper saltmarsh (Long and Innes, 1995b). The frequencies of Chenopodiaceae gradually increase upward, indicating that, at this time, a saltmarsh developed nearby and its influence increased upward through unit 3. Also, the proximity of a sedge fen is suggested by the presence of Cyperaceae and wetland herb types.

Thus, it would appear that a coastal reedswamp, as described by Waller (1994b), developed after the infilling of the gravel inlet by sedimentation at a rate faster than sea-level was rising. Fresh to brackish water conditions then prevailed and vegetation developed in the inlet with saltmarsh and sedge fen at the margins. The presence of saltmarsh throughout the peat unit indicates that some marine influence remained. Thus, the gravel inlet was not enclosed from the sea *via* the emplacement of a gravel barrier unless such a barrier was overtopped or rounded at each high tide.

At the top of unit 3, the marine influence appears to have increased considerably as the environment became dominated by brackish-fresh and brackish epiphytic/benthic diatoms with a significant allochthonous marine planktonic contribution. This demonstrates that the sea-level appears to have continued rising throughout the deposition of the peat of unit 3 and that at the contact of units 3 and 4, the rate of sea-level rise became faster than the rate of peat accumulation causing the peat surface to be inundated.

In unit 4, brackish water conditions prevailed with brackish epiphytic/benthic diatoms, particularly *Nitzschia navicularis*, being dominant and increasing upward in LDAZ AW63 - III. The *Navicula digitoradiata var. minima* group is dominant with a small number of the *Melosira sulcata* group. It appears from the low ratios of allochthonous marine diatoms and predominance of a brackish optimal diatom group, that the marine inputs to the site were reduced. The environment became more cut off from the open sea upward as brackish autochthonous diatoms increase relative to a marine allochthonous component. Consequently, an intertidal mudflat existed, with early stage brackish to brackish-marine water becoming brackish as the marine influence was removed further. Again, this was possibly due to the gradual emplacement of a gravel barrier allowing deposition to occur on an upper mudflat to lower saltmarsh (Denys, 1994) or, more likely, a low rate of sea-level leading to the sediments becoming sub-aerially exposed and allowing vegetation to colonise the intertidal mudflats, as suggested by Tooley and Switsur (1988).

Into unit 5, organic deposition again replaced minerogenic sedimentation as peat developed. At the base of this peat, in LPAZ AW63 - III at the contact between units 4 and 5, the pollen assemblage is dominated by Gramineae and Cyperaceae. The frequencies of Chenopodiaceae are initially high, but become reduced upward as the marine influence decreases. Therefore, at this point, the environment of deposition was that of a saltmarsh, as described by Long and Innes (1995a) for the initiation of peat at Brookland. The diatom assemblage at the contact of units 4 and 5 also indicates the presence of a saltmarsh to upper mudflat (Denys, 1994). Thus, the environment was initially characteristic of a saltmarsh, with grasses and sedges dominating as the marine influence was gradually reduced. The environment became dominated by fresh-brackish water conditions, as indicated by an optimal fresh-brackish diatom group reaching 72.4% TV.

Eventually the environment changed, with Gramineae and *Typha angustifolia* becoming dominant. This implies that a reedswamp had developed in the fresh-brackish conditions, with saltmarsh indicators still present as traces in LPAZ AW63 - IV, thus, indicating a transition from upper saltmarsh to freshwater reedswamp (Long and Innes, 1995b). However, towards the top of unit 5, in LPAZ AW63 - V, a change in environment is recorded as the freshwater table fell with the removal of marine conditions. This is suggested from the reduction in the frequencies of aquatic pollen types. A rise in frequency is recorded for Cyperaceae, Filicales, and also *Alnus* and the wetland pollen types, to the top of unit 5. This is similar to the environment described by Long and Innes (1993) at Midley, where a sedge fen with standing pools is considered to have developed. However, conditions were dry enough to support *Alnus* nearby. The environment at this point was still dominated by fresh-brackish water. The taxon *Epithemia turgida* indicates that freshwater pools were in existence on the peat surface at the top of unit 5 in LDAZ AW63 - IV, (Vos and de Wolf, 1988). The whole of this peat unit has been dated between altitudes of +0.59 and +0.74m OD to 3265-2930 cal. yrs. BP.

In unit 6, the fresh-brackish environment persisted, with the *Epithemia zebra* group continuing to dominate. The presence of *Diploneis interrupta* and *Pinnularia viridis* (Vos and de Wolf, 1988) and the relatively high frequency of Chenopodiaceae pollen, indicate that saltmarsh conditions prevailed nearby at this time. The high frequencies of Gramineae and aquatic pollen types, and a corresponding fall in the frequencies of Cyperaceae, indicate an increase in the waterlogging at the site. A fall in the dominance of the fresh-brackish optimal diatom group and a coincident increase in the ratio of marine allochthonous diatoms suggest that an increase in the marine influence was experienced, which caused a reversal in the vegetational succession. This is also indicated by an increase in the brackish influence at the top of LDAZ AW63 - IV, although fresh-brackish epiphytic/benthic diatoms are still the predominant types. The environment of deposition recorded in the upper contact of the peaty-clay (unit 6) was, therefore, a coastal reedswamp, as described by Waller (1994b), at the transition between marine and freshwater dominance.

It appears that marine conditions returned with the deposition of a peaty-clay (unit 6) and then a silty-sand (unit 7). This was facilitated by either the removal of a gravel barrier, allowing marine access to the site, or more likely due to the rate of sea-level rise overtaking the rate of peat accumulation. The return of marine conditions to the site was probably facilitated by the marine flooding of tidal channels to the north of Scotney Marsh, *i.e.* the Wainway Channel. The latter explanation is more likely due to the fact that the deposition of the orange oxidation mottled silts and sands (unit 7 in core AW63) was widespread across the Scotney Marsh area north of ridge 1. In addition, the Wainway Channel would, if sea-level had continued to rise throughout the period of peat development, be susceptible to flooding by the higher sea-level, thus, inundating the marsh areas in the immediate vicinity. These marine channels would have been flooding earlier, but until sea-level rose above a certain threshold, these floods would not have reached sufficient height to inundate the marsh.

Table 5.8 : Core AW63

Altitude m (OD)	Sediment Type	Predominant Diatoms	Predominant Pollen	Environment of Deposition	^{14}C Age (Cal. yrs. BP, 2 sigma)
+1.28 to +0.94	Unit 7: Orange, oxidation mottled silty-sand.	N/S	N/S	N/S	N/S
+0.94 to +0.77	Unit 6: Grey-brown peaty-clay.	Fresh-brackish to brackish diatoms dominated, especially *Epithemia turgida*.	Gramineae, Chenopodiaceae and some subordinate aquatic pollen types dominate.	Saltmarsh to coastal reedswamp conditions prevailed.	N/S
+0.77 to +0.57	Unit 5: Dark brown well humified peat.	At the upper contact the optimal diatom group, fresh-brackish diatoms dominated, especially *Epithemia turgida*.	At the base, Gramineae, Cyperaceae and Chenopodiaceae dominate. Upward, Sparganium and Gramineae became dominant. At the upper contact, Cyperaceae and the wetland pollen types dominate.	A saltmarsh to coastal reedswamp was superceded by a freshwater reedswamp as the marine influence was reduced. As the environment dried out, a sedge fen with some open pools prevailed.	3265 to 2930
+0.57 to +0.17	Unit 4: Blue-grey silty clay.	Brackish diatoms dominate, especially *Nitzschia navicularis*. The optimal diatom group is brackish throughout.	N/S	The rate of sea-level rise overtook the rate of peat accumulation allowing minerogenic sediments to be deposited on an intertidal mudflat.	N/S
+0.17 to +0.05	Unit 3: Brown clayey-peat, clay rich at upper contact.	At the base, fresh to brackish diatoms dominate, especially *Synedra pulchella* and *Synedra ulna*. Brackish diatoms increase upwards towards the middle and dominate at the upper contact with subordinate marine diatoms.	Gramineae and Cyperaceae with some wetland herb types. Sparganium increase from initially low levels upward. Chenopodiaceae are present at the lower contact but are reduced upward before increasing toward the upper contact.	The mudflat became subaerially exposed allowing a coastal to freshwater reedswamp to prevail with sedge fen and saltmarsh at the margins	3715 to 3565
+0.05 to -0.32	Unit 2: Grey silty-clay, organic rich at upper contact.	Brackish diatoms dominate, especially the *Navicula digitoradiata var. minima* group with subordinate *Melosira sulcata* group.	N/S	Intertidal mudflat with inputs from the open sea.	N/S
Below -0.32	Unit 1: Gravel.	N/S	N/S	N/S	N/S

5.6.1 Core AW-AX67, Lithostratigraphy

Table 5.9

Unit	Description	Altitude (OD)
8	Orange, oxidation mottled silty-sand.	+1.19 - +1.09m
7	Brown silty-clay with substantial organics.	+1.09 - +1.05m
6	Grey clay with silt.	+1.05 - +0.95m
5	Brown well humified peat.	+0.95 - +0.76m
4	Blue-grey silty-clay.	+0.76 - +0.21m
3	Brown well humified peat with some clay.	+0.21 - +0.16m
2	Blue-grey silty-clay.	+0.16 - -0.49m
1	Gravel (not recovered).	Below -0.49m

5.6.2 Core AW-AX67 : Palaeoenvironmental Interpretation

Overlying the basal gravel (unit 1), a blue-grey silty-clay (unit 2) was deposited. The deposition of the upper part of unit 2 occurred in an environment dominated by marine to brackish conditions, as indicated by high frequencies of the *Melosira sulcata* and *Navicula digitoradiata var. minima* groups. The large contribution made by planktonic diatoms and also allochthonous fresh-water and marine diatoms, make accurate palaeoenvironmental reconstruction problematic. However, it would appear that the site was an intertidal to subtidal mudflat, see table 5.10. Significantly, the back-barrier site must still have been exposed to the open sea as the frequencies of marine epiphytic/benthic diatoms are relatively high. Therefore, unit 2 of core AW-AX67 was deposited in the same environment as that of unit 2 in core AW63.

Upward through LDAZ AWAX67 - I, into unit 3, the marine influence decreases with a corresponding rise in the frequencies of fresh-brackish to brackish diatoms, *i.e.* the

Navicula digitoradiata var. minima group is still dominant and the optimal diatom group is brackish with a significant marine allochthonous component. At the contact between units 2 and 3, a saltmarsh environment was in existence, as indicated by the high frequencies of Chenopodiaceae and the dominance of Gramineae in LPAZ AWAX67 - I. An upper mudflat to lower saltmarsh environment is indicated by the diatom assemblage of LDAZ AWAX67 - I (Denys, 1994). However, through unit 3, a significant increase in the frequencies of aquatic pollen is observed in LPAZ AWAX67 - II, with a corresponding gradual decrease in the frequencies of Chenopodiaceae. This also demonstrates that the marine influence was reduced through unit 3, and that fresh-brackish to brackish water conditions prevailed with the accompanying colonisation by aquatic plants. Grasses, sedges and wetland herbs would also have been present at the margins. Long and Hughes (1995) described a similar vegetation assemblage at Wickmaryholme Pit as a brackish to freshwater shallow pond. At this site, as in core AW63 the accumulation of this peat unit (unit 3 in both cores) was facilitated by sedimentation at a more rapid rate than the rate of sea-level rise, thus, leading to the sub-aerial emergence of the sediments and colonisation of the substrate by vegetation. However, a significant marine influence still existed throughout the period of deposition of unit 3, as the vegetation was probably inundated at each high tide. The entire peat (unit 3) was dated between altitudes of +0.16 and +0.21m OD to 4060-4040 or 3990-3700 cal. yrs. BP.

At the contact between units 3 and 4, an increase in the marine influence appears to have occurred, as suggested by a reduction in the frequencies of aquatic taxa and a corresponding rise in frequencies of Chenopodiaceae and Gramineae in LPAZ AWAX67 - III. The contact between units 3 and 4 demonstrates a strong similarity to the contact between the corresponding units 3 and 4 in core AW63.

The above trend is further illustrated throughout unit 4, where the marine influence increases significantly upward and there is a move from a brackish to a more marine optimal diatom group. The frequencies of marine planktonic diatoms fall slightly before rising to the top of the LDAZ AWAX67 - II, with planktonic diatoms dominating the assemblage. Significant contributions are also made by brackish to marine-brackish diatoms throughout the zone. The *Melosira sulcata* group is dominant in unit 4, with subordinate frequencies of the *Navicula digitoradiata var. minima* group. Deposition of this unit probably occurred on an intertidal mudflat (Denys, 1994).

Thus, following the deposition of the biogenic unit 3, a return to marine dominance occurred, during which the minerogenic unit 4 was deposited. The return must have been relatively gradual as the vegetational succession at the top of unit 3 had sufficient time to go into a reversal, with a saltmarsh present at the contact. The cause of this was probably an increase in the rate of sea-level rise, which exceeded the rate of peat accumulation and, therefore, enabled inundation of the peat and deposition of an intertidal mudflat. However, at the contact of units 4 and 5, a relatively dramatic return to more fresh to brackish conditions is recorded at the base of LDAZ AWAX67 - III, with a corresponding reduction in frequencies of marine diatoms to <10% TV. The optimal diatom group is brackish, although a greater number of fresh allochthonous diatoms are present than those in the optimal diatom group. At the base of this unit (unit 5), the *Navicula digitoradiata var. minima* and *Diploneis interrupta* groups dominate, indicating that deposition was on an upper mudflat to saltmarsh (Vos and de Wolf, 1988). The return of these conditions indicates that the mudflat again became sub-aerially exposed due to the infilling of the gravel inlet by sediments and subsequent colonisation by salt-tolerant vegetation began to colonise the mudflat. The diatom assemblage indicates that deposition occurred on an upper mudflat to lower saltmarsh (Denys, 1994).

A less clear picture is evident from the pollen data, in which Gramineae and Cyperaceae dominate with low frequencies of saltmarsh, aquatic and wetland herb indicators. The assemblage suggests that a coastal reedswamp environment existed at the site, with high (up to 75% TLP) frequencies of Gramineae. The other taxa indicate the presence of saltmarsh and wet sedge fen nearby. The lower part of the peat (unit 6) forms a regressive contact which has been dated between altitudes of +0.76 and +0.85m OD to 3220-2795 cal. yrs. BP.

A very high peak of *Typha angustifolia* (~105% TLP) occurs in the middle of the LPAZ AWAX67 - IV. This is probably due to the inclusion of a macrofossil of this plant in the sample, but could also be due to the existence of a freshwater pool prevailing at the site at this time. Towards the top of LPAZ AWAX67 - IV, the assemblage becomes indicative of a saltmarsh, with an increase in the frequencies of Chenopodiaceae and *Plantago maritima* present as a trace. This pollen assemblage suggests the presence of a coastal reedswamp, as described by Waller (1994b), in unit 5 at the transition from marine and freshwater conditions, becoming more marine towards the top. The presence of a *Phragmites* saltmarsh environment (Denys, 1994) at this time is also indicated by the high frequencies (~75% TV) of *Diploneis interrupta* at the top of LDAZ AWAX67 - III, in which there is a brackish optimal diatom group with a very low contribution from either fresh-water or marine allochthonous diatoms. The upper contact of the peat (unit 6) forms a transgressive contact which has been dated between altitudes of +0.85 - +0.94m OD to 2950-2725 cal. yrs. BP.

The saltmarsh appears to have persisted into unit 6, with the *Diploneis interrupta* group still dominant. It would appear that a slight increase in the marine influence occurred, as indicated by increases in the frequencies of the *Melosira sulcata* group. This resulted in an increase in the numbers of allochthonous planktonic marine diatoms. This trend is continued into unit 7, where a considerable decrease in the frequencies of brackish epiphytic/benthic diatoms is observed with a coincident rise in the frequencies of marine-brackish epiphytic/benthic diatoms, which eventually become the optimal diatom group. A return to dominance of the *Navicula digitoradiata var. minima* group indicates a return to an intertidal mudflat environment with small but significant inputs from the open sea influencing deposition on a upper mudflat to lower saltmarsh (Denys, 1994).

Therefore, an increase in the marine influence is recorded by the brackish to marine-brackish dominance in unit 7. This was probably caused by a gradual sea-level rise with the marine water, probably flooding the tidal channels, such as the Wainway Channel to the north of Scotney Marsh, to a greater extent than before.

Table 5.10 : Core AW-AX67

Altitude m (OD)	Sediment Type	Predominant Diatoms	Predominant Pollen	Environment of Deposition	^{14}C Age (Cal. yrs. BP, 2 sigma)
+1.19 to +1.09	Unit 8: Orange, oxidation mottled silty-sand.	N/S	N/S	N/S	N/S
+1.09 to +1.05	Unit 7: Brown silty-clay with substantial organics.	Brackish to marine-brackish diatoms dominate, especially the *Navicula digitoradiata var. minima* group.	N/S	Intertidal mudflats. A return to the intertidal back-barrier environment with some open sea input.	N/S
+1.05 to +0.95	Unit 6: Grey clay with silt.	Brackish diatoms dominate, especially *Diploneis interupta*, with subordinate marine diatoms.	N/S	Initially, saltmarsh conditions prevailed which became intertidal mudflats upward.	N/S
+0.95 to +0.76	Unit 5: Brown well humified peat.	Fresh to brackish diatoms dominate, with contributions from the *Navicula digitoradiata var. minima* and *Diploneis interupta* groups. The optimal diatom group is brackish.	Gramineae and Cyperaceae dominate. A high value of Sparganium is recorded and the frequency of Chenopodiaceae gradually rises towards the upper contact.	Initially, intertidal mudflats existed which became more freshwater dominated and a coastal reedswamp developed. Saltmarsh conditions prevailed at the upper contact.	2950 to 2725 3220 to 2795
+0.76 to +0.21	Unit 4: Blue-grey silty-clay.	Marine planktonic diatoms dominate, especially the *Melosira sulcata* group with subordinate *Navicula digitoradiata var. minima* group.	At the base, Gramineae and Chenopodiaceae dominate with subordinate Cyperaceae and aquatic pollen types.	At the lower contact, a saltmarsh / coastal reedswamp existed, which became an intertidal mudflat upward.	N/S
+0.21 to +0.16	Unit 3: Brown well humified peat with some clay.	Fresh-brackish to brackish diatoms dominate, especially the *Navicula digitoradiata var. minima* group. Some allochthonous marine input exists throughout.	Initially Chenopodiaceae and Gramineae dominate. Gramineae, Cyperaceae and aquatic pollen types become dominant upward, with some Chenopodiaceae throughout.	A saltmarsh was replaced upward by a fresh to brackish coastal reedswamp colonised by aquatic plants.	4060 to 4040 or 3990 to 3700
+0.16 to +0.09	Unit 2: Blue-grey silty-clay.	Brackish diatoms are the optimal diatom group, with a high allochthonous marine input. *Melosira sulcata* and *Navicula digitoradiata var. minima* groups dominated.	N/S	Intertidal to subtidal marine-brackish mudflat exposed to the influence of the open sea.	N/S
Below +0.09	Unit 1: Gravel.	N/S	N/S	N/S	N/S

5.7 Palaeoenvironmental Reconstruction of Scotney Marsh.

To reconstruct the palaeoenvironment of Scotney Marsh, first the gravel deposits will be considered as these deposits form the surface upon which the marsh sediments were deposited. The evolution of each of the environments will then be studied in detail, *via* consideration of the typecores, the main back-barrier environment (cores AY17 and A-B27), the Scotney Marsh trough (core G60) and the gravel inlet (cores AW63 and AW-AX67).

5.7.1 The Evolution of the Gravel.

The basal gravel unit is present across Scotney Marsh at some depth in each of the cores, and is also at the surface as two major ridges that run approximately south-west to north-east through the area. These ridges are at the north-western margin of the study area, and are, therefore, one of the earliest of the groups of ridges that make up the Dungeness Foreland, *i.e.* the Midrips, The Forelands and Holmestone (Eddison, 1983a). The gravel of ridge 2, which marks the south-eastern border of Scotney Marsh, is at the north-eastern end of The Forelands gravel complex (figure 5.3). The remainder of the sediments of Scotney Marsh occur between the complex series of near parallel ridges this includes the area between shorelines 3 and 4 of Lewis (1932) (figure 5.4). The dating for the emplacement of these gravel ridges and swales of the study area is, thus, a matter of some debate, yet this is critical in determining the evolution of Scotney Marsh.

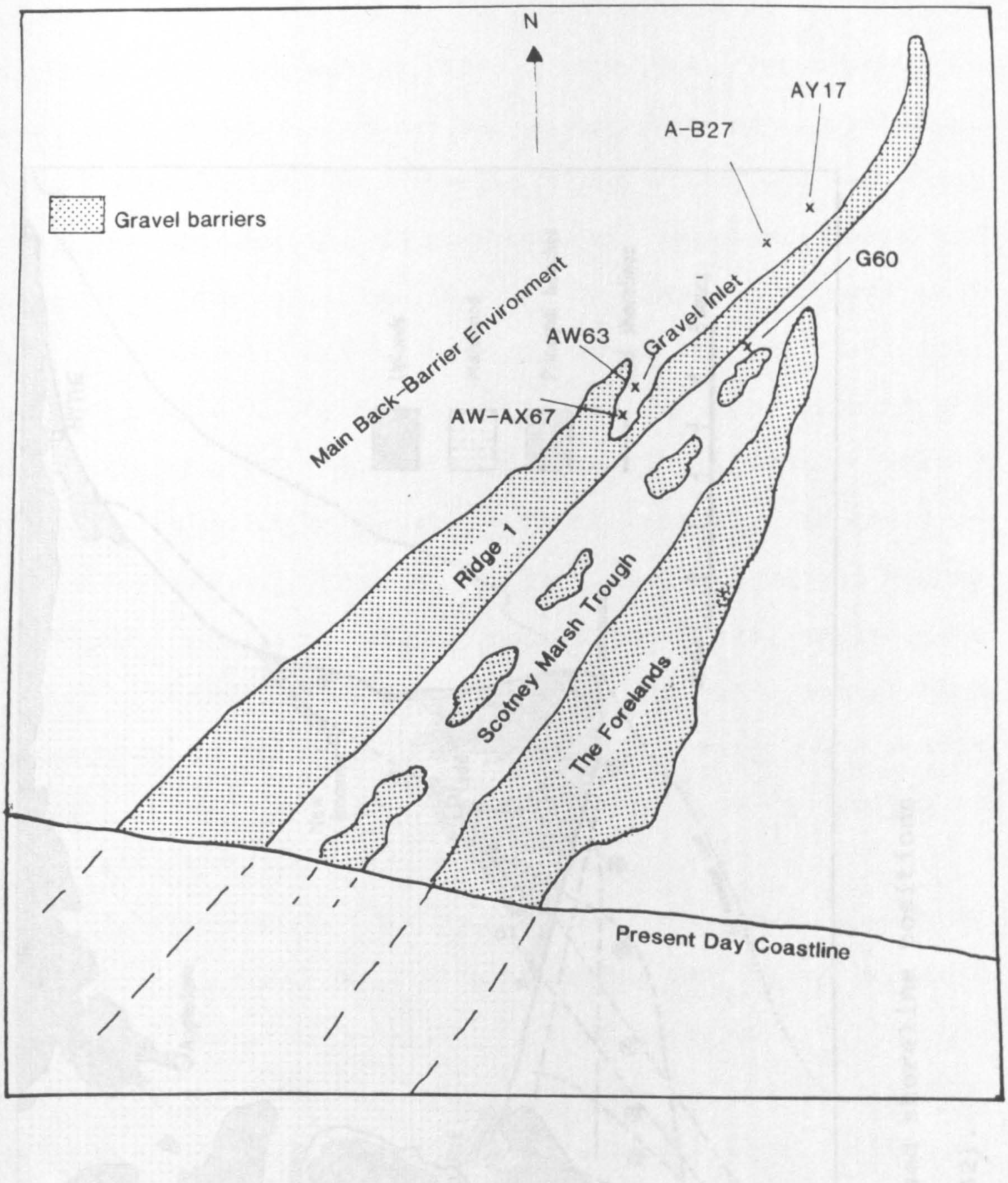

Figure 5.3: The locations of the important features in the Scotney Marsh area.

