2. Theme 1: The Regional Reconstruction of Submerged Landscapes

The regional reconstruction of submerged landscapes, has been a desire of archaeological, geological and geomorphological researchers for at least the last 150 years. However, although during this period a multitude of reconstructions have been presented in both academic and popular literature we rarely consider the accuracy of such representations. Even more disappointingly, if inherent problems of reconstruction are identified they are very rarely propagated through the modelling process and so the reader is left with what appears to be a definitive statement of palaeo-shoreline location.

Theme 1 aims to analyse the process of submerged landscape reconstruction in order to assess and if possible quantify the potential errors. Having provided a representative review of current approaches (Section 2.1) the Theme will explore the key drivers behind the process, namely:

- The processes of sea level change (Section 2.2)
- The documentation of sea level change (Section 2.3), and
- The modelling of sea level change (Section 2.4)

This critical review will then be used to illustrate the potential errors involved with the application of a variety of sea-level sources with an equivalent variety of base levels (a term used within this report to describe various proxies for ancient land surfaces: Section 2.5). This latter section will be quantitatively and qualitatively illustrated by a number of reconstructions for the NW European continental shelf.

2.1. A Review of Existing Palaeo-geographic Reconstructions

Critical to a re-assessment of continental shelf archaeology is a review of the way in which submerged landscape reconstructions have so far been approached by archaeologists, and other Quaternary scientists. Such an examination of extant palaeo-geographic maps, and critically the approaches and modes of thinking underpinning them, is deemed necessary to assess how we currently produce and most importantly use such reconstructions to inform our archaeological interpretations and even our strategy for archaeological exploration of the shelf. Considering the ubiquitous use of such maps in a wide range of Quaternary studies we felt it essential to extend this review to include a wide range of sources looking at both archaeological and environmental reconstructions. A review was undertaken as follows:

- 85 maps have been randomly selected from a wide-ranging set of articles from across the disciplines. They have been compared qualitatively and quantitatively in order to assess the variability between different authors' interpretations of palaeo-shorelines
- A sourcing exercise has also been undertaken to determine what information has gone into the creation of these palaeo-geographic maps. This process aims to highlight the way that information regarding past sea level change is being used, or misused.

Even this limited review highlights significant variability in shoreline position between reconstructions. For example, as can be seen from Figure 6, a comparison of four reconstructions for the North-West European continental shelf at c. 12 – 10 ka

BP (uncalibrated C^{14} dates) gives reconstructed coastline positions that vary by a maximum of 250 kilometres and more typically between 20 and 30 kilometres.

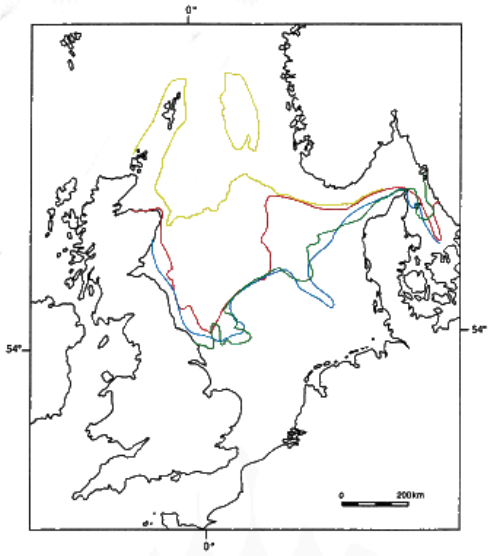


Figure 6. Comparisons of North Sea shorelines obtained from palaeo-geographic reconstructions found in archaeological publications. Yellow, = Bjerck, 1995 (c.12-11 ka BP); Red = Jonsson, 1995 (c.10-9 ka BP), Blue = Schild, 1996 (c.11-10 ka BP), Green = Newell & Constandse-Westermann, 1996 (12.8-10.3 ka BP). All of these dates are in uncalibrated C^{14} years. Modern shorelines are from Coles (1998).