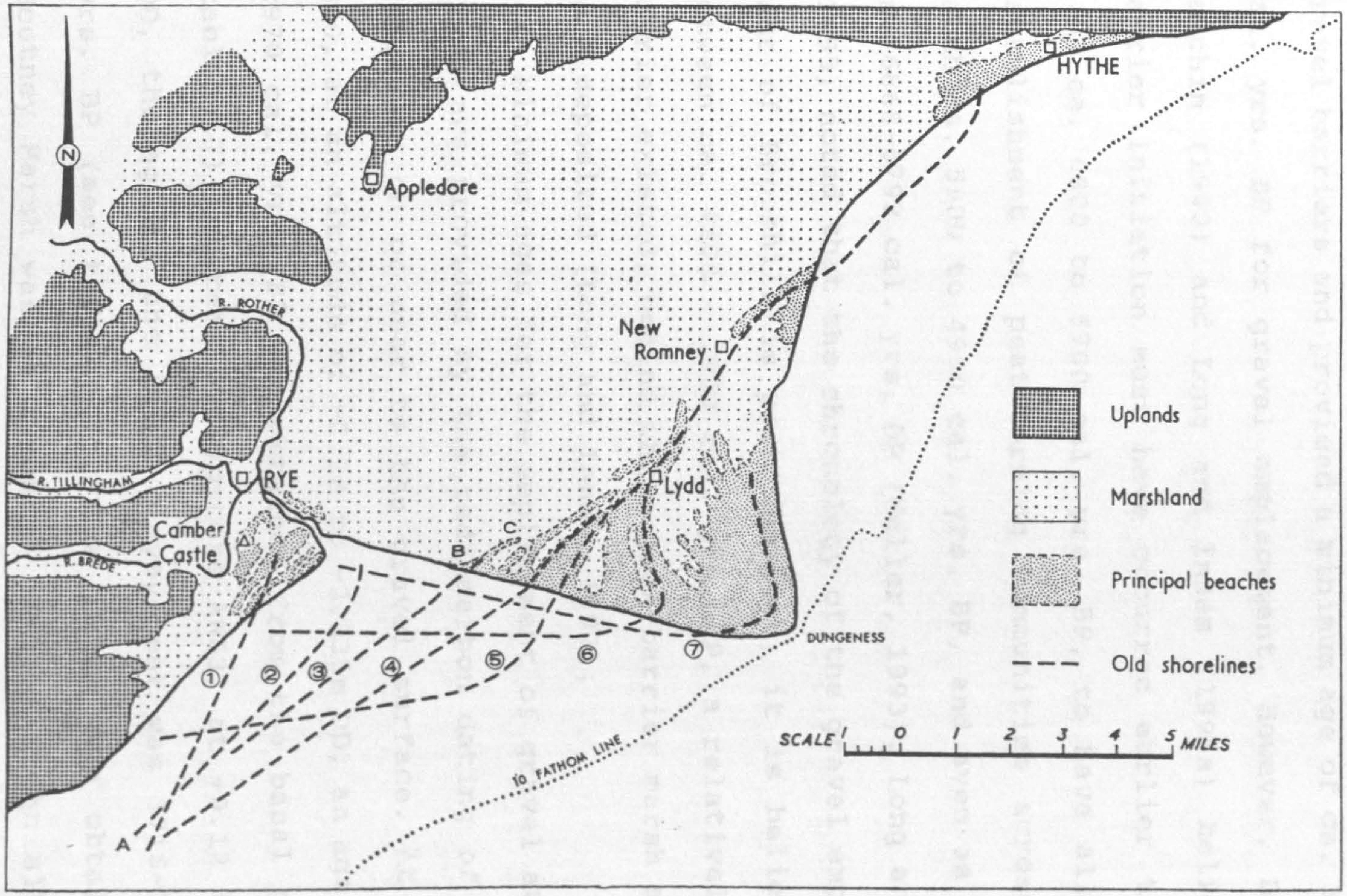

Figure 5.4: Proposed shoreline positions.

Source: Lewis (1932).

The dating of barrier emplacement has been attempted elsewhere from the back-barrier sedimentary record, particularly at Broomhill to the south-west of Scotney Marsh (Tooley and Switsur, 1988). Here, too, investigations were carried out in the swales between some of the earliest gravel barriers and provided a minimum age of ca. 3382-3345 cal. yrs. BP for gravel emplacement. However, Lewis and Balchin (1940) and Long and Innes (1995a) believe that barrier initiation must have occurred earlier than this *i.e.* ca. 6800 to 5700 cal. yrs. BP, to have allowed the establishment of peat-forming communities across Romney Marsh ca. 5900 to 4900 cal. yrs. BP, and even as early as ca. 6868-6792 cal. yrs. BP (Waller, 1993). Long and Hughes (1995) noted that the chronology of the gravel emplacement east of Broomhill is poor. However, it is believed that between ca. 4500 - 1950 cal. yrs. BP, a relatively stable barrier existed, behind which back-barrier marsh sediments were deposited (Long and Innes, 1993).

Minimum ages for the emplacement of gravel at Scotney Marsh are provided by the radiocarbon dating of biogenic sediments on or near to the gravel surface. At typecore G60, at an altitude of -2.28 to -2.235m OD, an age of 3370-2970 cal. yrs. BP was obtained from the basal peat (see table 4.7), whereas at typecore AW63, at +0.12 to +0.05m OD, the age on the lower clayey-peat was 3715-3565 cal. yrs. BP (see table 4.7). The oldest age obtained from Scotney Marsh was at typecore AW-AX67 at an altitude of +0.16 to +0.21m OD on the lower peat, giving an age of

4060-4040 or 3990-3700 cal. yrs. BP (see table 4.7). Therefore, a minimum age for the emplacement of ridge 1 in Scotney Marsh is ca. 4000 cal. yrs. BP, with a minimum age for the emplacement of ridge 2 of ca. 3150 cal. yrs. BP.

Due to the lack of stratigraphic detail from within the gravel unit, it is not possible to provide any detailed interpretation of its environment of deposition. However, the wealth of previous work on the gravel deposits of the Romney Marsh coast (Lewis, 1932; Lewis and Balchin, 1940; Eddison, 1983a Long and Fox, 1988) provide some insight to the environment of deposition (see section 2.1.2) Lewis and Balchin (1940) have proposed that the alternating areas of gravel and marsh, such as those recorded within Scotney Marsh, record periods of high sea-level, during which the gravel ridge develops, and then relatively low sea-level, during which the swale develops. However, Long and Innes (1993) and Plater and Long (1995) believe that it is hard to establish a relationship between the gravel barriers and sea-level changes and, thus, it is difficult to resolve whether long term sea-level change or the magnitude and frequency of storms control the gravel movement and deposition. Importantly, Long and Hughes (1995) suggested that the alternating areas of gravel and marsh did not record periods of sea-level change, but rather, episodic changes in gravel supply, possibly linked to changes in storm incidence.

The gravel surface morphology in Scotney Marsh does, however, provide some detail on the depositional history of the gravel. For example, the close interval borings reveal that the ridges tend to be asymmetrical, with the shallow lee-side to the north-western and the steeper stoss-side to the south-east. This suggests that the gravel ridges were deposited from the south-east, with the stoss-side being steeper due to the constant wave action (Lewis and Balchin, 1940). Successive ridges to the south-east become progressively younger, therefore, ridge 1 is likely to be older than ridge 2, and the smaller gravel ridges Ga, Gb and Gc were deposited in the period between the two.

5.7.2 Sediments Overlying the Gravel in Scotney Marsh.

Following the deposition of the gravel ridges, it is not clear how much time elapsed before the deposition of the overlying minerogenic sands, silts, clays and peats that occur across Scotney Marsh where the gravel surface is sufficiently depressed. Indeed, it is not certain whether deposition in the back-barrier environments of the two ridges was synchronous. However, it is suggested here that deposition in the Scotney Marsh trough occurred sometime after deposition in the main back-barrier environment.

5.7.2.1 The Main Back-barrier Environment.

In the main back-barrier environment, the blue-grey silts which directly overlie the gravel surface were initially deposited on a tidal mudflat from fully marine water. However, the marine dominance became reduced upward. Within the sediments being deposited on tidal mudflats, variations in the salinity, depth of water and tidal velocity are recorded. The variations in the depositional environment may be accounted for by changes in the connection of the back-barrier inlet to the open sea the location of which is not known but was probably between Lydd and New Romney to the north-east.

The blue-grey silts pass transitionally upward into a peaty-clay. Through this transition, deposition was initially on an intertidal mudflat and then an upper mudflat, which became colonised by saltmarsh plants as the marine influence was reduced. This trend was continued upward as more brackish to freshwater conditions prevailed and the marine influence was further removed, allowing organic peat to accumulate on a *Phragmites* saltmarsh, followed by a freshwater reedswamp to freshwater pool. The base of this peat unit was dated in typecore AY17 to 3355-2980 cal. yrs. BP on the regressive contact at an altitude of +0.52 to +0.57m OD (see table 4.7). The same contact between the saltmarsh peaty-clay and freshwater peat in core A-B27 yielded a date of 2795-2710 or 2620-2500 cal. yrs. BP (see table 4.7).

This regressive sequence may have been facilitated by either the gradual building / extension of a gravel barrier, or more likely due to the relatively stable or slowly rising sea-level present across Romney Marsh at this time. Long and Innes (1993) suggest that sea-level was relatively stable between ca. 4500-1950 cal. yrs. BP. This would have promoted the infilling of the back-barrier environment with the blue-grey silts which would have gradually emerged from the sea allowing the vegetation to colonise, and peat deposition to commence. Alternatively, sufficiently high rates of sediment accretion may have kept pace with or exceeded the rate of sea-level.

Gradually the freshwater table fell as marine conditions were further removed, allowing a sedge fen to develop across much of the back-barrier environment, with alder carr conditions nearby and standing pools on the peat surface. Upward, towards the top of the peat unit in the back-barrier environment, a reversal of the vegetational succession occurred with the return of saltmarsh and aquatic plants indicating a rise in the local water table and an enhancement in the marine influence. It appears that a *Phragmites* saltmarsh persisted across the area, with a significant increase in the inputs of marine water from the open sea. The upper contact of this peat unit has been dated in typecore AY17 at an altitude of +0.72 to +0.77m OD giving an age of 2710-2585 or 2510-2320 cal. yrs. BP for the transgressive contact (see table 4.7).

The cause of the reversal of the peat succession is not clear, but it appears that a gradual return of more marine conditions occurred and the vegetation had sufficient time to respond. However, the oxidation mottled silt truncates the upper peat contact in many cores and, consequently, palaeoenvironmental interpretation is somewhat equivocal. It is proposed by Long and Innes (1993) that sea-level was relatively stable until ca. 1950 cal.

yrs. BP. If this is accepted then, it would be unlikely that the end of peat deposition was associated with a rise sea-level. However, upper peat contacts, both within the main back-barrier environment (cores AY17 and A-B27) and also the gravel inlet (cores AW63 and AW-AX67) record a gradual lithological change which is indicative of a rise in relative sea-level, and which may have caused the inundation of the peats. If the cause of the inundation had been the removal of a protective gravel barrier, then the return of minerogenic deposition and the end of peat accumulation would probably been relatively rapid. This was not the case in the upper parts of the peat units from Scotney Marsh. Although the peat erosion and truncation in other places suggest that this gradual transition may have disturbed the stratigraphic record in certain locations the vegetational succession illustrates that the freshwater table appears to have risen towards the end of peat accumulation with aquatic pollen dominating. This could be caused by a rising sea-level ponding back freshwater drainage. This trend is continued upward to the transgressive contact as coastal reedswamp to saltmarsh conditions are recorded. Inundation of the peat, when it did occur, may have been relatively rapid as the saltmarsh phase of the vegetation succession appears to have been short-lived. This relatively rapid inundation of the peat may have been facilitated by the rising sea-level leading to the flooding with marine to brackish water from the tidal channels, such as the Wainway Channel, to the north of Scotney Marsh. An alternative hypothesis is that a barrier was overtopped, or partially degraded, allowing marine conditions to reach the back-barrier environment. However, this would not explain the rise in the water table recorded in the reversal of the vegetational succession.

5.7.2.2 The Scotney Marsh Trough.

As has already been outlined, ridge 1 appears to have had a very strong control on the evolution of Scotney Marsh, with the sediments of the back-barrier environment being deposited, to the north-west of ridge 1 and the sediments in the Scotney Marsh trough deposited to the south-east. Therefore, it is necessary to further investigate the evolution of the Scotney Marsh trough in order to examine the differences in the palaeoenvironments of these two adjacent (~200 metres apart), morphologically and altitudinally similar features, and to propose some explanation for the observed differences.

Typecore G60 was atypical of the sediments in the Scotney Marsh trough in that it contained a sequence of biogenic sediments at an altitude of approximately -2.00m OD. However, the sediments above the peats were the same as those recorded throughout the remainder of the trough, *i.e.* blue-grey silts overlain by oxidation mottled silts.

In core G60, directly overlying the gravel surface, a peaty-clay was present. This indicated that intertidal mudflat deposition progressively gave way to an intertidal saltmarsh as plants colonised the muddy substrate. This environment of deposition then became a wet sedge fen to freshwater pool. The dating of this peat unit has given an age of 3370-2970 cal. yrs. BP (see table 4.7) for the regressive contact. Towards the top of unit 3, the marine influence increased and a reversal of the vegetational succession was recorded, with the presence of a coastal saltmarsh environment indicating a transgressive contact with the overlying minerogenic sediments, *i.e.* between units 3 and 4. An age of 3375-3070 cal. yrs. BP has been obtained for the upper part of this peat unit (see table 4.7). Above the biogenic sediments, a sequence of silts and gravels were deposited under marine-dominated conditions. This suggests that tidal deposition predominated, but with some periods of high energy able to transport the gravels. Slumping of the surrounding gravel ridges into the hollow is not likely as this would not explain the deposition of the gravels within a silt matrix. Instead, gravel input occurred possibly due to the overtopping of a gravel barrier that had previously provided the relative shelter which enabled peat deposition. It appears that after this high energy disturbance, possibly a storm or a rise in the base level for storms *via* gradual sea-level rise, the environment of deposition returned to relatively low energy sedimentation, with a detrital peat deposited above the minerogenic sediments.

The low energy environment in which the upper peat accumulated continued upward, with the deposition of a peaty-clay on an intertidal saltmarsh with a strong marine influence throughout. Upward, however, the environment became a freshwater-dominated pond with a widespread occurrence of aquatic plants. Eventually a coastal reedswamp developed, depositing the upper peat (unit 7), which then gave way to a tidal inlet in which a silty-clay was deposited.

The environment of deposition throughout the sedimentation of the lower part of core G60 was similar to that of Wickmaryholme Pit (Long and Hughes, 1995). Critically, the location of core G60 in a hollow in a gravel swale is morphologically similar to that of Wickmaryholme Pit. It would appear, therefore, that the lower sediments of core G60 were deposited in an 'open pit' (Eddison, 1983a). Long and Hughes (1995) propose that gravel entering a drift-aligned system tends to be transported down drift until wave competence is no longer able to transport it. This limit is the 'null point'. The position of the null point is dependent on a number of variables, e.g. during a period of storm conditions wave competence is high and so the null point moves further down drift. This null point extension allows distal extension of the beach ridges (figure 5.5). Conversely, during a period of low storm incidence, the gravel beaches would tend to be shorter as a consequence of null point contraction, and are built up at the proximal end of the barrier. The formation of the open pits is believed to have been facilitated when a period of relative low storm incidence, which promotes the build up of a number of relatively short ridges, is

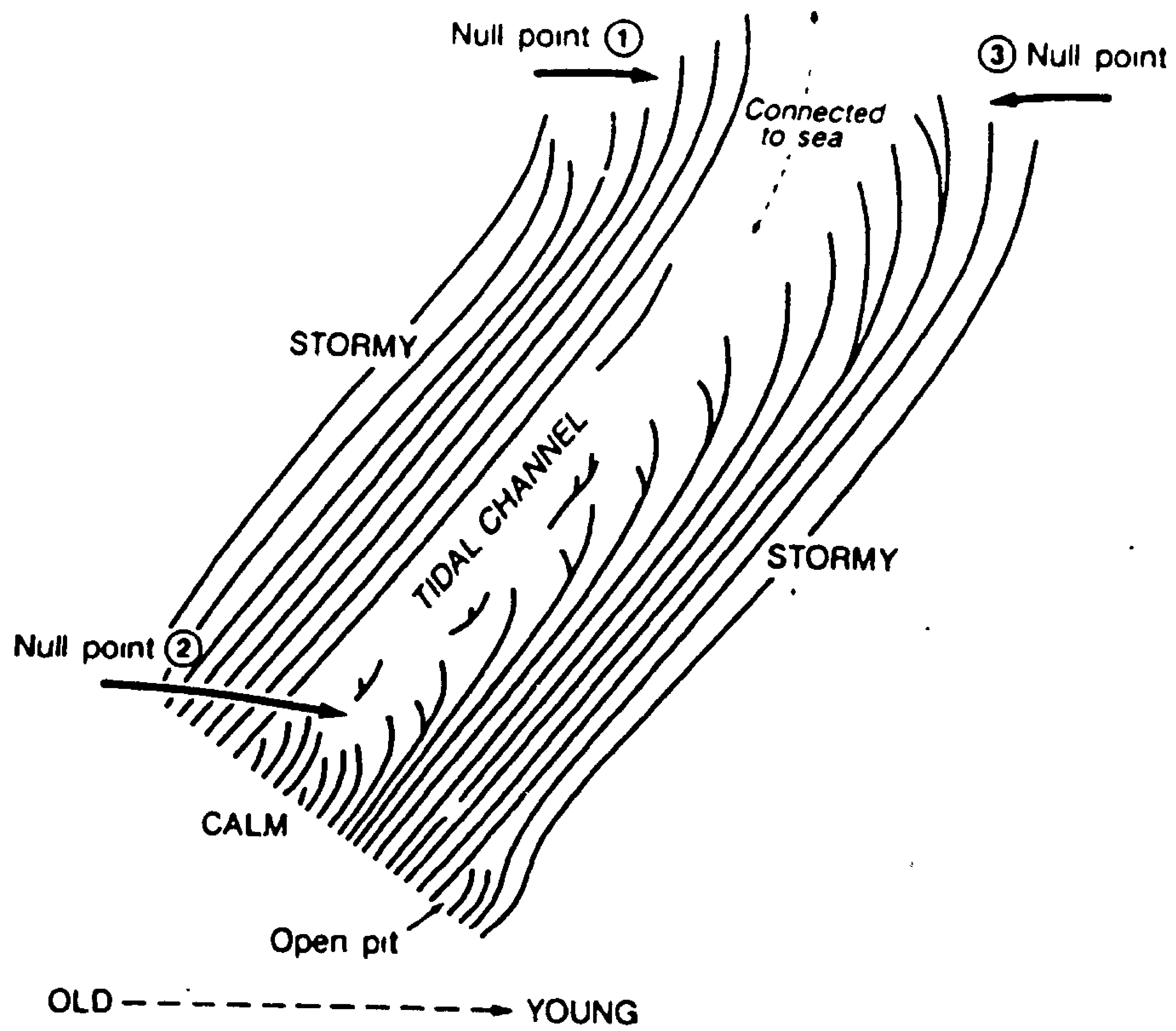

Figure 5.5: Coastal evolution under periods of high and low storm incidence and 'null point' migration.
Source: Long and Hughes (1995).

followed by a period of increased storms. This causes the deposition of a longer beach which overlaps the shorter ones, behind which an open pit may develop (figure 5.5). Long and Hughes (1995) suggest that this mode of barrier development explains not only the evolution of the open pits but also the alternating areas of gravel and marsh, such as those encountered within Scotney Marsh with the marsh sediments being deposited between the gravel ridges sometime after their deposition. Consequently, this would appear to provide the cause of deposition for both the large gravel swale, between ridges 1 and 2 (the Scotney Marsh trough), and also the sediments at the base of core G60.

The sediments from the upper part of typecore G60 (-1.84m OD) are typical of the sedimentary sequence recorded throughout the Scotney Marsh trough. In this case the environment became a marine inlet / tidal channel (Vos and de Wolf, 1993) as marine planktonic diatoms predominate in the preserved assemblages. Therefore, the Scotney Marsh trough, following the deposition of the basal part of typecore G60 (during which time the Scotney Marsh trough was relatively enclosed), appears to have been open to the direct influence of the sea. The deposition in the tidal inlet was very strongly marine with sedimentation at slack water during high tide.

The location of where this marine inlet was open to the sea is crucial to the palaeoenvironmental reconstructions at this time. To the south-west, the barrier must have been intact as The Forelands gravel complex extended in that direction, with the lows of the Midrips between them and the southerly extension of ridge 1 (figure 5.3). Equally, to the west and the north, ridge 1 is present. Thus, marine water could not have gained access from the north, south or west. Any linkage with the open sea must, therefore, have been to the north-east as any breach along the length of The Forelands would only have existed temporarily or would have developed into a more significant opening.

The location of an opening to the sea at the north-eastern end of the Scotney Marsh trough was probably where The Forelands gravel complex butted ridge 1 (figure 5.3). There may always have been a gap between ridge 1 and The Forelands. However, it is more likely that the distal end of The Forelands gravel complex abutted ridge 1 initially, leaving a sheltered environment in their lee to the west, and allowing the basal peat units to develop in an open pit. It appears that throughout the deposition of the basal peat units in core G60, sea-level was rising. This is suggested, first from the rise in the freshwater table indicated in the predominance of aquatic pollen in LPAZ G60 - IV, and secondly, from the eventual inundation of the basal peat unit and subsequent deposition of silts and gravel. This suggests that sea-level continued to rise, explaining the recorded changes in the vegetational succession. Eventually, however, a storm overtopped the protective gravel barrier. This caused silts to be deposited followed by gravels in a silt matrix. Indeed, the presence of gravels in the sediments may indicate that the gravel barrier (ridge 2) was still prograding at this time. This may have created a relatively low energy environment in which the upper organic sediments (units 6 and 7) were then deposited. Biogenic sedimentation was eventually replaced by minerogenic sedimentation with time. The age of the upper contact of the peat is unknown, but the cause of the inundation of the upper peat sediments was probably due to the continued sea-level rise again overtopping the protective gravel barrier (ridge 2). This allowed marine conditions to predominate in the Scotney Marsh trough as the sediments were deposited in an environment typical of a tidal inlet.

Throughout the Scotney Marsh trough, rare biogenic units were noted in the stratigraphic survey. These units tended to exist along the margins of the gravel ridges or in the sheltered sub-troughs within trough B. It would appear that, at some time during the deposition of the blue-grey silts, conditions at the margins became conducive to the growth of peat-forming communities in sheltered areas, suggesting that shallow water and low energy conditions prevailed in the Scotney Marsh trough prior to the deposition of the oxidation mottled silts.

5.7.2.3 The Gravel Inlet.

Within the main back-barrier environment a gravel inlet was recorded. Green (1968) previously noted a gravel inlet in the region of ridge 1, cutting across the ridge and extending from the main back-barrier environment to the Scotney Marsh trough. In this former gravel inlet, the typecores AW63 and AW-AX67 were situated. However, due to the location of the Brett Gravel plant site in this area, no stratigraphic data for the north-eastern end of this gravel inlet are available.

Minerogenic deposition began within this inlet, following the emplacement of the gravel, on an intertidal to subtidal marine-brackish mudflat. Importantly, the site was accessible by inputs from the open sea, with marine epiphytic/benthic diatoms present throughout the lower parts of cores AW63 and AW-AX67. This would not be expected if the access of marine water was gained *via* the Scotney Marsh trough, but rather more likely if the sea gained access from the main back-barrier environment, where, at this time, intertidal muds were being deposited. Upwards, the marine influence was reduced and fresh to brackish conditions began to dominate, with an allochthonous marine input throughout. This change in environment allowed a peaty-clay to be deposited as a coastal reedswamp developed, with saltmarsh and sedge fen conditions present at the margins.

The regressive contact between the intertidal muds and the reedswamp peat has been dated to 3715-3565 cal. yrs. BP in core AW63 and 4060-4040 or 3990-3700 cal. yrs. BP in core AW-AX67 (see table 4.7). These ages are of crucial importance as this regressive contact pre-dates the deposition of peat in core G60, which occurred directly on to the gravel surface in the Scotney Marsh trough. Therefore, at this time, the Scotney Marsh trough must have been a relatively dry gravel swale, cut off from the influence of the sea. Therefore, the access of the sea to the tidal gravel inlet of cores AW63 and AW-AX67 must have been through the main back-barrier environment, where, at this time, intertidal mudflats were being deposited, prior to the period of widespread peat accumulation in Scotney Marsh. Peat forming conditions did, however, clearly prevail in this gravel inlet which was relatively sheltered and protected from the intertidal conditions that existed in the main back-barrier environment at this time.

The conditions that promoted the establishment of peat-forming communities within the gravel inlet may have been a consequence of sediments infilling the sheltered inlet and becoming sub-aerially exposed, leading to salt tolerant plants (halophytes) colonising the mudflats.

Eventually, peat deposition was brought to an end with the deposition of a blue-grey silty-clay as the saltmarsh conditions at the upper contact of the peat were replaced by deposition of silt and clay on an intertidal marine to brackish mudflat. The cause of the return to minerogenic deposition in the sheltered inlet was probably continued sea-level rise overtaking the rate at which peat could accumulate. This continued sea-level rise at ca. 4000 cal. yrs. BP is in contradiction to the proposal of Long and Innes (1993), *i.e.* that sea-level was relatively stable between 4500 and 1950 cal. yrs. BP.

The marine influence in the gravel inlet was reduced once more, resulting in the prevailing salinity becoming more fresh to brackish. Hence, another regressive contact is recorded by the succession of intertidal mudflat, saltmarsh and coastal reedswamp within the gravel inlet. The entire upper peat unit was dated to 3265-2930 cal. yrs. BP in core AW63, and a regressive contact gave an age of 3220-2795 cal. yrs. BP in core AW-AX67 (see table 4.7). These ages are similar to the ages of the regressive contacts marked by peat initiation in the main back-barrier environment in Scotney Marsh. This synchroneity suggests a strong linkage between the evolution of the main back-barrier and the gravel inlet. The similarities of the ages for peat initiation in both of these areas indicate that this period of peat deposition was facilitated by a relatively widespread process. This process was probably a stabilisation or slowing of the rate of sea-level rise, along with the infilling of the back-barrier environments with sediment leading to sub-aerial exposure and eventual colonisation by vegetation of the mudflats and, thus, accumulation of organic deposits.

Peat accumulation eventually came to an end as a reversal of the vegetational succession is observed, with saltmarsh conditions present in a brackish water environment at the upper peat contact where it gives way to intertidal mudflats. This transgressive contact was dated to 2950-2725 cal. yrs. BP (see table 4.7); an age similar to the transgressive contacts of the main back-barrier environment in Scotney Marsh (cores AY17 and A-B27). This age also post-dates the end of peat deposition in the Scotney Marsh trough (at core G60) indicating that intertidal sedimentation had already begun in the trough by this time.

The close correspondence between the ages for the end of peat deposition in both the main back-barrier environment in Scotney Marsh and the gravel inlet again suggests that a relatively widespread change in environment took place. It appears that sea-level may have continued to rise until such a time as it exceeded the rate of peat accumulation and, thus, marine conditions inundated the peat. This is recorded by the transgressive contacts in core AY17, A-B27, AW63 and AW-AX67. The route by which the marine water gained access to Scotney Marsh may have been the more extensive flooding, driven by sea-level rise, of one of the tidal channels, such as the Wainway Channel to the north of Scotney Marsh.

The uppermost sediments of the gravel inlet, the oxidation mottled silts, have sedimentary characteristics similar to those observed for the oxidation mottled silts across the whole of Scotney Marsh, *i.e.* the orangey-grey silty-clay to silty-sand. It appears that the deposition of this facies was synchronous across the whole of Scotney

Marsh, bringing to an end biogenic deposition in this area. In many cores, the deposition of this mottled unit truncated the upper surface of the underlying peat unit. Indeed sharp, eroded, upper contacts of the peat were observed in many of the cores sampled in the stratigraphic survey. The cause of the deposition of this uppermost unit is not clear. This is primarily due to the lack of palaeoenvironmental indicators within the sediments. However, deposition appears to have taken place on a brackish tidal mudflat with a significant open marine influence. An increase in the energy of the environment is recorded upward as mean grain size increase and the sediments become strongly fine-skewed, indicating an increase in tidal flow velocity. However, the environment was still relatively protected from the direct influence of the sea as no indicators of wave action were observed, *i.e.* no consistent negative skew in the particle size distributions.

The deposition of the oxidation mottled silts was probably caused by rising sea-level as outlined above. The transgressive contact between the peat and the start of minerogenic deposition across Scotney Marsh pre-dates the renewed period of sea-level rise as proposed by Long and Innes (1993), *i.e.* ca. 1950 cal. yrs. BP. Therefore, this provides further evidence that sea-level rise had continued throughout the period during which Long and Innes (1993) proposed that it was stable (ca. 4500 to 1950 cal. yrs. BP).

6 Sediment Source Determination for Scotney Marsh.

6.1 Introduction

An attempt is made here to determine the source of the sediments that make up the Holocene sediments of Scotney Marsh, as evaluation of sediment provenance may assist in the palaeoenvironmental reconstruction concerning the influence of the sea on back-barrier sedimentation. This will be completed by the utilising mineral magnetic analysis. The methods utilised in the completion of the magnetic analysis are outlined in section 3.1.

6.2 Core AY17, Mineral Magnetic Analysis : Results

The results of the mineral magnetic analysis for core AY17 are presented in figure 6.1 and summarised in table 6.1. It is clear that from figures 6.1 that four zones of magnetic properties exist. Due to the very high values of χ, SIRM and χARM recorded in zones AY17/I and III, the detail of any variation in zones AY17/II and IV is masked. However, on removal of these extremes, and inspection of the remaining data, it is clear that little variation is actually present and, thus, these sections of relatively low values of χ, SIRM and χARM are identified as coherent magnetic zones.

AY17/I : -2.11 - -1.79m OD

The values χ are relatively low, whereas SIRM and χARM are relatively high. This suggests a relatively low

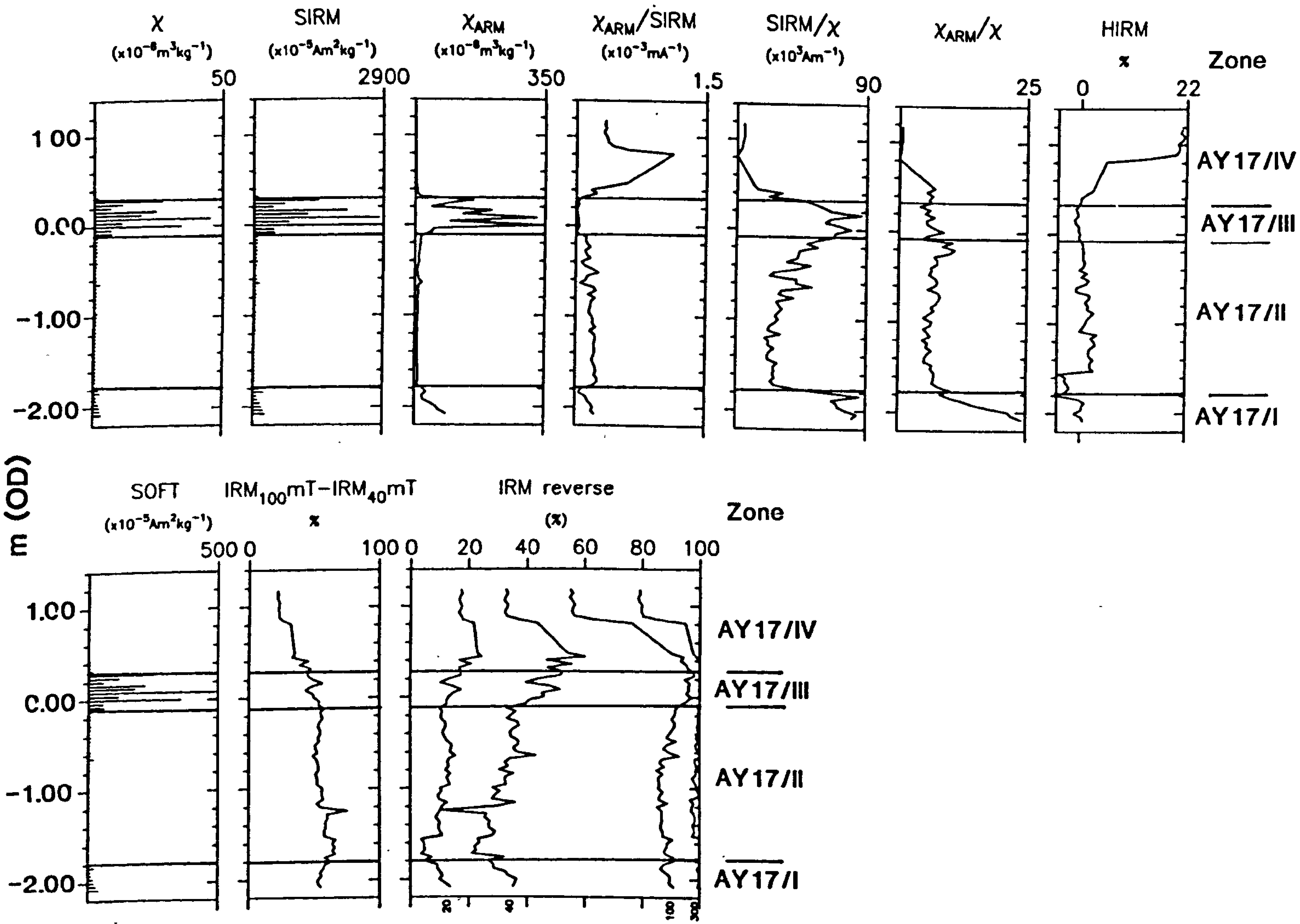

Figure 6.1: Mineral magnetic analysis, core AY17.

concentration of magnetic minerals, with grains of fine single domain size dominating. The values of χARM/χ and χARM/SIRM are both relatively high but decrease upward. This upward trend indicates a fall in the proportion of fine single domain grains. Very low values of HIRM indicate low contributions from haematite and goethite, whilst divergence of the backfield ratios together with very high values of SIRM/χ (>40 KAm^{-1}), indicate a strong contribution of authigenic greigite.

AY17/II : -1.79 - -0.11m OD

Very low values are recorded for χ, SIRM and χARM throughout the zone. This indicates a very low concentration of magnetic minerals. High values are recorded for SIRM/χ which increases upward, with many samples >40 Kam^{-1}. Significantly, divergence of the backfield IRMs is also present throughout the zone, indicating the continued presence of greigite with its influence on the mineral magnetic character of the sediment increasing upward. The low values of HIRM indicate that little haematite or goethite is present. Little variation is recorded for either χARM/SIRM or χARM/χ, suggesting limited variation in magnetic grain size through the zone.

AY17/III : -0.11 - +0.29m OD

High values of χ, SIRM and χARM are recorded, indicating a high concentration of magnetic minerals. Very low values for χARM/SIRM indicate that multidomain magnetic grains predominate, with low proportions of fine single domain grains. Very high values of SIRM/χ (>60 Kam^{-1}) are present throughout the zone, as well as divergence of the backfield IRMs, indicating a marked degree of overprinting by greigite.

Zone (m OD) / Units	χ ($x10^{-8}m^3kg^{-1}$)	χfd %	SIRM ($x10^{-5}Am^2kg^{-1}$)	χARM ($x10^{-8}m^3kg^{-1}$)	χARM/SIRM ($x10^{-3}mA^{-1}$)
I : -2.11 to -1.79	2.46	3.41	188.33	44.03	0.23
II : -1.79 to -0.11	1.16	2.70	41.26	9.82	0.26
III : -0.11 to +0.29	19.02	1.24	1312.13	147.78	0.11
IV : +0.29 to +1.17	0.73	2.16	12.50	3.40	0.47

Zone (m OD) / Units	SIRM/χ ($x10^3Am^{-1}$)	χARM/χ ($x10^3Am^{-1}$)	HIRM ($x10^{-5}Am^2kg^{-1}$)	HIRM %	SOFT ($x10^{-5}Am^2kg^{-1}$)	GREIGITE ($x10^{-5}Am^2kg^{-1}$)
I : -2.11 to -1.79	75.17	17.21	-0.70	-0.45	19.51	212.83
II : -1.79 to -0.11	34.01	8.47	0.19	0.63	4.57	46.47
III : -0.11 to +0.29	68.74	7.56	-8.76	-0.64	189.43	1296.20
IV : +0.29 to +1.17	12.44	4.23	0.43	12.50	2.22	10.42

Table 6.1 : Mean values for mineral magnetics, core AY17.

Zone (m OD) / Units	χ ($x10^{-8}m^3kg^{-1}$)	χfd %	SIRM ($x10^{-5}Am^2kg^{-1}$)	χARM ($x10^{-8}m^3kg^{-1}$)	χARM/SIRM ($x10^{-3}mA^{-1}$)
I : -2.33 to -1.01	27.60	6.64	1628.81	224.40	0.21
II : -1.01 to -0.89	54.15	1.52	5183.02	478.40	0.09
III : -0.89 to +0.07	19.88	2.35	1121.17	214.67	0.22
IV : +0.07 to +0.11	117.96	0.13	10602.82	1140.61	0.11

Zone (m OD) / Units	SIRM/χ ($x10^3Am^{-1}$)	χARM/χ ($x10^3Am^{-1}$)	HIRM ($x10^{-5}Am^2kg^{-1}$)	HIRM %	SOFT ($x10^{-5}Am^2kg^{-1}$)	GREIGITE ($x10^{-5}Am^2kg^{-1}$)
I : -2.33 to -1.01	45.44	9.00	8.30	1.30	205.13	1744.107
II : -1.01 to -0.89	94.26	8.83	-2.21	0.16	443.17	6171.02
III : -0.89 to +0.07	51.75	10.850	16.15	1.79	109.89	1286.557
IV : +0.07 to +0.11	89.88	9.67	-43.40	-0.41	1012.50	12459.22

Table 6.2 : Mean values for mineral magnetics, core G60.

AY17/IV : +0.29 - +1.18m OD

Very low values are recorded for χ, SIRM and χARM, reflecting very low concentrations of magnetic minerals. The χARM/SIRM ratio is initially low, indicating a high proportion of multidomain magnetic grains, but rises sharply to a high value of 1.14 KmA^{-1} as the proportion of fine single domain grains increases before falling again at the top of the zone. An increase in the proportion of hard magnetic minerals, e.g. haematite and goethite, is recorded upward through the zone. In the lower part of the zone, relatively high values for SIRM/χ, with divergence of the backfield IRMs illustrate the presence of greigite. The contribution from greigite decreases upward, as relatively low values of SIRM/χ and near parallel backfields are recorded.

6.3 Core G60, Mineral Magnetic Analysis : Results

The results of the mineral magnetic analysis of core G60 are presented in figure 6.2 and summarised in table 6.2. It can be seen (figures 6.2) that relatively little variation was recorded upward through the core, although a number of significant exceptions make up the separate zones, e.g. G60/II and IV.

G60/I : -2.37 - -1.01m OD

The values of χ, SIRM and χARM are generally low. This illustrates that relatively low concentrations of magnetic minerals are present. Notable exceptions to this exist at

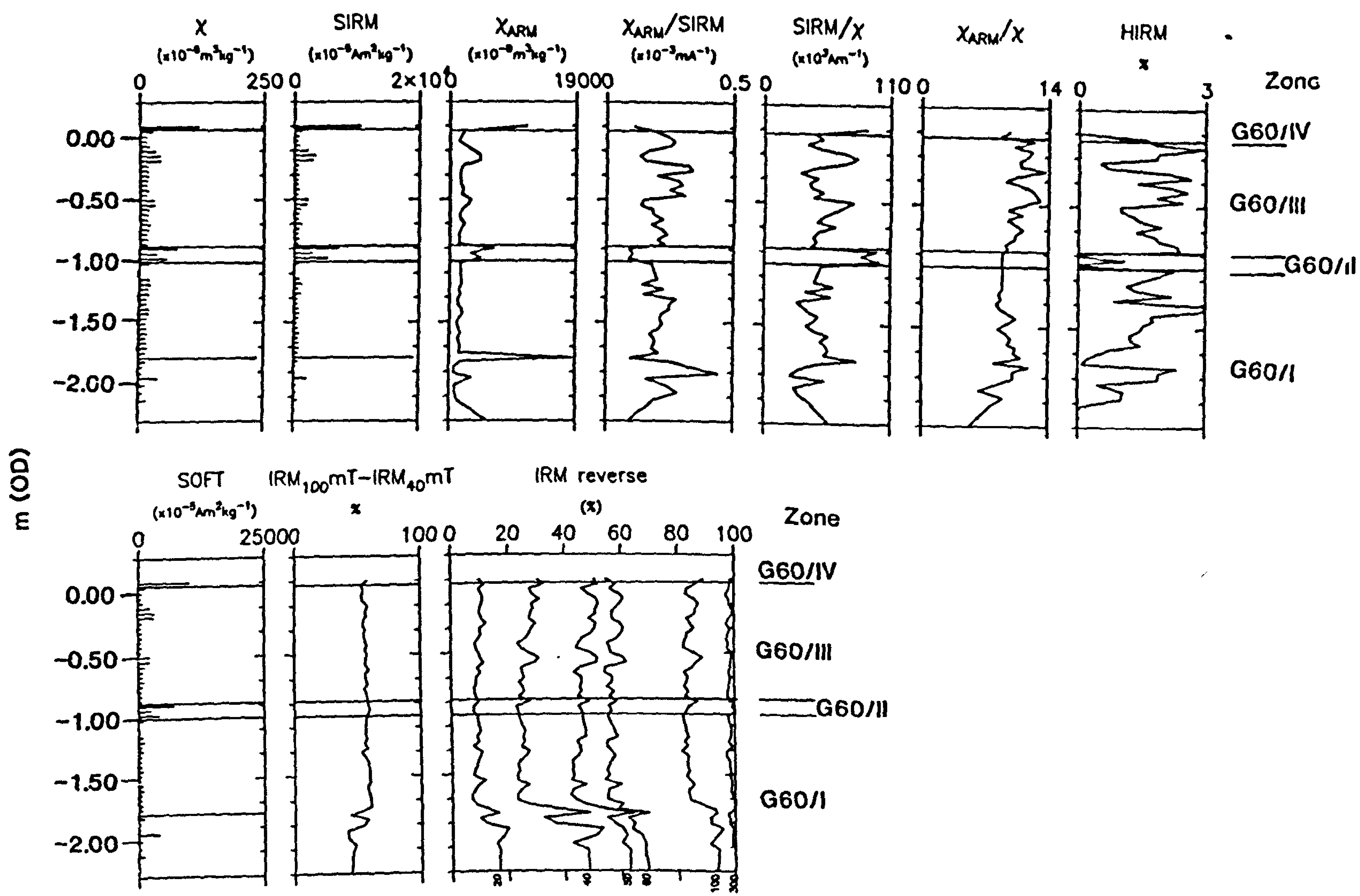

Figure 6.2: Mineral magnetic analysis, core G60.

altitudes of -2.31m OD and -1.80m OD, which reflect high concentration of magnetic minerals at these levels. The opposite trend is observed for χARM/SIRM and χARM/χ, as the values are relatively high throughout the zone and indicate that fine single domain magnetic grains predominate. Conversely, the samples that exhibit high magnetic concentrations, *i.e.* -2.31m OD and -1.80m OD, have relatively low values for the above quotients and, therefore, are dominated by multidomain magnetic grains.