Similar scales of shoreline position variability could be seen to occur between suites of papers from a global range of geographical areas. Analysis of the accompanying literature suggests a number of potential sources of these large discrepancies in coastline reconstructions. The two key sources of error appear to be the choice of sea-level data and the stratigraphic time horizon used to represent the ancient landscape. Figure 7 demonstrates the range and popularity of sea-level sources used in these reconstructions.

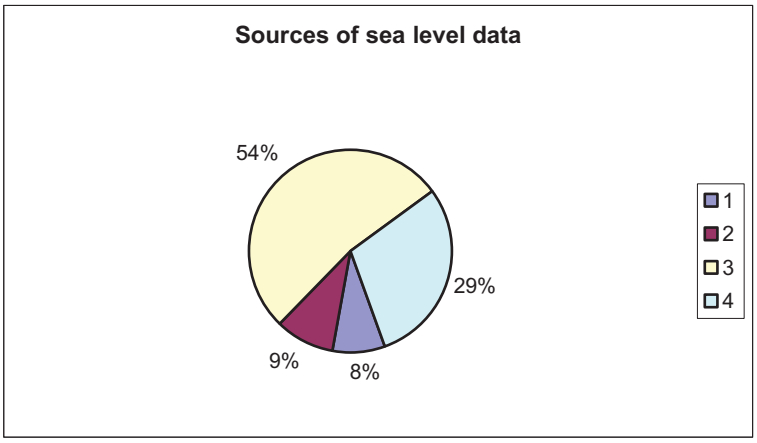


Figure 7. A pie chart to illustrate the range of information sources used in 85 palaeo-geographic reconstructions. 1) Global-eustatic data. 2) Sea Level models 3) Non mainstream sea level data source and 4) No data (n = 85).

This simple analysis suggests that only 17% of authors actually gave any indication of what sea-level data had been used for their palaeo-shoreline reconstructions. Within this group there was an effectively even split between those using a global

eustatic curve (Section 2.2.3) and those using sea level models (Section 2.4). By far the majority of the papers (54%) although alluding to the use of sea-level data in their reconstructions (as well as frequently discussing its complexities) used non mainstream sea level data. For example, Bocquet-Appel & Demars (2000: Figure 3) reference Adams & Faure (1997) a paper which deals primarily with palaeo-vegetation. Palaeo-shorelines in the source reference are dealt with only very briefly and from only a background to the main issue. Finally, and even more disappointingly, twenty-nine percent of the articles provided no indication as to where the sea level data and resulting palaeo-shoreline position had come from (e.g. Dolukhanov, 1993; Housley et al, 1997).

The second major issue identified from this analysis is the idea that the continental shelf is purely a relict landscape. This manifests itself in the selection of the modern seabed as being the most accurate representation of the past sub-aerial landscape (i.e. that the modern bathymetric contours equate to past shorelines). Several authors do mention (but do not factor into their reconstructions) that sedimentary and erosive processes are likely to have altered this surface (e.g. Coles, 1998; Shennan et al, 2000b), but rather than taking the next step and attempting to incorporate stratigraphic complexities into palaeo-geographic reconstructions they are ignored in favour of an approach which simply 'drains the landscape'.

However, it is worth pointing out that when these issues are seen in context, the use or mis-use of reconstructions is often not as dire as it seems. Frequently, an exhaustive review of the sea level change process or an exact position of the shoreline is not essential. Van Andel (1989) has for instance highlighted this very point and made use of OIS based reconstructions to draw out broad general conclusions with respect to the settlement- subsistence systems and migration of hunter gatherers since the Last Glacial Maximum. Similarly, the recent Stage 3 project used the –100m contour as a proxy for the OIS 3 shoreline (which was on average located at –80m) in its Global Circulation Models for palaeo-climate modelling (Van Andel & Davies, 2003). For higher resolution work though, more accurate shorelines are required. Hence in the case of the Stage 3 project, modelled shorelines (e.g. Lambeck, 1995) were used 'meso-scale' palaeo-climate models (Barron et al, 2003).