Very low values are recorded for HIRM throughout the zone; generally less than 3% with many samples significantly lower. This illustrates a low contribution from haematite or goethite to the magnetic assemblage. The values for SIRM/χ are relatively high, especially for the two samples at altitudes of -2.31m OD and -1.80m OD, with many values >40 Kam^{-1}. Importantly, there is also divergence of the backfield IRMs which corroborates the evidence from the SIRM/χ that greigite dominates the mineral magnetic characteristics of this zone.

G60/II : -1.01 - -0.89m OD

Notable increases in the values of χ, SIRM, and χARM are recorded in this zone, indicating an increase in the concentration of magnetic minerals. The relatively low values for χARM/SIRM suggest that the proportion of fine single domain magnetic grains is reduced compared with the underlying zone. Low to very low values are recorded for HIRM %, suggesting that the sediments of this zone contain a very low proportion of haematite or goethite. Very high values of SIRM/χ, (~90 Kam^{-1}) and divergence of backfield IRMs are clear illustrations of a strong greigite component.

G60/III : -0.89 - +0.07m OD

The values recorded for the concentration parameters χ, SIRM and χARM are relatively low and fluctuate through the zone. The quotients χARM/χ and χARM/SIRM are relatively high and variable, indicating slight fluctuations in the magnetic grain size with a high proportion of fine single domain magnetic grains throughout. The values for SIRM/χ are high, predominantly >40KAm^{-1}, and significant divergence of the backfields IRMs illustrates that greigite dominates the magnetic characteristics of this zone.

G60/IV : +0.07 - +0.11m OD

In this uppermost zone, a single sample exhibits significantly different mineral magnetic characteristics from those sediments of the underlying zone. The concentration parameters are relatively high, illustrating an enhanced concentration of magnetic minerals. Conversely, relatively low values are recorded for χARM/χ and χARM/SIRM, indicating a low proportion of magnetic grains of fine single domain size. A high value of SIRM/χ and divergence of the backfield IRMs clearly illustrates the predominance of greigite, whilst a very low HIRM reflects a limited contribution from either haematite or goethite.

6.4 Mineral Magnetic Analysis : Summary

The mineral magnetic analyses for both typecores demonstrate that greigite dominates the magnetic signature throughout the two cores. Greigite has a very strong magnetic signal and, due to the fact that greigite forms authigenically in the sediment, overprints any detrital magnetic signal that may have been present in the sediment at the time of deposition. As it is the detrital signal that is sought in order to determine the source of the marsh sediments, the presence of authigenic greigite presents significant problems.

Similar overprinting was observed by Spencer (1992) in the Denge Marsh area of Romney Marsh. Here, too, greigite masks the detrital magnetic signal and, thus, prevents the determination of sediment source. An attempt was made by Spencer (1992) to chemically remove the authigenic greigite in order to measure the detrital signal. Some qualified success was achieved in that some of the greigite was removed. However, it was not determined whether all of the greigite was removed or to what extent the detrital magnetic signal was altered by the extraction process.

In the case of Denge Marsh, authigenic greigite presented a significant problem, particularly as only very subtle differences existed between the magnetic signals of the potential source areas for marsh sediment. The potential sources of the sediment according to Spencer (1992) were the Wealden sediments to the north and west of Romney Marsh and the marine sediments from the English

Channel. However, the latter source area would have included reworked Wealden sediments from either the Holocene rivers or coastal glacial deposits, as well as sea bed sediment from west of Romney Marsh. Equally, Wealden sediments from the Romney Marsh catchment would have been transported to the sea throughout the evolution of the marsh and, to some extent, mixed with the sea bed sediments during deposition, thus making the separation of the potential source areas difficult using magnetic properties. Therefore, it was clear that the potential sediment sources had relatively weak and somewhat similar detrital magnetic signals, whereas the authigenic greigite overprint possessed a relatively very strong magnetic signal such a situation makes sediment source ascription very difficult.

In the light of this, further research was necessary in this study to attempt to remove the greigite overprint. Chemical extraction had previously proved to be unsuccessful (Spencer, 1992) and, thus, another approach was investigated. This approach included the separation of various particle size classes of a number of samples to determine whether greigite was absent in any size classes. If this proved to be the case, it would then have been possible to analyse the detrital magnetic signal of the particle size classes not significantly overprinted, and to compare the results with similar grain sizes from the potential source areas.

Samples were selected (as described in section 3.5.6), and the selected phi sizes classes were then separated, (as described in section 3.5.7). Samples ranging from 3ϕ (fine sand) to 9ϕ (clay) were subjected to the typical suite of mineral magnetic measurements to determine their magnetic properties (figures 6.3 and 6.4 and table 6.3).

6.5.1 AY17G, Particle Size Separation : Results

Sample AY17G was made up of a number of samples from core AY17 whose mineral magnetic properties were dominated by greigite. The finer sediments, *i.e.* 4-9ϕ, exhibit relatively low values for the concentration parameters (figure 6.3 and table 6.3), whereas the 3ϕ size class has a significantly higher concentration of magnetic minerals. Importantly, it is clear that greigite dominates the magnetic properties throughout the particle size range. This is illustrated by the clear divergence of the -20mT and -40mT backfields from the -100mT and -300mT backfields, as demonstrated by the high (>40%) values of IRM_{100mT}-IRM_{40mT}. The values of SIRM/χ are also relatively high for all but the 9ϕ particle size class. This suggests, therefore, that the mineral magnetic properties are dominated by greigite for all particle sizes except the 9ϕ size class.

6.5.2 AY17N, Particle Size Separation : Results

Sample AY17N comprised a number of samples from core AY17 whose mineral magnetic properties demonstrated relatively little evidence of greigite. However, the mineral magnetic characteristics of the range of particle size classes appear to have some significant contribution

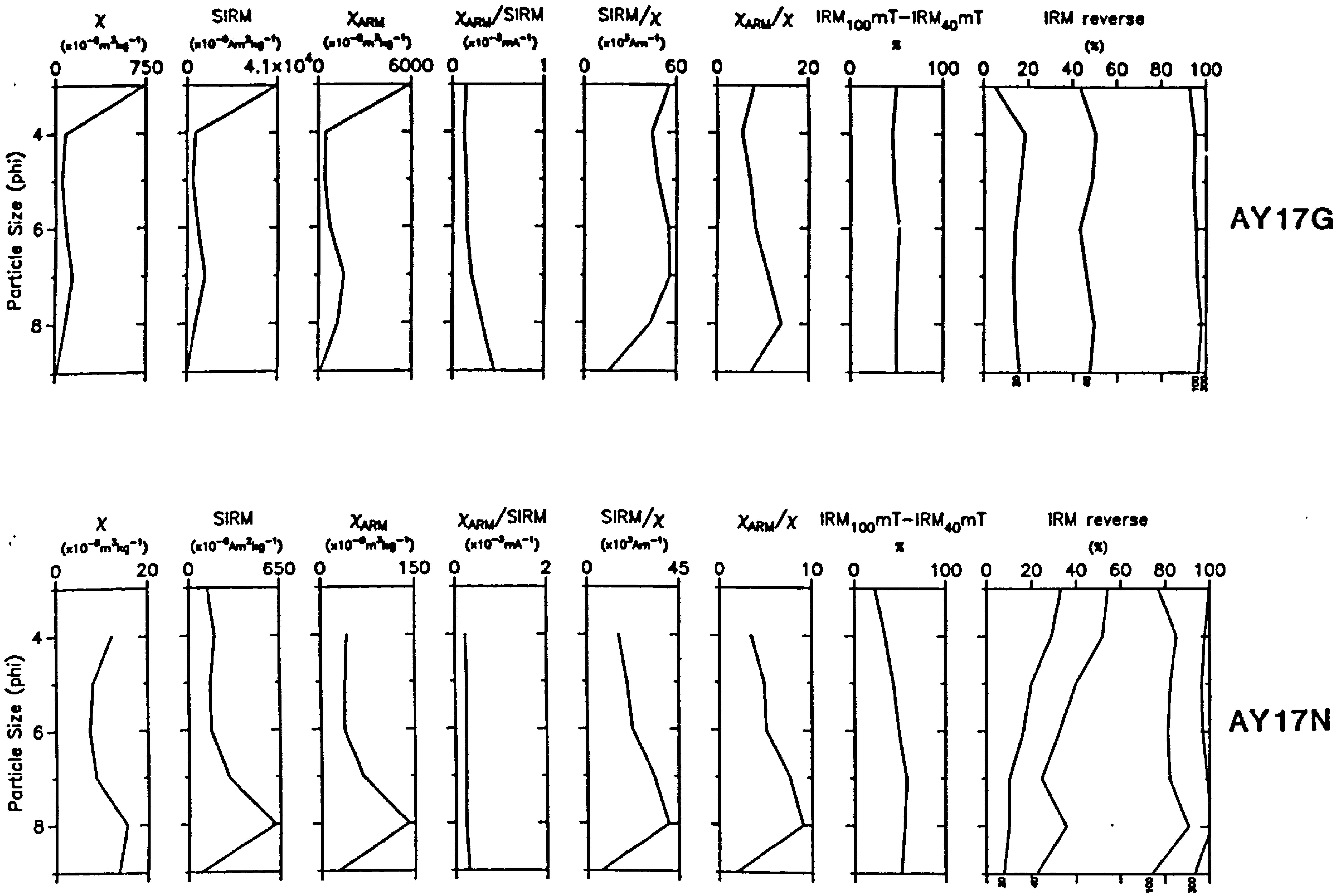

Figure 6.3: Mineral magnetic analysis on separated particle size fractions AY17G and AY17N.

from greigite (figure 6.3 and table 6.3). The divergence of the -20mT and -40mT backfields from the -100mT and -300mT backfields is relatively large for the finer particle sizes, but is reduced in the 3ϕ and 4ϕ classes. Similarly the values of SIRM/χ fall as the sediments coarsen. Significantly lower values were recorded in sample AY17N for χ, SIRM and χARM than for sample AY17G, indicating that the concentration of magnetic minerals in this sample is relatively low. Therefore, much of the relatively weak magnetic signal was dominated by greigite in all but the 3ϕ and 4ϕ particle sizes, and even here some greigite contribution cannot be discounted as some divergence of the backfield IRMs may still be present.

Sample Name	χ $\times10^{8}m^3kg^{-1}$	SIRM $\times10^{5}Am^2kg^{-1}$	χARM $\times10^{8}m^3kg^{-1}$	χARM/SIRM $\times10^{-3}mA^{-1}$	SIRM/χ $\times10^{3}Am^{-1}$	χARM/χ $\times10^{3}Am^{-1}$	GREIGITE $\times10^{-5}Am^2kg^{-1}$
AY17N	9.29	232.17	99.31	0.58	19.48	4.59	44.31
AY17G	173.32	9180.15	1513.35	0.22	45.39	8.86	48.54
G60N	20.22	1197.02	177.49	0.20	53.78	9.81	54.92
G60SG	44.99	3253.40	321.19	0.13	62.00	6.81	58.10

Table 6.3 : Mineral magnetic mean values of separated particle sizes.

6.5.3 G60G, Particle Size Separation : Results

This sample was made up of a number of samples from core G60 whose magnetic properties were dominated by greigite. A similar trend is exhibited for each of the concentration parameters, which gradually rise as the sediment coarsens but fall again for the 3ϕ and 4ϕ samples

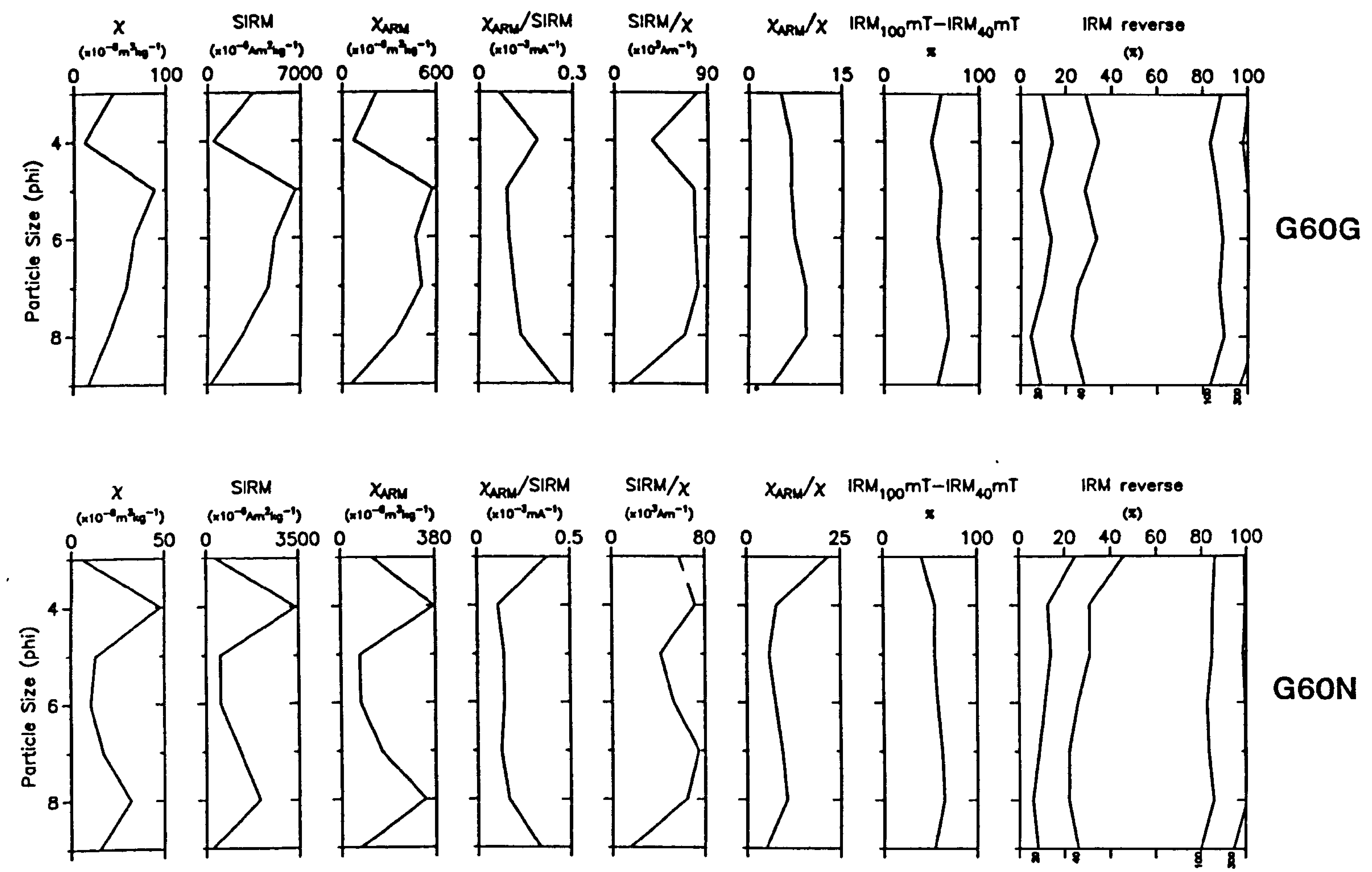

Figure 6.4: Mineral magnetic analysis on separated particle size fractions G60G and G60N.

(figure 6.4 and table 6.3). Importantly, considerable divergence occurs in the backfield IRMs with the IRM_{100mT}-IRM_{40mT} being high at >48% throughout. The values of SIRM/χ are also relatively high (at >75% Kam^{-1}) for all except the 4ϕ and 9ϕ samples. This clearly illustrates that greigite dominates the magnetic properties in most of the particle size classes and is present, although with a slightly weaker magnetic signal, in the 4ϕ and 9ϕ samples.

6.5.4 G60N, Particle Size Separation : Results

G60N was made up of a number of samples whose magnetic properties demonstrated a relatively weak greigite component. The values for the concentration parameters are variable throughout the range of particle sizes, but are significantly lower than the values for the concentration parameters of G60G (figure 6.4 and table 6.3). The backfields diverge for all of the particle sizes, but to a reduced extent for the 3ϕ particle size. This is also illustrated by the high values (>40%) for IRM_{100mT}-IRM_{40mT} throughout the particle size range. In addition, very high values of SIRM/χ (>40 Kam^{-1}) are recorded for all particle sizes except 9ϕ. Therefore, it appears that greigite dominates the mineral magnetic character of all of the particle sizes to a greater or lesser extent.

6.6 Particle Size Separation : Summary

Following the particle size separation and mineral magnetic analysis, it was concluded that greigite was present to some extent in all the but the finest particle size classes and, indeed, dominated the magnetic properties of most. Clearly the greigite component was not confined to one particle size. Equally no particle size class was identified in which one could be confident that greigite did not make a contribution to the magnetic properties. In the 9ϕ fractions of each of the samples, relatively low values of SIRM/χ are recorded. However, in these samples the IRM_{100mT}-IRM_{40mT} is at ~50% and, thus, some contribution by greigite to the magnetic signal cannot be discounted. In addition, the 9ϕ fraction forms only a small proportion of the core and source sediments and, therefore, sediment source ascription is not possible for much of the core material or potential source areas.

7 The Evolution of Romney Marsh : Introduction.

In the previous chapters conclusions have been reached regarding, first, the site-specific evolution of the barrier / back-barrier environment at Scotney Marsh, and secondly, the overall succession of the marsh sediments after considering a number of key site-specific studies completed on Romney Marsh. These findings will now be utilised to discuss a number of key evolutionary elements, such as the initial protective barrier and the back-barrier drainage and to formulate a model of the evolution of Romney Marsh.

7.1 The Initial Protective Gravel Barrier.

It was determined by Long and Innes (1993; 1995b) that the Midley Sand of Green (1968) was not part of an initial protective gravel barrier. This theory had been adopted previously by a number of workers (Cunliffe, 1980; Lake and Shephard-Thorn, 1987; Greensmith and Gutmanis, 1990). The belief that a barrier must have been present in order to facilitate the deposition of the low energy deposits, *i.e.* mudflats, and the proliferation of peat communities is suggested in a number of studies, e.g. Waller *et al.* (1988). However, Long and Innes (1993) identified that direct stratigraphic or biostratigraphic evidence indicating that marsh sedimentation had occurred behind a protective gravel barrier is sparse. They proposed that were it not for the presence of the gravel of the Dungeness foreland, many of the sequences recorded across the marsh would be interpreted as a progressive decrease and then increase in the marine influence.

It was, however, noted by Long and Innes (1993) that the study of Tooley and Switsur (1988) at Broomhill did indicate that a gravel barrier had provided the protection that facilitated marsh sedimentation. In addition, Long and Innes utilised the discovery of Early Bronze Age axes at Lydd (Needham, 1988) and, also, the fact that the blue clay and the peat units are not recorded to the east of Lydd (Long and Fox, 1988; Plater, 1992; Plater and Long, 1995; Long and Hughes, 1995) to cautiously interpret the gravel at Broomhill and Lydd as part of an initial gravel barrier.

The detailed stratigraphic sampling completed in this study now enables the location of the initial protective gravel barrier to be determined with increased accuracy. This stratigraphic research has indicated that a sedimentary sequence, typical of the back-barrier environment in Romney Marsh (blue clay, main marsh peat, and oxidation mottled silts), was recorded to the north-west of ridge 1. Conversely, in the Scotney Marsh trough this typical sedimentary sequence was largely absent. The palaeoenvironmental reconstructions of the main back-barrier environment (cores AY17, A-B27, AW63, AW-AX67) and the Scotney Marsh trough (core G60) also indicate that ridge 1 appears to mark a boundary between the two environments, whose evolution would appear to have been separate.

Therefore, it seems that an initial protective gravel barrier existed on an approximate line from Broomhill extending north-eastwards towards Scotney Marsh (as illustrated in figure 7.7). Minimum ages for the emplacement of gravel are provided by dates obtained from peats that directly overlie the gravel barriers. Tooley and Switsur (1988) dated a peat unit present in a gravel swale at Broomhill to 3817-3622 cal. yrs. BP. In the present study, peat units found to be overlying the gravel have been dated, with the earliest age of 4060-4040 or 3990-3700 cal. yrs. BP indicating a minimum age for gravel emplacement.

An alternative method for the dating of the emplacement of the initial barrier is provided by the proxy record of datable back-barrier sediments. For example Waller *et al.* (1988) suggest that the age of the emplacement of the initial barrier must have been ca. 6850 cal. yrs. BP so as to facilitate the deposition of the back-marsh peat units in the Brede Valley after this time. Similarly, Long *et al.* (1996) obtained an age of 6497-6280 cal. yrs. BP for the initiation of peat at Rye, and Long and Hughes (1995) suggested that barrier emplacement allowed peat to develop at Horsemarsh Sewer at 6164-5736 cal. yrs. BP (Tooley and Switsur, 1988). Long and Innes (1995a), therefore, propose that the barrier initiation began between ca. 6850 and 5750 cal. yrs. BP, creating a morphological barrier behind which the subsequent marsh sediments were deposited. Although this cannot be supported by the present study, a back-barrier environment would appear to have been well-established by ca. 4000 cal. yrs. BP.

7.2 Back-barrier Drainage.

Throughout the evolution of the back-barrier environment, large quantities of both freshwater, from the Weald, and also marine water, following each high tide, would have been present in the back-barrier. Consequently, a gap in the protective gravel barrier must have existed, and in addition, a tidal channel or channels must have acted to transport this water on and off the marsh. The wide ranging debate regarding the drainage of the back-barrier environment has been summarised in section 2.1.5. Long and Innes (1995a) demonstrated that large areas of sediments, that they interpret as channel sands, exist in Walland Marsh (figure 7.1). Figure 7.1 illustrates the extent of stratigraphic work completed in Romney Marsh previous to this study, and shows that a number of large channels appear to have dissected the marsh. The work completed in the present study, *i.e.* stratigraphic transects I and II, has added significant detail to the broad stratigraphic knowledge of the Romney Marsh region, and, in particular, the determination of the location and dimensions of the tidal channels (figure 7.2).

Long and Innes (1995a) identified two distinct areas of channel sands, one extending south of Brookland for ca. 2.0km towards Old Cheyne Court, the other further south

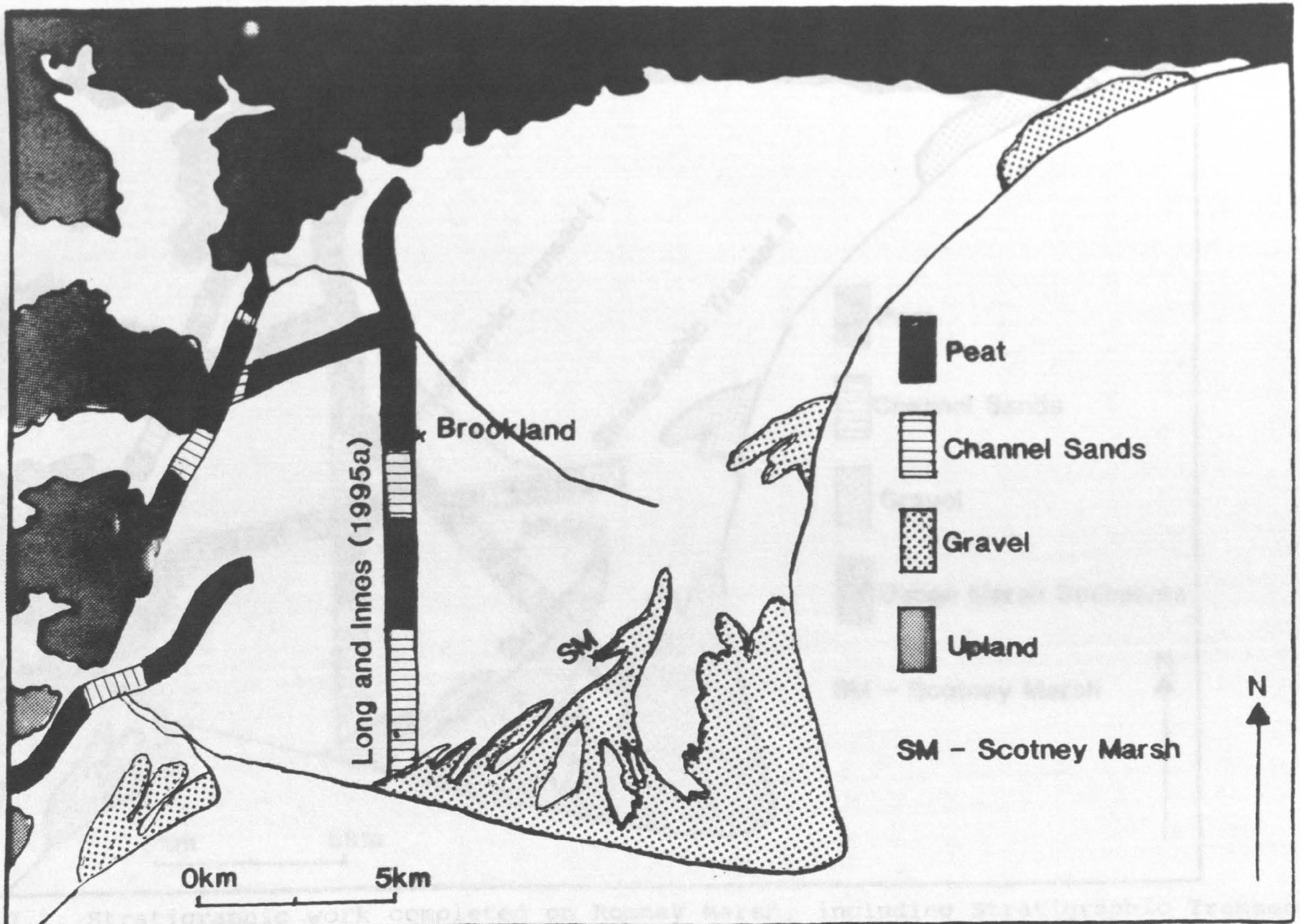

Figure 7.1: Stratigraphic work completed on Romney Marsh previous to this study.

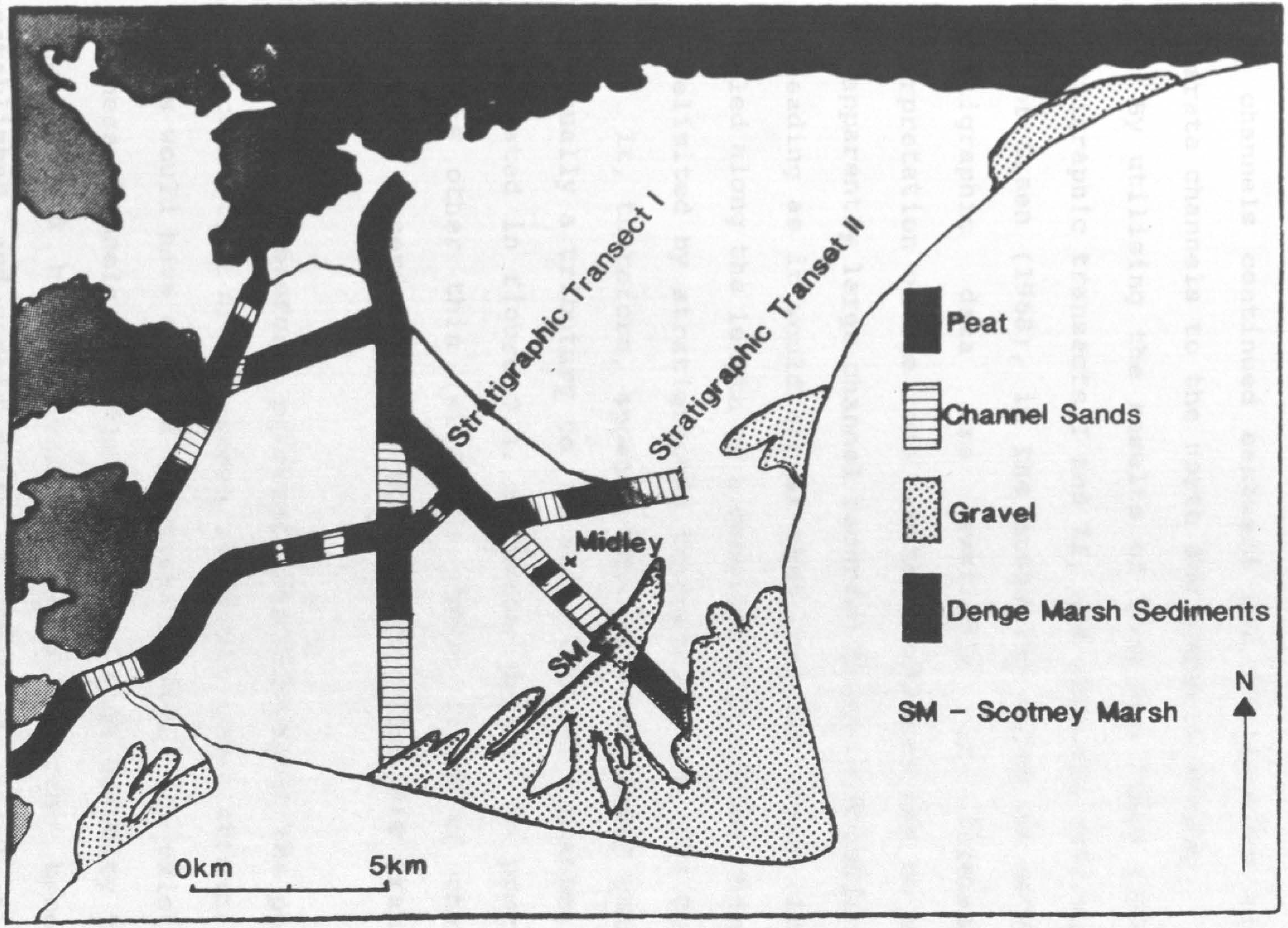

Figure 7.2: Stratigraphic work completed on Romney Marsh, including Stratigraphic Transects I and II.

extending from Little Cheyne Court towards Broomhill for ca. 2.5km. The addition of stratigraphic transects I and II in this study assists greatly in determining the locations of these tidal channels. Clearly the more southerly of the two channels continued eastward and is recorded as two separate channels to the north and south of Midley.

By utilising the results of Long and Innes (1995a), stratigraphic transects I and II, and also the soil survey map of Green (1968), in the areas for which no detailed stratigraphic data are available an interesting interpretation of the more northerly channel can be made. The apparently large channel recorded south of Brookland is misleading as it would appear that Long and Innes (1995a) sampled along the length of a considerably thinner channel, as delimited by stratigraphic transect II near Old Cheyne Court. It, therefore, appears that this northerly channel is actually a tributary to the major southerly channel, as illustrated in figure 7.3. A similar pattern is proposed for the other thin (<200m to <500m) areas of channel sediments recorded to the west in stratigraphic transect II.

It is, therefore, proposed that throughout the period of accumulation of the marsh sediments tidal streams and rivers would have crossed the marsh surface. The existence of these channels, and the relatively high velocity tidal flows, would have prevented vegetation from becoming established, and consequently, explains the distinct lack of peat in these areas. A southerly channel in the Walland

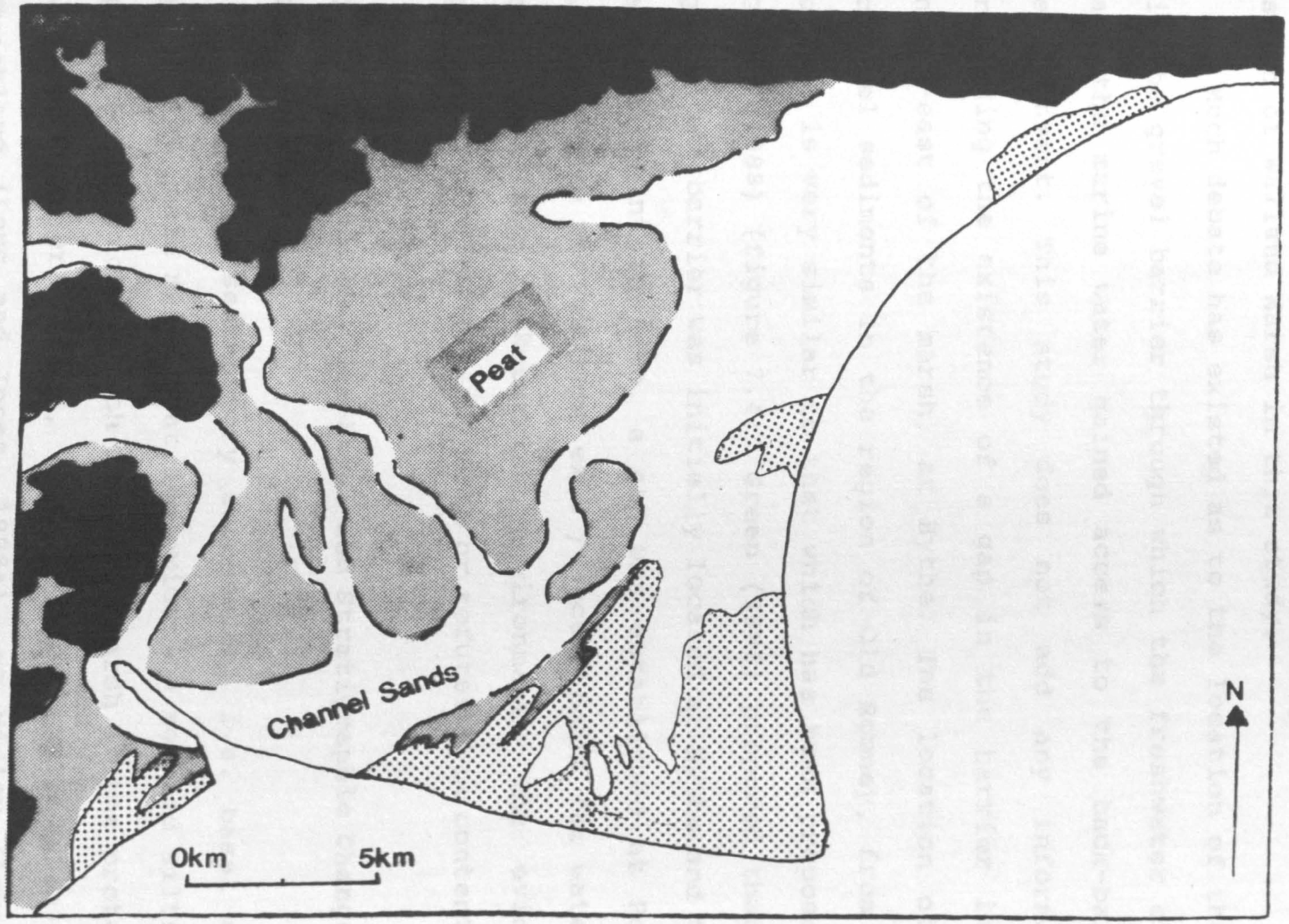

Figure 7.3: Interpretation of the stratigraphic data from Romney Marsh, delimiting the locations of the channel sediments.

Marsh area has been referred to previously as the 'Wainway Channel' in a number of studies, e.g. Gardiner (1988) and Long and Innes (1993). Therefore, it is proposed that this name by applied to the channel sediments recorded in the south of Walland Marsh in this study.

Much debate has existed as to the location of the gap in the gravel barrier through which the freshwater exited and the marine water gained access to the back-barrier environment. This study does not add any information regarding the existence of a gap in the barrier in the north-east of the marsh, at Hythe. The location of the channel sediments in the region of Old Romney, from this study, is very similar to that which has been proposed by Green (1988) (figure 7.4). Green (1988) proposed that the gap in the barrier was initially located at Hythe and that, at some point in time, a gap was established at Romney which became the dominant exit / access point for water to and from the back-barrier environment. The evidence presented here neither supports or refutes this contention.

7.3 Time / Altitude, Sea-Level and Stratigraphic Changes in Romney Marsh.

A typical sedimentary succession, *i.e.* basal sand, blue clay, main marsh peat and oxidation mottled silts has been recorded across much of Romney Marsh in the protected valleys (Waller *et al.*, 1988; Waller 1993; 1994a), at Brookland (Long and Innes, 1995a), at Midley (Long and Innes, 1993; 1995b), across much of Scotney Marsh and in

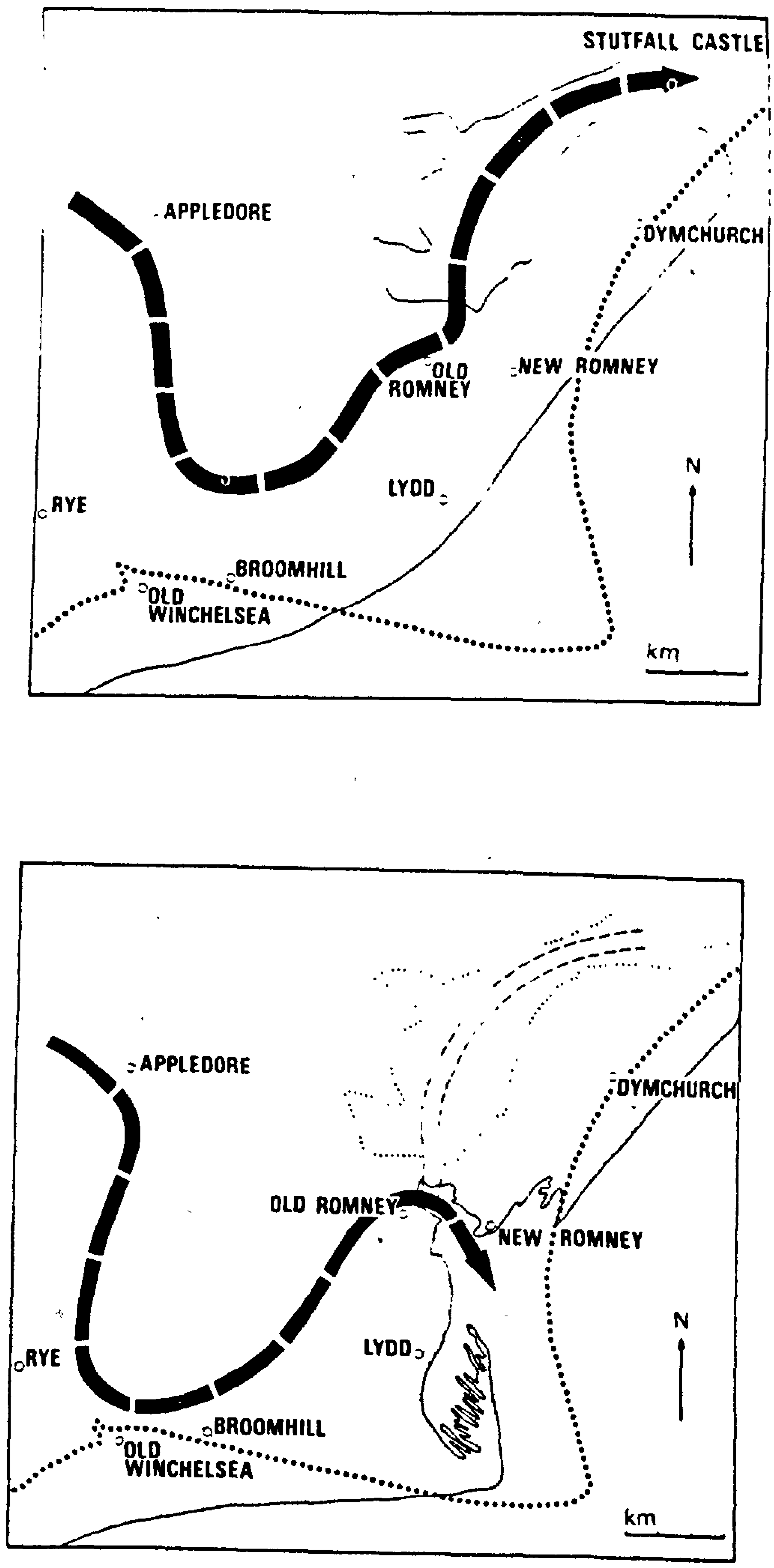

Figure 7.4: Suggested location of channel sediments in the region of Old Romney.

Source: Green (1988).

stratigraphic transects I and II. The environments of deposition of the various sedimentary units recorded in the Romney Marsh region have been determined, both in this study and in a number of previous works. The successive environments of deposition are summarised as follows.

7.3.1 The Basal Sands.

The lowermost sedimentary unit is a sand recorded across much of the marsh. Innes and Long (1992) and Long and Innes (1993; 1995b) have demonstrated that, at the typesite of Midley the sand outcrop is a relatively young deposit on the marsh and is, thus, neither part of an initial barrier nor the same sedimentary unit as the basal sand as had been proposed by Green (1968).

Detailed palaeoenvironmental analysis of these early Holocene basal sands in Romney Marsh has been relatively sparse. However, Waller *et al.* (1988) have investigated these sediments in Pett Level and have demonstrated, from foraminiferal evidence, that the sediments where deposited under a marine influence. In addition, Long *et al.* (1996) completed a study on a deep borehole reaching these sediments between altitudes of -25m and -12m OD near Rye. A coarsening upwards sequence was recorded, and it was proposed that the most likely explanation for this rapid influx of sand, was a rapid rise in relative sea-level that occurred after ca. 7900 cal. yrs. BP (Long and Innes, 1993). Long *et al.* (1996) suggest that this would have been accompanied by a landward migration of the shoreline and deposition under intertidal to subtidal conditions. As this happened in the nearshore, water would deepen with an increase in the tidal velocity and, thus, the sediments would become progressively coarser at a particular location.

Similar sediments to those described by Long *et al.* (1996) have been encountered below -2.00m OD at many sites on Romney Marsh, for example at Brookland (Long and Innes, 1995a), Midley Church Bank (Long and Innes, 1993), Walland Marsh (Waller *et al.*, 1988), and in the present study in stratigraphic transects I and II. Thus, it appears that a period of increased energy, associated with a rise in relative sea-level, occurred across much of Romney Marsh. Long *et al.* (1996) suggest that the deposition of these sediments took place between ca. 7878-7866 or 7813-7753 and 6868-6792 cal. yrs. BP at Rye, and probably across much of Romney Marsh. From this period onward, they believe that coastal evolution in the Romney Marsh region was strongly influenced by changes in barrier dynamics, with the resulting deposition of the finer grained marsh sediments.

7.3.2 The Blue Clay.

Following the emplacement of the initial gravel barrier, a distinct change is recorded in the sediments of Romney Marsh. In many of the back-marsh areas, a change to finer grained sedimentation occurred with the deposition of the blue clay. For example, Waller *et al.* (1988) recorded this unit in the protected valleys, as do Long *et al.*

(1996) at Rye. Deposition of the blue clay appears to have occurred on an intertidal mudflat (Long and Innes, 1995a).

In the Scotney Marsh area this sedimentary unit has been demonstrated to have been deposited under tidal lag conditions on an intertidal mudflat. The upper contact of this unit tends to contain organic matter, and the diatom and pollen assemblages demonstrate that this sedimentary contact represents the transitional boundary between an upper intertidal mudflat and saltmarsh.

7.3.3 The Main Marsh Peat.

A peat unit has been recorded over wide areas of the back-barrier environment (Waller *et al.*, 1988; Tooley and Switsur, 1988; Long and Innes, 1993; 1995a&b, Long *et al.*, 1996). Some distinct differences are noted within these organic sediments from one site to another, which will be outlined in detail below. However, some strong spatial similarities are also recorded.

The palaeoenvironmental reconstructions completed on these organic sediments have demonstrated that a progressive removal and a subsequent progressive return of the marine influence has occurred. This is demonstrated in the vegetational successions, recording a transition of environments as the marine influence was, first, gradually removed, *i.e.* saltmarsh, freshwater reedswamp, sedge fen, alder carr and in some areas acid bog. Eventually, a return of the marine influence is recorded in the organic sediments. This appears to have been relatively rapid, although a reversal of the vegetational succession is recorded at most sites. The vegetational succession that indicates an increase in the marine influence in Romney Marsh tends to be an increase in the aquatic taxa followed by the development of saltmarsh conditions, e.g. Long and Innes (1993; 1995a&b). This demonstrates an increase in the height of the water table followed by an increase in its salinity (Long and Innes, 1993). A sharp, eroded upper contact is recorded in the stratigraphy across wide areas of Walland Marsh (Long *et al.*, 1996), and also Scotney Marsh. However, due to the gradual return of marine conditions indicated by the biostratigraphy, this erosion is suggested to be a consequence of a later event as the sea inundated the upper peat surface.

Due to the variable timing of the initiation of peat across the marsh (to be discussed below), the vegetational successions achieved different stages of development in different areas of the marsh before the return of the marine influence. Long *et al.* (1996), for example, have recorded a change from alder carr, that had persisted across much of the back- to mid-marsh, to more acid bog vegetation. A similar succession has also been recorded by Waller (1993) in Pannel Valley, and to some extent, by Long and Innes (1995a) at Brookland. Alternatively, Long and Innes (1993; 1995b) have recorded a maximum extent of vegetation succession to be alder replaced by *Betula*, *Quercus*, and *Corylus*-type and an eventual drying out of the peat surface. In the fore-marsh the maximum extent of vegetation succession was a sedge fen with nearby alder carr at Scotney Marsh, whereas, at Broomhill Tooley and Switsur (1988) recorded freshwater reedswamp with saltmarsh throughout.

Following the time of the maximum hydroseral succession (Pearsall, 1918), which may have been synchronous across the Romney Marsh, the increase in the marine influence is clearly illustrated in the vegetational successions recorded across the marsh. The continued increase in the marine influence led to the eventual inundation of the organic sediments and associated intertidal mudflat deposition. The trends described above have been recorded at a number of sites in Romney Marsh and in this study in Scotney Marsh (cores AY17, A-B27, AW63, AW-AX67).

7.3.4 The Oxidation Mottled Silts.

The deposition of this uppermost sediment across Romney Marsh appears to have been in a relatively similar environment across the entire marsh, although relatively few palaeoenvironmental indicators have been found or analysed from within it. At Scotney Marsh, this upper silty-sand to silty-clay was deposited on a brackish intertidal mudflat with significant marine inputs and a possible upward increase in the energy of the environment.

At Brookland, Long and Innes (1995a) recorded a mixed environment with marine, marine-brackish and brackish indicators present. In addition, Waller *et al.* (1988)

recorded a mixed environment, ranging from freshwater to near-shore, in the foraminiferal assemblage. Tooley and Switsur (1988), also utilising foraminifera, demonstrated that brackish conditions existed during the deposition of these upper sediments. Also, at Broomhill, Tooley (1990) examined the diatom assemblage and illustrated that a brackish to marine environment prevailed with inputs from the open sea.

7.3.5 Time / Altitude and Stratigraphic Differences.

Some critical spatial anomalies in the general stratigraphic record are observed. A distinct altitudinal gradient can be observed for the regressive contact between the intertidal silty-clay (blue clay) and the peat (units MM2 and MM3) across the marsh. This contact is at an altitude of ca. -3.30m OD at Horsemarsh Sewer (Tooley and Switsur, 1988), rising to ca. -1.00m OD at Brookland (Long and Innes, 1995a), ca. +0.20m OD at Midley (Long and Innes, 1993), and finally to ca. +0.15m OD for the relatively sheltered sub-trough and to between ca. +0.50 and +0.95m OD in the back-barrier environment in Scotney Marsh. In addition, the peat unit tends to thin towards the fore-marsh, from a thickness of up to 3.50m in the back-marsh to ca. 0.40m at the fore-marsh in Scotney Marsh.

Long and Innes (1995a) proposed that the increase in the altitude of the regressive contact from the back- to the fore-marsh was due, in part, to the fore-marsh sites, such as Scotney Marsh, being underlain by gravel that will have compacted relatively little. Similarly, the peat units at Midley are underlain by a relatively thin bluey-grey silty-clay and sands, which will have compacted relatively little compared to the thicker, and more compactable, silts of the mid- and back-marsh sites, such as those found at Brookland and Horsemarsh Sewer.

There is also a chronological gradient of the ages of the regressive contact marked by peat initiation. An age of 6164-5736 cal. yrs. BP at Horsemarsh Sewer was recorded by Tooley and Switsur (1988), whereas at Brookland, an age of 5046-4861 cal. yrs. BP was obtained by Long and Innes (1995a). In addition, an age of 4143-3896 cal. yrs. BP obtained for the regressive contact at Midley (Long and Innes, 1993). From Scotney Marsh, ages of between 4060-3700 cal. yrs. BP for the initiation of the lower peat in the sub-trough, and between 3355-2500 cal. yrs. BP for the main peat of the back-barrier environment were obtained. In addition, Tooley and Switsur (1988) recorded the age of the initiation of peat at Broomhill to be 3817-3622 cal. yrs. BP. Therefore, clearly the ages of the regressive contact become progressively younger from the back-marsh to the fore-marsh.

The increases in the altitudes of the regressive contacts and also the progressive younging of the age of this regressive contact from the back- to the fore-marsh must be considered as a critical element of marsh evolution. It is clear, therefore, that conditions conducive to the development of peat were first established

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in the back-marsh areas of Romney Marsh, then in relatively quick succession, peat-forming communities migrated out across the marsh. The fact that time had elapsed between the initiation of peat at the back-marsh, mid-marsh and fore-marsh, helps explain the increase in altitude of the regressive contact. This is because whilst peat initiation had occurred at the back-marsh, intertidal mudflat deposition continued in the mid- to fore-marsh. Consequently, by the time peat became established in these areas, the ground surface was relatively higher due to sediment accumulation on the back-barrier mudflats (figure 7.5).

This interpretation of the regressive contacts of the peat removes the necessity to interpret the variable altitude of the base of the peat as being due to sediment compaction. In addition, it is suggested that the rise in the altitude of the regressive contact of the peat between the back-marsh and the fore-marsh (a distance of ca. 13km) provides an excellent record of a rise in the MHWS between 6164-5736 cal. yrs. BP and ca. 3000 cal. yrs. BP respectively. The MHWS is demonstrated to have risen through the Holocene in figure 7.6. The new SLIPs from this study occur within the sea-level band proposed by Long *et al.* (1996). The two SLIPs that occur outside the sea-level band are from core G60, the buried open pit which explains their anomalously low altitude.

Thus, the initiation of peat within the back-barrier environment appears to have been primarily controlled by a

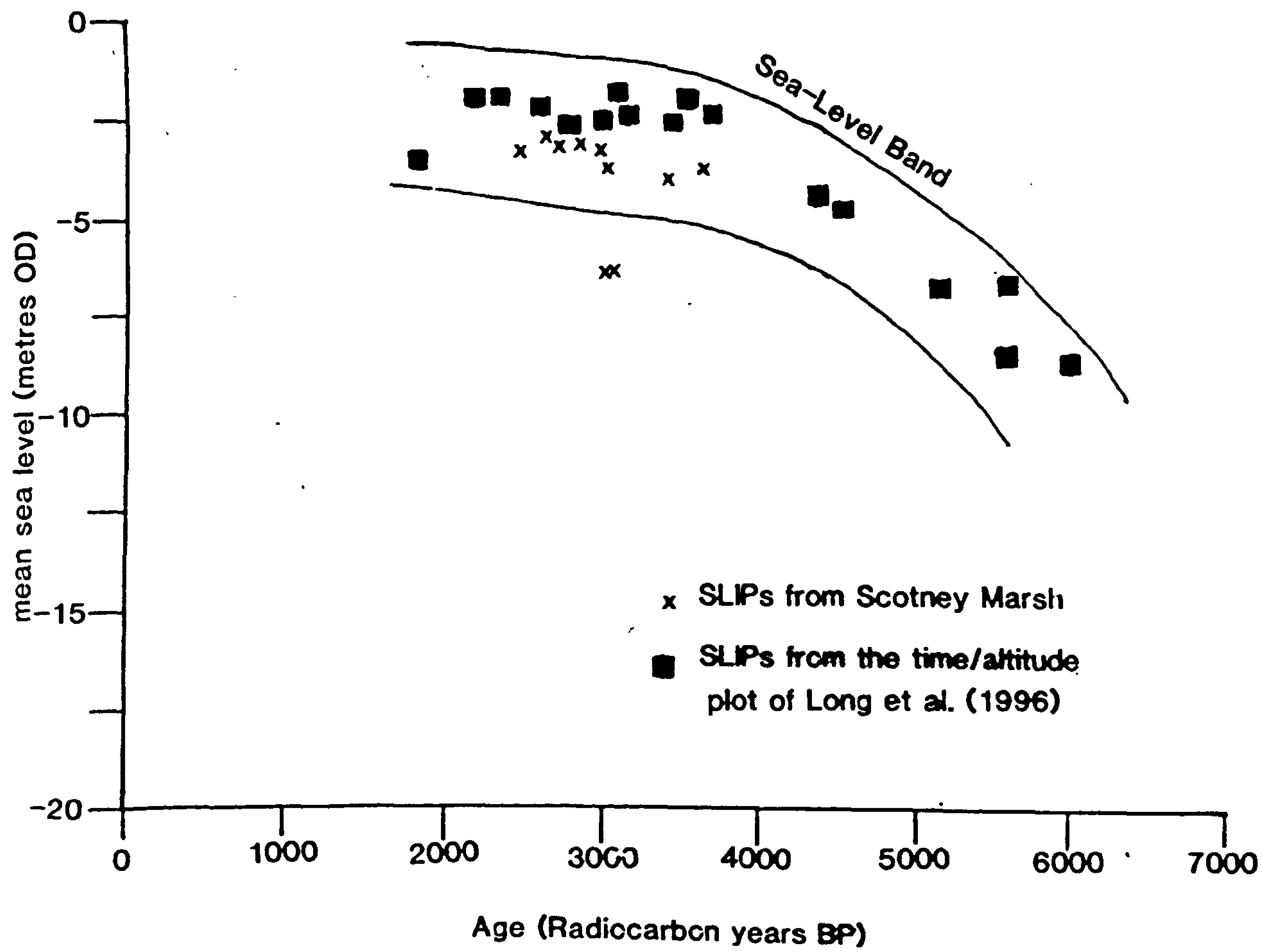

Figure 7.6: Time / Altitude graph of transgressive or regressive contacts from the Romney Marsh region, including the index points from Scotney Marsh. Each index point is corrected to mean sea-level.

combination of sea-level and sedimentation rates. Peat initiation occurred as sedimentation progressed across the marsh to a point where the sedimentary surface became sub-aerially exposed and vegetation colonised the sub-strate, occurring firstly at the back-marsh and progressing towards the fore-marsh. The regressive contact of the peat, therefore, provides an indirect record of sea-level change between 6164-5736 cal. yrs. BP (the age of peat initiation at Horsemarsh Sewer) and ca. 3000 cal. yrs. BP (the age of peat initiation of the peat at Scotney Marsh).

It is suggested above that the rate of sea-level rise and the sedimentation rate in the back-barrier environment were the controlling influences on the evolution of the back-barrier sediments. Consequently, it is possible to apply the model of back-barrier sedimentation proposed by Nichols (1989) to the back-barrier environment of Romney Marsh. Nichols (*op. cit.*) suggested that in a back-barrier lagoon in which the rate of sedimentation exceeds the rate of sea-level rise, the lagoon would tend to become infilled with sediment. Nichols termed this a surplus lagoon. In the Romney Marsh area it appears that the rate of sedimentation exceeded the rate of sea-level rise in the back-barrier environment between ca. 6000 and 3000 cal. yrs. BP, which led to the infilling of the back-barrier environment to a point at which the sediment became sub-aerially exposed and colonised by vegetation. Therefore, the back-barrier environment in the Romney Marsh area was a surplus lagoon until ca. 3000 cal. yrs. BP.

It appears, therefore, that between the time of initiation of peat at Horsemarsh Sewer (at an altitude of -3.30m OD and 6164-5736 cal. yrs. BP) and the initiation of peat at Scotney Marsh (at an altitude of between +0.52 and +0.96m OD and ca. 3000 cal. yrs. BP) that sea-level continued to rise. This demonstrates that Long and Innes (1993) are incorrect in suggesting that sea-level rose until ca. 4500 cal. yrs. BP and was then stable until ca. 1950 cal. yrs. BP, as a perceptible rise in sea-level is recorded up until ca. 3000 cal. yrs. BP. However, as table 7.1 indicates, the rate of sea-level rise appears to have been slowing throughout this time.

Location.	Time elapsed between regressive contacts.	Difference in altitude of the regressive contacts	Rate of inferred sea-level rise.
Horsemarsh Sewer to Brookland.	ca. 1000 cal. years.	2.30m	2.3mm/yr
Brookland to Midley.	ca. 1000 cal. years.	1.20m	1.2mm/yr
Midley to Scotney Marsh.	ca. 1000 cal. years.	0.80m	0.8mm/yr

Table 7.1 : Illustrating the inferred temporal variation in the rate of sea-level rise in Romney Marsh.

Further evidence of continued sea-level rise after ca. 4500 cal. yrs. BP is provided from within the gravel inlet in Scotney Marsh (cores AW63 and AW-AX67). Here, a record of peat initiation, at ca. 3700 cal. yrs. BP, continued sea-level rise and, subsequent, peat inundation is preserved. This was followed by a period of intertidal mudflat deposition, before the main period of peat deposition in the back-barrier environment in Scotney Marsh at ca. 3000 cal. yrs. BP.

The widespread increase in the marine influence recorded in the biostratigraphy in studies from across the marsh appears, therefore, to be due to an acceleration in the rate of sea-level rise. Sea-level rise is hypothesised to have been the cause of the inundation, despite the fact that the upper contacts of a many of the peat units in Walland Marsh (Long *et al.*, 1996) and Scotney Marsh are eroded, as the nature of this peat inundation is proved to be gradual from the vegetational successions.

The ages of the transgressive contacts (the end of peat deposition) across the marsh are also variable and tend to occur in two groups. The youngest group includes dates of 1938-1543 cal. yrs. BP at Old Place, in the Brede Valley (Waller *et al.*, 1988), 1890-1630 cal. yrs. BP at Brookland (Long and Innes, 1995a), and 1391-1161 cal. yrs. BP at Rye (Long *et al.*, 1996). The second, older, group includes dates of 2569-2474 to 1814-1747 cal. yrs. BP at Midley (Long and Innes, 1993) and 3006-2948 to 1814-1747 cal. yrs. BP at Broomhill (Tooley and Switsur, 1988). The ages for the transgressive contacts in Scotney Marsh also occur in this older group at 2710-2585 or 2510-2320 cal. yrs. BP (core AY17) and 2950-2725 cal. yrs. BP (core AW-AX67). The ages from Scotney Marsh are, therefore, similar those obtained from Broomhill and Midley, but significantly older than comparative ages from Brookland, Rye and Old place.

It was suggested by Long and Innes (1995a) that the observed variability in the ages of the upper peat contacts may be due to differential erosion of the peat surface. However, due to the care taken in selecting samples for analysis, they believed this to be unlikely. Consequently, Long and Innes (*op.cit.*) concluded that the end of peat deposition occurred at an earlier age at some sites than it did at others.

It is not certain what process resulted in the end of peat deposition and, indeed, different causes may account for the return to minerogenic deposition at each site. As already stated, the nature of the transgressive contact appears to be relatively similar across the marsh, with some limited reversal of the vegetational succession prior to minerogenic deposition indicating a gradual transgression. At the upper contact, however, the limited development of saltmarsh communities at many of the transgressive contacts investigated across the marsh (Long and Innes, 1995a), and also at Scotney Marsh, indicates that inundation of the peat appears to have been relatively rapid once marine conditions had returned. This rapid inundation may be accounted for by a period of rapid sea-level rise, relative to that which had been occurring previously during peat deposition. This period of sea-level rise would have led, first, to the rise in the freshwater table causing a reversal of the vegetation succession, and secondly, to the rapid inundation of the peat surface and incipient saltmarsh once sea-level had risen sufficiently. Alternatively, the peat surface at the time of transgression may have been very flat, allowing the peat to be inundated rapidly once the rate of sea-level rise exceeded the rate of peat accumulation. The evidence from the time / altitude plot, figure 7.6, is inconclusive when considering the transgression of the main marsh peat unit.

The most simple explanation is that as sea-level rose, the peats in the fore-marsh were inundated first, then mid-marsh sites, and then, lastly, the peats of the back-marsh once sea-level had risen sufficiently. However, the altitudes of the transgressive contacts of the back-marsh peats (which were inundated last) range in altitude from ca. -2.00m OD to 0.00m OD, whereas the transgressive contacts of the mid- to fore-marsh (which were inundated earlier) range in altitude from ca. +0.50m OD to +1.00m OD. Therefore, the mid- to fore-marsh sites are significantly higher than those in the back-marsh. As a consequence, it appears that sea-level rise alone cannot explain the critical differences in the ages of the transgressive contacts between back- and mid- to fore-marsh areas.

An alternative explanation for the differences in the ages of the transgressive contacts may be that as the peats in the mid- to fore-marsh areas were relatively higher than those in the back-marsh as a result of depositional regression (Curray, 1964). Consequently, sea-level would have had to overtop the mid- to fore-marsh peat before it could gain access to the back-marsh. However, as rivers appear to have drained across the marsh throughout its evolution, a rise in sea-level would have gained access to the back-marsh *via* these river channels.

Another explanation for the variability in the ages for the end of peat accumulation may be due to the locations of the tidal channels on the marsh. The exact ages (and locations) of these channels are not known but the proximity of Broomhill, Midley and Scotney Marsh to the Wainway Channel, and the similarities of the ages for the end of peat deposition at these sites, may point to flooding of the channels, driven by accelerated sea-level rise, as the cause of the transgressive overlap in these mid- to fore-marsh areas.

It appears that at some time during the late Holocene, the back-barrier environment ceased to be a surplus lagoon and became a deficit lagoon (Nichols, 1989), *i.e.* when the rate of sea-level rise is greater than the rate of sedimentation. This would account for the return of marine conditions to the back-barrier environment. In addition, the model of Nichols (*op.cit.*) also introduces the possibility that the reason for the transgressive contact may have been due to a combination of both sea-level rise and a reduction in the rate of sedimentation, relative to those rates which had prevailed as the back-barrier infilled. Indeed, the question must be asked as to why the rate of sedimentation was overtaken by the rate of sea-level rise when, during earlier periods of relatively rapid sea-level rise, the rate of sedimentation had exceeded that of sea-level rise. It is not clear whether an increase in the rate of sea-level rise or a decrease in the rate of sedimentation, or indeed a combination of the two occurred.

One explanation may be that the mode of sediment transport to the back-barrier environment had changed from that which had existed prior to peat accumulation. For example, before peat accumulation large areas of the back-barrier environment were accessible by the sea, and also by the rivers flowing into the back-barrier environment. Eventually, however, both riverine and also marine access to the back-barrier environment became confined to channels. As a consequence, the volume of water and, therefore, the supply of sediment to the back-barrier environment would have been reduced. Indeed, this confinement may have raised the prevailing flow velocities and reduced the degree of tidal lag sedimentation.

It can be seen, therefore, that questions remain unanswered as to the cause of the transgressive contacts across Romney Marsh, and that the resolution of these questions will necessitate further study. Such study will probably need to consider differential compaction rates, both in different locations across the marsh and also in different sediment types, indeed, Allen (1996) has suggested that differential compaction may present significant problems for sea-level reconstruction. However, the balance of evidence, particularly the reversals of the vegetational successions recorded across the marsh, suggests that an enhanced rate of sea-level rise provides the most likely cause of the transgressive contacts across Romney Marsh. An acceleration in the rate of sea-level rise at ca. 3000 cal. yrs. BP is, therefore, suggested to have occurred in the Romney Marsh area, this has not previously been recorded.

7.4 Evolution of the Environment Seaward of the Initial Protective Gravel Barrier.

Throughout the period of the evolution of the main back-barrier environment, behind the initial protective gravel barrier, changes were also experienced on the seaward side of this barrier. Scotney Marsh is ideally situated to study the evolution of the sediments on the seaward side of the barrier as it cuts across the boundary of the fore-marsh, the initial protective gravel barrier and the sediments of the main back-barrier environment.

Figure 7.9 illustrates that further barriers accumulated on the seaward side of the initial barrier, the Midrips and The Forelands. It can clearly be seen that the form of these barriers varies, *i.e.* the ridges of the Midrips tend to be relatively shorter than those of either the initial gravel barrier or The Forelands. It has been described by Long and Hughes (1995) how the migration of the null point, due to changes in storm incidence, results in variations in the distal extension of the gravel barriers. In addition, Long and Hughes (*op.cit.*) have illustrated how the variable position of the null point can lead to the development of open pits, such as Wickmaryholme Pit and the buried open pit recorded in the basal sediments of core G60.

It is suggested, therefore, that at some time after the deposition of the initial protective gravel barrier (ridge 1 in Scotney Marsh), further gravel barrier construction occurred with the accumulation of the Midrips. These ridges are relatively short and, perhaps, were deposited during a period of low storm incidence (Long and Hughes, 1995). Eventually, further gravel ridges, The Forelands, accumulated which had greater distal extension than the Midrips and, thus, extended around these shorter ridges. As a consequence of the alternating distal extension of the gravel ridges both the Scotney Marsh trough and the buried open pit (recorded in the base of core G60) were formed.

Initially, following the deposition of the gravel ridges of the Midrips and The Forelands, organic deposition prevailed in the buried open pit, with a removal and then a return of the marine conditions recorded by the changes in the vegetation. The deposition of this organic unit has been dated to ca. 3000 cal. yrs. BP. At this time peat had been spreading across the main back-barrier environment and had reached the fore-marsh sites, such as Broomhill (Tooley and Switsur, 1988) and Scotney Marsh. After this time, the environment of deposition became strongly marine with silts and gravels being deposited, possibly representing the temporary inundation of the site by storm overwash or barrier breaching. The environment became relatively protected from the sea following this period of minerogenic deposition as a gradual removal and then a return of marine conditions is again recorded in the vegetational succession. Importantly, the morphological situation and the vegetation recorded from the buried open pit are similar to those recorded by Long and Hughes (1995) at Wickmaryholme Pit.

Following the upper period of organic sedimentation in the buried open pit in the Scotney Marsh trough, the trough was infilled by inorganic sediments under marine intertidal conditions. Long and Hughes (1995) have described a similar environment of deposition for the South Brooks; an area of marsh between two gravel ridge populations to the south-east of Scotney Marsh. Successive gravel ridges to the south-east become younger (Lewis and Balchin, 1940; Eddison, 1983b). However, the poor chronology (due to the lack of datable organic sediment from within the successive tidal channels) make it impossible to identify whether the fine-grained sediments were deposited between each of the gravel barriers (in tidal channels) following its construction, or if, following the deposition of the gravel ridge populations, fine-grained sedimentation was synchronous in all of these tidal channels.

The only radiocarbon dates that have been obtained from the sediments outside the main back-barrier environment are from the buried open pit in the Scotney Marsh trough and from Wickmaryholme Pit (Long and Hughes, 1995). In the buried open pit the initiation of peat has been dated to 3370-2970 cal. yrs. BP, whereas at Wickmaryholme Pit, the base of the peat unit has been dated to 2038-1732 cal. yrs. BP. These two dates provide minimum ages for the emplacement of gravel at each of the sites, and, thus, provide a chronology for the minimum ages of gravel accumulation between these sites. However, as the upper contact of the peat, and the beginning of marine intertidal channel sedimentation, has not been dated in Scotney Marsh, it is not possible to determine whether or not the fine-grained inter-ridge deposition was synchronous across the area. Plater and Long (1996) have proposed that the sediments of Denge Marsh, which are outside the initial gravel barrier, bear a strong similarity to the post-peat sediments of the main back-barrier environment. This might, therefore, suggest that a period of synchronous minerogenic sedimentation occurred across the whole of Romney Marsh associated with the rise in sea-level which is demonstrated to have occurred causing the end of peat accumulation.

It can be seen, therefore, that following the emplacement of the initial gravel barrier in Scotney Marsh, gravel barriers continued to accumulate on the seaward side of the initial barrier. As a consequence of the emplacement of further gravel barriers, additional back-barrier environments were created, for example the Scotney Marsh trough and also the buried open pit.

7.5 The Evolution of Romney Marsh.

In the previous sections conclusions have been reached regarding, the site-specific evolution of the barrier / back-barrier environment at Scotney Marsh. In addition, by stratigraphically linking a number of site-specific studies from the Romney Marsh area a number of the key elements that have controlled the evolution of Romney Marsh have also been discussed. Consequently, it is now possible to present a model of the evolution of Romney Marsh.

7.5.1 Pre-barrier Sedimentation: pre- ca. 6850 cal. yrs. BP.

The lowermost sedimentary unit recorded across much of the marsh is a basal sand. These sediments were deposited under marine conditions (Waller *et al.*, 1988) and coarsen upwards which is suggested to indicate a rapidly rising sea-level and the landward migration of the shoreline with sediments deposited under intertidal to subtidal conditions (Long *et al.*, 1996).

The basal sands have been encountered below -2.00m OD at many sites on Romney Marsh, including those at Brookland (Long and Innes, 1995a), Midley Church Bank (Long and Innes, 1993), Walland Marsh (Waller *et al.*, 1988), and in the present study in stratigraphic transects I and II. Therefore, it appears that a period of increased energy, associated with a rise in relative sea-level, occurred across much of Romney Marsh. Long *et al.* (1996) suggest that the deposition of these sediments took place between ca. 7870 and 6850 cal. yrs. BP at Rye and probably over much of Romney Marsh. From this period onward, it is believed that coastal evolution in the Romney Marsh region was strongly influenced by changes in barrier dynamics.

7.5.2 Emplacement of the Initial Protective Gravel Barrier: ca. 6850 cal. yrs. BP.

It has been demonstrated by Forbes *et al.* (1991) and Orford *et al.* (1991) that, depending on sediment availability under conditions of rising sea-level, both swash and, to a lesser extent, drift aligned barriers will generally migrate landwards. This process is, therefore, suggested by Long *et al.* (1996) to be a possible cause of the accumulation of an initial protective gravel barrier across Rye Bay. The gravel was probably transported from the floor of the English Channel (Smith, 1989) with the landward advance of the swash zone, during which time the barrier would probably have been swash-aligned. Eventually this barrier will have encountered the cliffline, probably at Fairlight (Eddison, 1983a). From this time onwards, however, the predominant west to east movement of the sediment in the English Channel would have led to the north-eastward extension of the gravel barrier under a more drift-aligned regime. Indeed, Nicholls (1991) has stated that longshore transport has been critical in the redistribution of the gravel in this region during the Holocene.

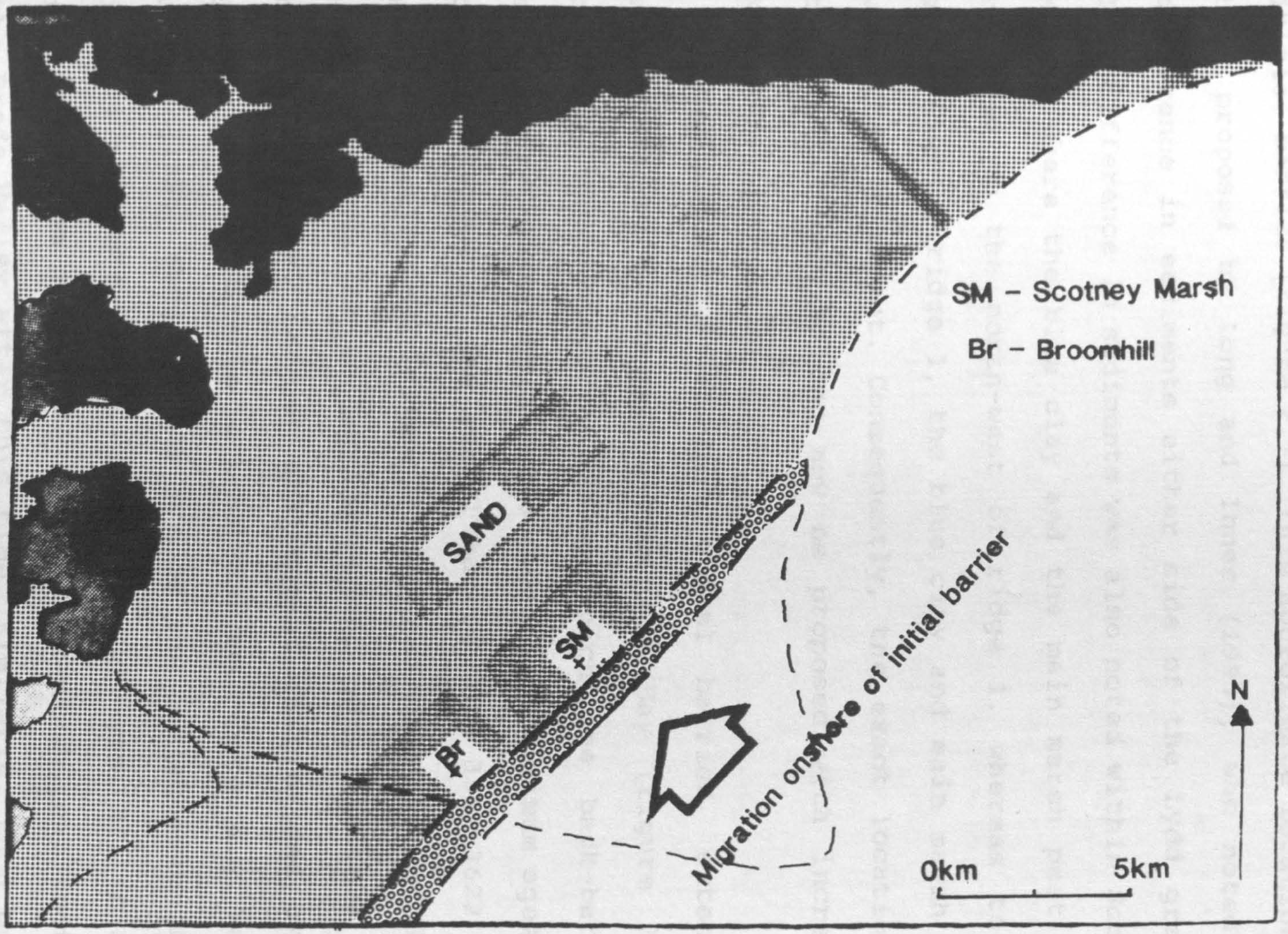

Figure 7.7: The location of the initial protective gravel barrier across Rye Bay, with sandflats prevailing in the back-barrier environment. (ca. 6850 cal. yrs. BP).

The location of the initial barrier appears to be in an approximate line which includes Fairlight, Broomhill, and runs south-west to north-east across Rye Bay to intersect Scotney Marsh, with ridge 1 being part of the initial barrier (figure 7.7). This approximate location had been proposed by Long and Innes (1993), who noted the difference in sediments either side of the Lydd gravel. This difference in sediments was also noted within Scotney Marsh, where the blue clay and the main marsh peat were recorded to the north-west of ridge 1, whereas to the south-east of ridge 1, the blue clay and main marsh peat were largely absent. Consequently, the exact location of the initial barrier can now be proposed with increased accuracy.

The deposition of this gravel barrier, extending south-west to north-east across Rye Bay (figure 7.7), provided the protection behind which the back-barrier sediments of Romney Marsh were deposited. Minimum ages for the emplacement of the gravel barrier are 3817-3622 cal. yrs. BP (Tooley and Switsur, 1988) and 4060-4040 or 3990-3700 cal. yrs. BP in Scotney Marsh. Alternatively, the timing of gravel barrier emplacement is provided by the proxy record of datable back-barrier sediments, e.g. Waller *et al.* (1988) suggests the age of the emplacement of the initial barrier to be ca. 6868-6792 cal. yrs. BP, so as to facilitate the deposition of the back-marsh peat units in the Brede Valley after this time. Similarly, Long *et al.* (1996) obtained an age of 6497-6280 cal. yrs. BP for the initiation of peat at Rye, and Long and Hughes (1995) suggested that barrier emplacement allowed peat to develop at Horsemarsh Sewer at 6164-5736 cal. yrs. BP (Tooley and Switsur, 1988).

It appears, therefore, that a swash-aligned gravel barrier was transported with rising sea-level until it encountered the cliffline, from which time it became more drift-aligned and extended across Rye Bay, through Broomhill and Scotney Marsh. Long and Innes (1995a) propose that the initiation of the barrier began between ca. 6850-5700 cal. yrs. BP.

7.5.3 Back-barrier Sedimentation: ca. 6850-5000 cal. yrs. BP.

Following the emplacement of the initial gravel barrier, a distinct change is recorded in the sediments of Romney Marsh. Across most of the back-marsh areas (Brede Valley, Rye, Brookland, Midley and Scotney Marsh) a change to fine-grained sedimentation occurred as the blue clay was deposited on a brackish to marine intertidal mudflat. At this time the back-barrier environment was a surplus lagoon, becoming infilled with sediment (Nichols, 1989). Eventually, in some of the protected valleys the sediments became sub-aerially exposed, leading to the colonisation of the inter-tidal mudflats by peat-forming communities and, thus, recording a regressive contact. This occurred first in the Brede Valley at 7180-6445 cal. yrs. BP (Waller et *al.*, 1988).

The peat-forming communities spread out from the protected valleys into the back-marsh areas, as the rate of sedimentation continued to be greater than the rate of sea-level rise. This has been demonstrated to have occurred, first, in the back-marsh areas, *i.e.* at Rye at 6503-6213 cal. yrs. BP, and then further out towards the mid-marsh areas, *i.e.* at Brookland at 5046-4861 cal. yrs. BP. A progressive younging of the regressive contact and its occurrence at a progressively higher altitudes further out on to the marsh is believed to provide a record of the rising level of the MHWS throughout the period of peat initiation. It is suggested that between ca. 6000 and 5000 cal. yrs. BP a rate of sea-level rise of ca. 2.3 mm/yr was experienced across Romney Marsh. Figure 7.8 illustrates that, at the time of peat initiation in the back-marsh, intertidal mudflat and, eventually, intertidal sandflat conditions prevailed in the mid- and fore-marsh area. In addition, at this time *i.e.* before ca. 5000 cal. yrs. BP, it is suggested that some tidal channels would have dissected the tidal mudflats, probably in the approximate location of the later Wainway Channel.

At some time following the formation of the initial protective gravel barrier, further gravel beaches began to accumulate on its seaward side. The exact timing of this is not known. However, it is likely that the gravel ridges of the Midrips began to accumulate, with relatively low distal extension, at some time before ca. 5000 cal. yrs. BP.

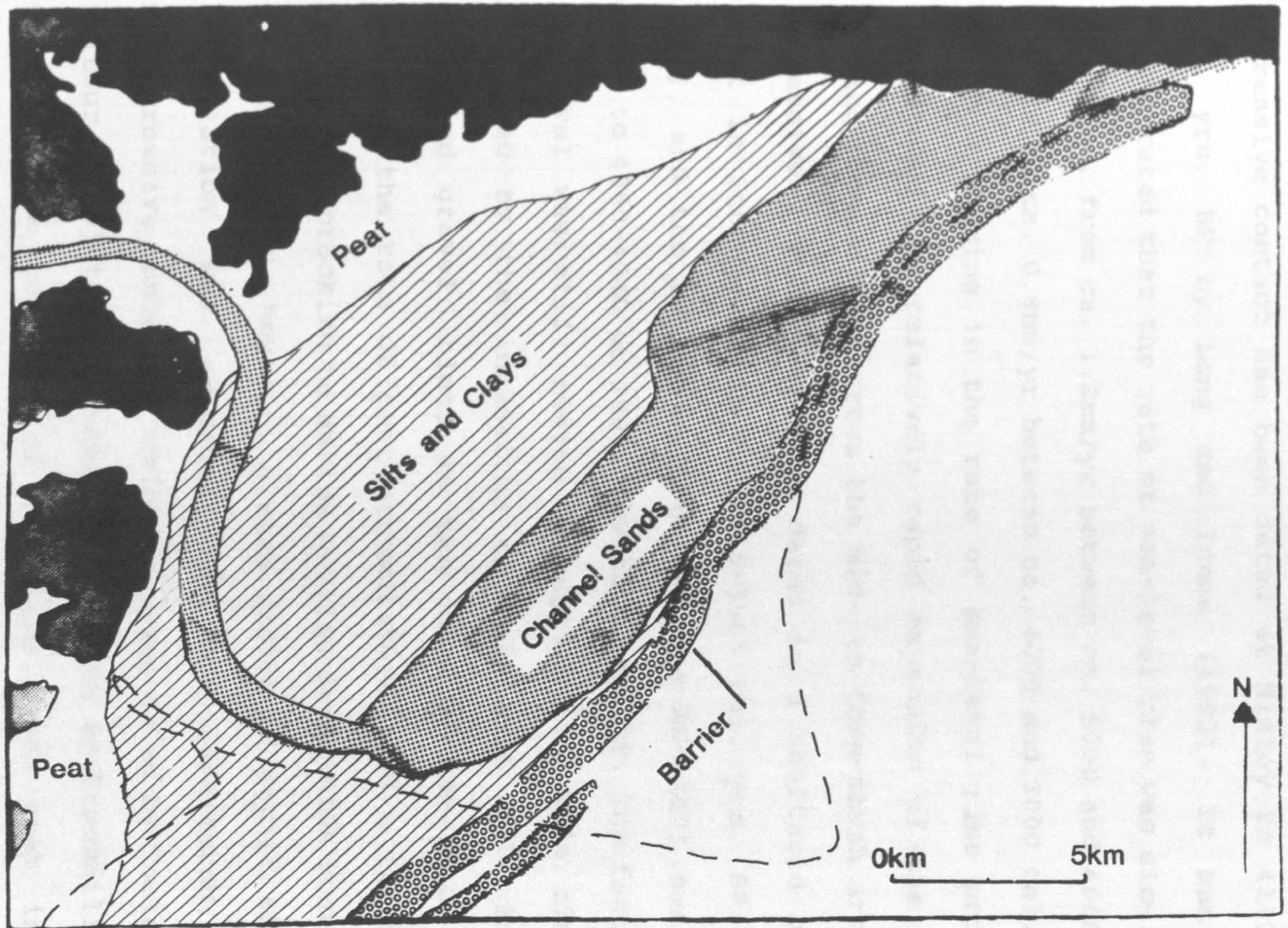

Figure 7.8: Romney Marsh at ca. 5000 cal. yrs. BP, with peat initiated in the back-marsh and tidal mudflat and eventually sandflat conditions prevailing in the mid- and fore-marshes.

7.5.4 Back-Barrier Sedimentation: ca. 5000-3000 cal. yrs. BP.

The peat-forming communities continued to spread out further across the marsh towards the mid-marsh areas. The regressive contact has been dated at Midley to 4143-3896 cal. yrs. BP by Long and Innes (1993). It has been demonstrated that the rate of sea-level rise was slowing at this time, from ca. 1.2mm/yr between ca. 5000 and 4000 cal. yrs. BP to ca. 0.8mm/yr between ca. 4000 and 3000 cal. yrs. BP. This slowing in the rate of sea-level rise probably facilitated the relatively rapid expansion of the peat-forming communities across the mid- to fore-marsh areas. A regressive contact has been dated in a sheltered gravel inlet in Scotney Marsh to 3715-3565 cal. yrs. BP (core AW63), and the entire peat unit (core AW-AX67) has been dated to 4060-4040 or 3990-3700 cal. yrs. BP. The fact that sea-level was still rising perceptibly at this time is evidenced by the inundation of this peat unit in the sheltered gravel inlet, as the rate of sea-level rise exceeded the rate of peat accumulation, and a return to inter-tidal brackish to marine mudflat deposition occurred.

Eventually, however, a more prolonged period of peat accumulation was recorded in the fore-marsh, the transgressive contact of which has been dated by Tooley and Switsur (1988) to 3817-3622 cal. yrs. BP at Broomhill. The transgressive contacts of the main marsh peat in the Scotney Marsh area have provided a similar age of 3355-2980 cal. yrs. BP (core AY17). At the barrier / back-barrier interface in Scotney Marsh, the development of the saltmarsh was replaced by coastal and then freshwater reedswamp conditions. This transition has been dated to 2795-2710 or 2620-2500 cal. yrs. BP (core A-B27). Eventually, as peat accumulated, the environment began to dry out enough in some areas for a sedge fen to develop with some standing pools, representing the maximum extent of vegetational succession in the fore-marsh.

Throughout the time that peat had been expanding across the marsh, the hydroseral succession in the back-marsh sites had continued to develop. In different areas of the marsh, different stages of maximum extent of the succession were achieved. For example, at Rye Long *et al.* (1996) recorded a change from alder domination, that had been present across much of the back- to mid-marsh (Waller *et al.*, 1988; Tooley and Switsur, 1988; Waller, 1993; 1994a; Long *et al.*, 1996) to more acidic vegetation. A similar change has also been recorded by Waller (1993) in Pannel Valley and, to some extent, by Long and Innes (1995a) at Brookland. In the mid-marsh, at Midley, Long and Innes (1993) have recorded the maximum extent of vegetation succession to be alder being replaced by *Betula, Quercus* and Corylus-type and a drying of the peat surface. Long *et al.* (1996) suggest that, for the existence of acid bog conditions, in particular, to have been present in the back-barrier environment, the barrier at this time must have been a stable feature.

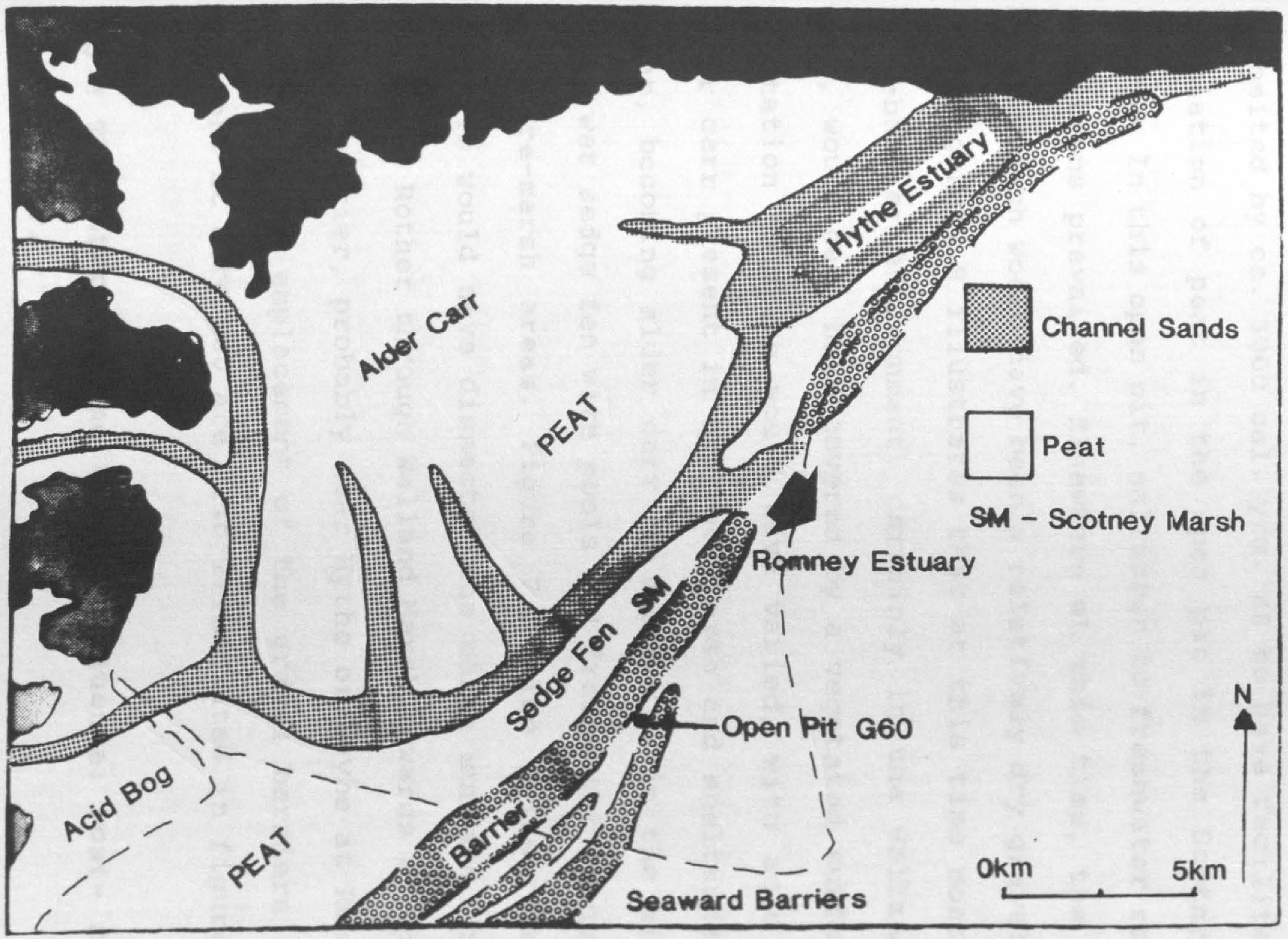

Figure 7.9: Romney Marsh at ca. 3000 cal. yrs. BP, with peat initiated in the fore-marsh and alder carr present across much of the back- to mid-marshes. Tidal channels dissect the back-barrier environment, probably with an estuary at Hythe or New Romney.

On the seaward side of the initial gravel barrier, continued gravel ridge development occurred. The relatively short gravel ridges of the Midrips, and also the first of the longer gravel ridges of The Forelands, must have been deposited by ca. 3000 cal. yrs. BP to have facilitated the initiation of peat in the open pit in the Scotney Marsh trough. In this open pit, saltmarsh to freshwater reedswamp conditions prevailed. Elsewhere at this time, the Scotney Marsh trough would have been a relatively dry gravel swale.

Figure 7.9 illustrates that at this time most of the back-barrier environment, certainly in the Walland Marsh area, would have been covered by a vegetated surface. The vegetation existing would have varied, with acid bog to alder carr present in the back-marsh and sheltered valley areas, becoming alder carr to sedge fen in the mid-marsh and a wet sedge fen with pools and freshwater reedswamp in the fore-marsh areas. Figure 7.9 also illustrates that channels would have dissected the marsh surface, carrying the River Rother through Walland Marsh towards a gap in the gravel barrier, probably near Hythe or maybe at Romney. In addition, the emplacement of the gravel barriers and the open pit at core G60 are also illustrated in figure 7.9.

7.5.5 The Return of the Marine Influence: Post- ca. 3000 cal. yrs. BP.

Following the time of the maximum extent of the hydroseral vegetation succession, which may have been synchronous across the Romney Marsh, an increase in the marine influence is clearly illustrated in the vegetational successions recorded across the marsh, by increases in the heights of the local watertable. Eventually, the increased marine presence lead to the establishment of saltmarsh conditions, which were short-lived at some sites before inundation of the organic / minerogenic interface by mudflat deposition. This transgressive sequence has been recorded at a number of sites in Romney Marsh: in Scotney Marsh (cores AY17, A-B27, AW63, AW-AX67); at Midley Church Bank (Long and Innes, 1993; 1995a); at Broomhill (Tooley and Switsur, 1988); at Brookland (Long and Innes, 1995a) and at Rye (Long *et al.*, 1996).

The widespread increase in the marine influence observed across the marsh appears, therefore, to be due to the rate of sea-level rise exceeding peat accumulation, as the back-barrier environment became a deficit lagoon (Nichols, 1989). Renewed sea-level rise, possibly coupled with a reduced sedimentation rate, is hypothesised to have been the cause of the inundation. Despite the fact that the upper contacts of a many of the peat units in Walland Marsh (Long *et al.*, 1996) and Scotney Marsh are eroded, the inundation of the peat is proved to have been gradual from the vegetational successions. It is not clear, however, what led to the later inundation of the peats in back-marsh areas.

On the seaward side of the initial protective gravel barrier, minerogenic deposition began inundating the hitherto protected organic sediments of the open pit as the tidal channels between the gravel ridges became infilled. The timing of this is not known. However, it appears likely that the deposition of these sediments is approximately synchronous with the transgressive contact recorded by the vegetation in the mid- to fore-marsh, *i.e.* ca. 2800 cal. yrs. BP.

8 Conclusions.

The conclusions of this study will be considered here in terms of the contributions made to the understanding of the evolution of Romney Marsh during the Holocene, Holocene sea-level changes in south-east England, and processes of barrier / back-barrier sedimentation. In addition, suggestions will also be made as to how future research should be most effectively directed in the light of the findings of this thesis.

8.1 Advances in the Understanding of the Evolution of Romney Marsh.

This study has resolved a number of issues regarding the evolution of Romney Marsh. First, the detailed stratigraphic analysis in parallel with the directed palaeoenvironmental reconstruction has enabled the exact location of the initial protective gravel barrier to be determined for the first time, *i.e.* ridge 1 in Scotney Marsh. The determination of the location of this initial gravel barrier elsewhere, particularly to the west of Scotney Marsh, is an issue that needs to be resolved to further understand this key stage in the evolution of Romney Marsh and its influence on marsh sedimentation.

In addition, it has only been possible to obtain minimum ages for the emplacement of this initial gravel barrier *i.e.* 4060-4040 or 3990-3700 cal. yrs. BP in core AW-AX67. However, the dating of back-barrier organic sediments has provided a proxy record for the earlier emplacement of the initial gravel barrier, *i.e.* 7180-6445 cal. yrs. BP (Waller *et al.*, 1988). The development of dating techniques that do not rely on organic sedimentary units, such as luminescence dating, is needed in order to obtain an age for the early emplacement of the initial protective gravel barrier since minerogenic sediments abut the gravel in most areas.

In this study it has been possible to reconstruct, in some detail, the evolution of the back-barrier interface in the Scotney Marsh area. By the integration of this site with the other site-specific studies completed in the Romney Marsh area it has, therefore, been possible to propose a model of the evolution of Romney Marsh. This demonstrates the necessity of such stratigraphic integration over a broad area for any future palaeoenvironmental research in the Romney Marsh area.

The completion of stratigraphic transects across Walland Marsh has also made it possible to identify with increased accuracy, the locations of areas of sediments believed to be former tidal channels (figure 7.3). Previously a number of authors (Elliot, 1862, 1874; Livett, 1930; Ward, 1952; Parkin, 1973; Green, 1988) have suggested that a southerly-arcing channel existed in Walland Marsh. However, this research was based on historical documents. Until the work of Long and Innes (1995a), no stratigraphic basis for this channel existed. The extensive stratigraphic transects of this study have demonstrated that the more southerly of the two channels, as identified by Long and Innes (1995a), extended further to the east, and appears to have be the former Wainway Channel. Conversely, the more northerly channel feature proved to be a relatively small tributary channel flowing north to south through Walland Marsh.

Until this time there has been an uneven spatial distribution of research into the evolution of Romney Marsh. Indeed, research thus far, has been focused on the Walland Marsh and Denge Marsh areas, yet little geomorphological research has been completed to the north-east of the Rhee Wall, on Romney Marsh proper. Therefore, to further resolve both the drainage of the back-barrier environment and the evolution of Romney Marsh, it is necessary that further research be focused on this area.

The attempt made in this study to determine the source of the marsh sediments has been unsuccessful due to the presence of authigenic magnetic minerals within the marsh sediments. The presence of authigenic greigite in the sediments from Scotney Marsh suggests that an alternative technique to environmental magnetism must be utilised to determine the source of the marsh sediments. Due to the fact that the two potential source areas, *i.e.* the fluvially-derived Wealden sediments and marine sediments (probably initially from a Wealden source), will be relatively similar. Hence, the resolution of the provenance of the marsh sediments may prove to be problematic. It is suggested, therefore, that attention should be directed to ascertain whether any mineralogical or geochemical differences exist between the sediments of the two potential source areas. Once this has been established a more focused investigation of the marsh sediments may be carried out.

8.2 The Relative Importance of Local Versus Regional Processes on Back-barrier Sedimentation.

During recent studies of coastal evolution in south-east England, debate has centred on whether local processes, *i.e.* barrier dynamics, or regional processes, *i.e.* sea-level change, have been the primary control on coastal sedimentation. It has been suggested by Jennings and Smyth (1982, 1985, 1987, 1990) that local processes have controlled coastal sedimentation, in particular protective gravel barriers. However, Burrin (1982) and Tooley and Switsur (1988) believe that sea-level changes have been the main controlling influence on coastal sedimentation. Alternative views have been proposed by Long (1992), who suggested that a combination of both local and regional processes have persisted, whereas Long *et al.* (1996) propose that both local and regional factors have controlled coastal sedimentation at different times during the Holocene.

This study has demonstrated the presence and stability of a protective gravel barrier has clearly exhibited a major control on coastal sedimentation as it provided a sheltered back-barrier environment in which sediments could accumulate. Equally, the rate of sediment supply to the back-barrier environment was a critical controlling influence in terms of the sedimentary response to temporal variations in the rate of sea-level rise. Hence, there is a clear combination of both local and regional factors in the evolution of the back-barrier. However, during certain periods of the evolution of Romney Marsh, different processes appear to have been the predominant controls on coastal sedimentation. For example, Long *et al.* (1996) have identified that coastal sedimentation was dominated by rapidly rising sea-level in the early-Holocene, prior to the emplacement of the protective barrier complex.

Hence, it appears that the primary control on the Holocene evolution of Romney Marsh has been sea-level, and that the sedimentary response of the back-barrier has been influenced by sediment budget. Most importantly, there is no unequivocal evidence of the influence of storms on back-barrier sedimentation, although these events will obviously have shaped the barrier and resulted in periodic breaching, and consequently, will have influenced the pattern of marsh drainage.

8.3 Reconstruction of Holocene Sea-levels.

Previous to this study, the pattern of Holocene sea-level change in the Romney Marsh region had been studied by a number of authors (Tooley and Switsur, 1988; Waller *et al.*, 1988; Long and Innes, 1993, 1995a&b; Long *et al.*, 1996). It has been established by Long *et al.* (1996) that between 7878-7866 or 7813-7753 cal. yrs. BP and 5269-5047 cal. yrs. BP sea-level rise was recorded across Romney Marsh, which Long and Innes (1993, 1995a) suggest was rapid (ca. 4-5mm/yr). After this time, however, Long and Innes (1993) argued that the sea-level was stable between ca. 4500 and 1950 cal. yrs. BP. Similarly, Long and Innes (1995a) proposed that sea-level was effectively stable, with a rise of ca. <0.003mm/yr^{-1}. Long *et al.* (1996) suggest merely that after 4507-4416 cal. yrs. BP the rate of sea-level rise fell.

From the present study, it is suggested that the rate of sea-level rise fell gradually, from ca. 2.3mm/yr to ca. 0.8mm/yr between ca. 6000 cal. yrs. BP, when peat was initiated at Horsemarsh Sewer (Tooley and Switsur, 1988), until sometime after ca. 3000 cal. yrs. BP, *i.e.* the time of peat initiation in Scotney Marsh. The fact that sea-level was still rising perceptibly after ca. 4000 cal. yrs. BP is evidenced by the fact that a relatively short-lived period of peat accumulation is recorded in the sheltered gravel inlet within Scotney Marsh at ca. 3800 cal. yrs. BP. This peat was eventually inundated by the continued sea-level rise before the main period of peat accumulation at the fore-marsh.

At some time after the initiation of peat at the fore-marsh in Scotney Marsh (ca. 3000 cal. yrs. BP), it appears that the rate of sea-level rise accelerated and overtook the rate of peat accumulation, the rate of sea-level rise at this time is, however, not known. As a consequence, the peat was inundated. It is suggested in this study, and also by Waller *et al.* (1988), Long and Innes (1993, 1995a&b) and Long *et al.* (1996), that the return of marine conditions was somewhat gradual and, therefore, sea-level driven, due to the recorded reversal of the vegetational succession. However, inundation has been demonstrated to have occurred in two distinct stages, with mid- to fore-marsh areas inundated by ca. 2400 cal. yrs. BP, whereas, back-marsh sites were not inundated until after ca. 1900 cal. yrs. BP, the reason for which remains unclear.

It is now possible, due to the improved resolution of the locations of the tidal channels, such as the Wainway Channel, to propose that the proximity of mid- to fore-marsh sites to the tidal channel would have led to the inundation of peats in these sites as sea-level rise resulted in flooding of these tidal channels. To resolve the cause of the return of marine conditions to the back-marsh sites further research will be necessary.

It can be seen, therefore, that an increasingly detailed framework of time / altitude changes is becoming available for the Romney Marsh region. However, little detail of sea-level changes is available for certain periods, especially the last 2000 years. It has been suggested by Long *et al.* (1996) that further resolution of sea-level changes will require the development of new dating techniques that do not necessitate organic sediments. This suggestion is endorsed by this study.

As yet, a lack of data on sea-level changes in the last 2000 years is evident for the Romney Marsh area and, indeed, for south-east England. It is suggested here that some of the organic units identified in the Scotney Marsh trough may have formed during the last 2000 years. However, considerable additional palaeoenvironmental and chronostratigraphic analysis is required to determine whether these sediments formed during this critical time period, and whether they were formed as a result of sea-level change or storm events.

8.3.1 Holocene Sea-level Changes in South-east England.

Long and Tooley (1995) have investigated changes in sea-level in south-east England during the Holocene. The conclusions of this thesis essentially re-affirm their findings. Long and Tooley (*op.cit.*) identified that differential patterns of sea-level change are recorded in south-east England. In the Thames Estuary and the East Kent Fens mean sea-level rose to ca. -5.0m OD by ca. 5000 cal. yrs. BP after which time sea-level rise slowed. Conversely, in Hampshire, East Sussex and Romney Marsh, Long and Tooley (1995) suggest that mean sea-level rose a further ca. 2.5m to ca. -2.5m OD by ca. 4000 cal. yrs. BP. Importantly, Long and Tooley (*op.cit.*) suggest that a small rate of sea-level rise occurred between ca. 4000-2000 cal. yrs. BP in all areas of south-east England. However, it has been demonstrated here that a perceptible, although slowing, rate of sea-level rise is recorded in Romney Marsh up to ca. 3000 cal. yrs. BP.

It can be seen, therefore, that this study adds detail to the conclusions of Long and Tooley (1995), but does not assist in resolving the cause of the differential spatial patterns of sea-level rise identified in south-east England.

8.4 Barrier / Back-barrier Sedimentary Evolution.

This study has contributed little to the study of barrier dynamics, except to essentially re-affirm the conclusions of Long and Innes (1995a), in which the three phase barrier model of Orford *et al.* (1991) was used to establish whether this model left a predictable pattern of sedimentary and vegetational changes in the back-barrier environment. It was suggested by Long and Innes (*op.cit.*) that it is difficult, and may be impossible, to link changes in the back-barrier to changes in the barrier, due to the inter-relationship of other processes, such as the pattern of tidal drainage and ingress, sediment supply and sea-level changes.

It has been suggested by a number of authors (e.g. Duffy *et al.*, 1989; Shaw and Forbes, 1987) that in the back-barrier environments, changes in marine / fresh water dominance can occur independent of sea-level tendency. Equally, Orford *et al.* (1991) suggest that sediment deposited behind a gravel barrier may confuse sea-level reconstruction. Conversely, Rampino and Sanders (1981) believe that the rate of sea-level rise is the most important control on coastal sedimentation in a back-barrier environment.

In this study, it has been established that the dominant control on the back-barrier sedimentation in Romney Marsh has been sea-level, as peat-forming communities became established once the rate of sedimentation overtook the rate of sea-level rise, which was gradually falling, and the sedimentary surface became sub-aerially exposed. Consequently, the back-barrier environment can be used as a reliable source of data for sea-level reconstruction.

The process controlling the end of peat accumulation in the Romney Marsh region is less well understood. However, it is demonstrated, both in this study and elsewhere (Long and Innes, 1993, 1995a&b; Long *et al.* 1996), that sea-level rose and, thus, led to the inundation of the peat units of the back-barrier environment. It is not clear, however, how the back-barrier environment evolved after the inundation of the peats. It is known that marine conditions predominated across the back-barrier environment at this time, but it is not known whether this was a result of sea-level rise alone or if this was accompanied by the breakdown of the protective gravel barrier.

It can be seen, therefore, that the rates of sea-level rise and sedimentation appear to have been the dominant controls on the initiation and subsequent inundation of the peat units. However, the barrier provided the sheltered environment in which sedimentation occurred, and also the protection necessary to facilitate peat development. Consequently, at certain times during the Holocene (especially following the inundation of the peat) barrier dynamics may have been the dominant controlling process on back-barrier sedimentation.

This study has also demonstrated that the model of Nichols (1989) effectively explains the evolution of back-barrier sediments in Romney Marsh. A surplus lagoon exists when the rate of sedimentation exceeds that of sea-level rise, and, thus, the back-barrier becomes infilled; as has been demonstrated here to have occurred in the back-barrier environment of Romney Marsh between ca. 6000 cal. yrs. BP and ca. 3000 cal. yrs. BP. Conversely, a deficit lagoon is described as a lagoon where the rate of sea-level rise exceeds the rate of sedimentation, and, thus, the lagoon deepens. This stage is recorded by the progressive inundation of the peat units by minerogenic sediments across Romney Marsh some time after ca. 3000 cal. yrs. BP.

It can be seen, therefore, that this study has contributed to a wide range of research, most comprehensively to the study of the evolution of Romney Marsh, but also to the reconstructions of sea-level changes in south-east England and the study of barrier / back-barrier processes. In terms of methodologies for Holocene sea-level reconstructions this study has demonstrated the need to combine the results of detailed palaeoenvironmental reconstructions from single cores dispersed over a wide area. Indeed, it is only through the combination of the Scotney Marsh research with the results of previous work that the clear control of sea-level and sediment supply on back-barrier evolution can be recognised. Furthermore there is unequivocal evidence here that the back-barrier can provide a true record of sea-level trends.

8.5 Executive Conclusions.

* The initial protective gravel barrier, behind which the sediments of Romney Marsh were deposited, is present in Scotney Marsh. The location of this feature to the north-east and south-west of the Scotney Marsh area is not known.

* In Scotney Marsh, at the barrier / back-barrier interface, a record of intertidal mudflat deposition succeeded by marine regression and the establishment of first saltmarsh, then coastal reedswamp is recorded at ca. 3000 cal. yrs. BP. After this time sedge fen prevailed, eventually marine dominance returned as a reversal of the vegetational succession occurred and a marine transgression was recorded at ca. 2700 to 2500 cal. yrs. BP, with a return to intertidal mudflat conditions.

* The detailed palaeoenvironmental reconstruction of the Scotney Marsh area, and the combination of these results with those from other detailed palaeoenvironmental investigations, has enabled a comprehensive model of the evolution of Romney Marsh in the Holocene to be formulated.

* A tidal channel existed in southern Walland Marsh and appears to have been the major route by which water both gained access to and exited the back-barrier environment.

* The dominant control on the evolution of the back-barrier during the Holocene was the rate of sea-level rise, with changes in the rate of sea-level rise in comparison with sediment supply facilitating the initiation and subsequent inundation of peat conditions.

* It has been demonstrated that a reliable record of sea-level trends can be resolved from sediments deposited behind a protective gravel barrier. Indeed, it has been demonstrated that in the Romney Marsh area, the rate of sea-level rise slowed from ca. 2.3mm/yr to ca. 0.8mm/yr between ca. 6000 and 3000 cal. yrs. BP, an acceleration in the rate of sea-level rise is then suggested to have occurred after ca. 3000 yrs. BP.

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APPENDIX 1

The complete data set from the initial stratigraphic investigations of the Scotney Marsh area are presented in Appendix 1. These data are presented utilising the computer package TSPPLus (SEE SECTION 3.1.2. and saved onto four diskettes. Transects are presented from east to west *i.e.* transect SM-1 in the east to transect SM136 in the west.

Directories on diskettes :-

Diskette A: -1TO09, 10TO19, 20TO29, 30TO39, 40TO49, 50TO59.

Diskette B: 60TO69, 70TO79.

Diskette C: 80TO89, 90TO99.

Diskette D: 100TO109, 110TO119, 120TO129, 130TO136.

Within each directory the stratigraphic data are presented as a series of stratigraphic transects the locations of which are illustrated in figure A1.

Stratigraphic transects within directories :-

Directory -1TO09

SM-1.TS
SM01.TS
SM02.TS
SM03.TS
SM04.TS
SM05.TS
SM06.TS
SM07.TS
SM08.TS
SM09.TS

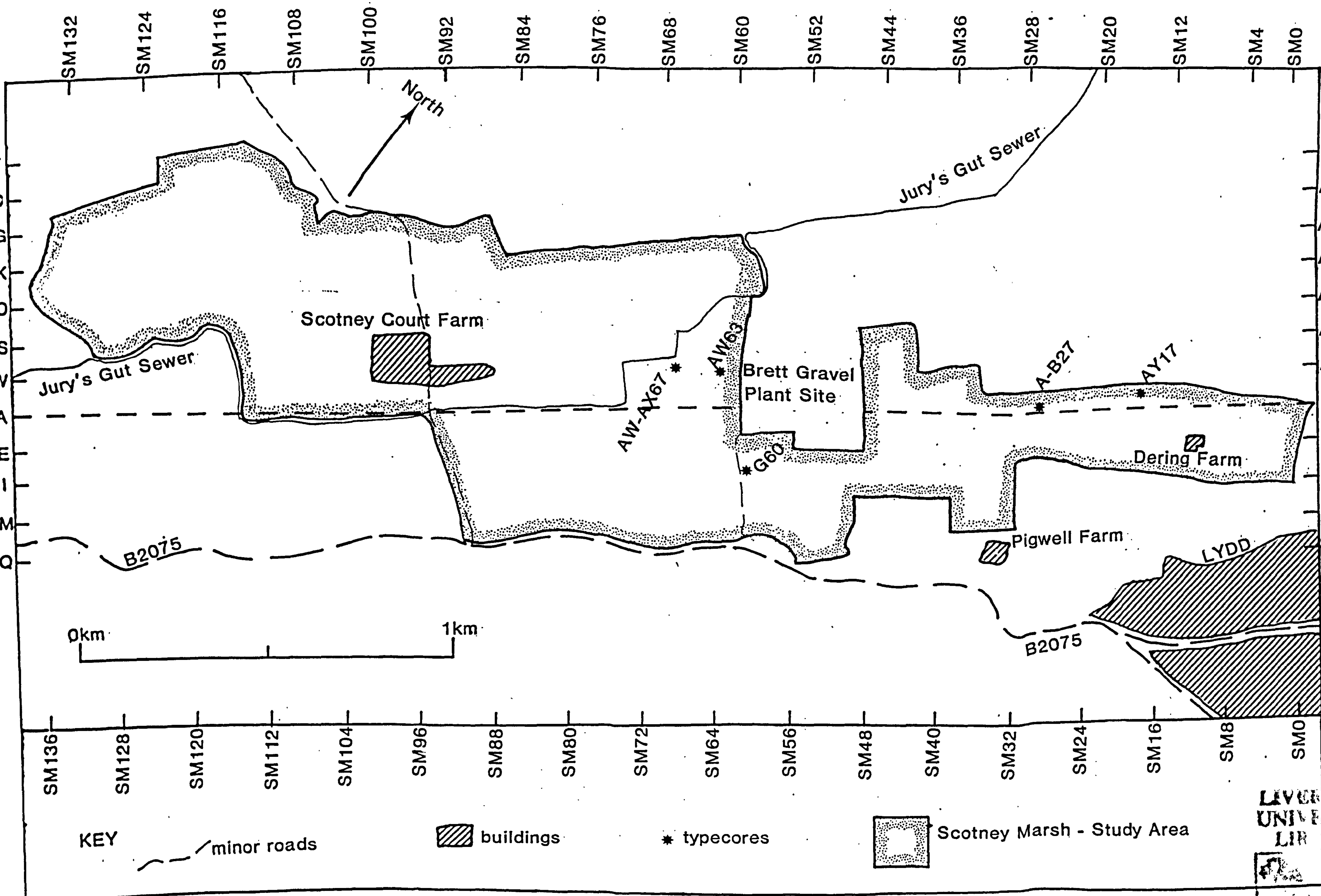

Figure A1 : Scotney Marsh study area, with the locations of the typecores and stratigraphic transects (on diskettes in Appendix A)