Therefore, while the above criticisms of existing work may seem unduly harsh, it is probably worth noting Van Andel and Tzedakis's comments in relation to the reconstructions and the nature of archaeological research:

"Ours is a historical science where…incomplete information produces tentative syntheses, which generate the inspiration for new observations that modify the existing syntheses as comprehension deepens in a circular fashion" (Van Andel & Tzedakis, 1996:481)

The reconstructions and papers examined in this study therefore constitute the 'tentative syntheses' (with respect to the study of submerged landscapes) in question. However, while this situation may have been adequate in the past, it is no longer so in the light of initiatives to take the study of submerged landscapes to the next level, that of actively studying them, managing them and integrating the archaeological resource they contain with the mainstream body of terrestrial archaeological work.

Consequently, if we are to move to the next level of submerged landscape reconstruction, it is necessary to provide both an overview of: the sea level change phenomena (Section 2.2); and how we can assess the applicability of such data to a

variety of stratigraphic horizons (Section 2.3). Particular emphasis, in this project, has been placed on quantifying variability between different models in an attempt to assess the errors in both extant and future reconstructions.

2.2 Sea Level Change: an Overview

2.2.1. Introduction

To express the sea level change phenomena in the most general terms possible, it is the superimposition of two independent movements - that of the sea surface and that of the land surface (Pirazzoli, 1996; Douglas, 2001a).

Factors influencing these movements are a diverse complex of processes that interact on a number of spatial and temporal scales. They range from astronomical factors operating from outside the Earth; such as variations in the planet's angular velocity, to factors operating on the Earth's surface; like the global distributions of glaciers and meltwater and finally, factors working within the Earth's interior; such as displacements of mantle material (Pirazzoli, 1993 and 1996; Douglas, 2001a). Many of these factors are interconnected and operate in conjunction with each other. The identification of the dominant influence on sea level change at a particular place and time depends on the temporal and spatial scale on which it is observed.

Before launching into a more detailed explanation of the processes involved, it is worth pointing out that there does exist a certain amount of confusion and ambiguity in the ideas and terminology surrounding sea level change (Mörner, 1976; Clark, 1980). Factors affecting sea level change can be classified in terms of the spatial extent of their outcomes (i.e – global versus local processes), the temporal extent of their outcomes (i.e. short term versus long term processes), or the medium in which they operate (i.e vertical movements of the sea surface versus vertical movements of the land surface).

Problems do arise over the use of these classifications. With respect to the first two categories, definitions of short and long term or global and regional may vary according to the scale and scope of each researcher's approach. For instance, although the term 'global' might be assumed to imply a uniform change affecting the entire planet, Long and Roberts (1997) define global factors as those influencing "one or more ocean basins" (1997:25). Therefore, throughout this document, for the purposes of consistency 'long term' will refer to timescales of thousands to millions of years and 'short term' to scales of days to hundreds of years. Spatially, 'global' will refer to scales of tens of thousands of kilometres or more, 'regional' to hundreds to thousands of kilometres, and 'local' to tens of kilometres or less.

Over a time scale of millions of years the dominant factors involved in global sea level fluctuations consist of plate tectonic induced changes in ocean basin geometry. As the oceanic crust spreads out from submarine ridges, it tends to thicken, increase in density and subside, thus taking the level of the water surface down with it (Pirazzoli, 1996; Lambeck & Chappell, 2001). The long term movement of continental and oceanic crustal plates can result in changes of up to several hundred metres as ocean basins are created or destroyed and expand or shrink (see Figure 8A; and section 2.2.2).

On timescales of tens of thousands of years, the periodic exchange of mass between the Earth's ice sheets and oceans as a result of glacial-interglacial cycles provides the dominant contribution. This includes both eustatic (changes in ocean volume and its

distribution) and isostatic (movement of continental and oceanic crust due to shifting loads) components (see Figure 8B; and Sections 2.2.2 and 2.2.3 respectively).

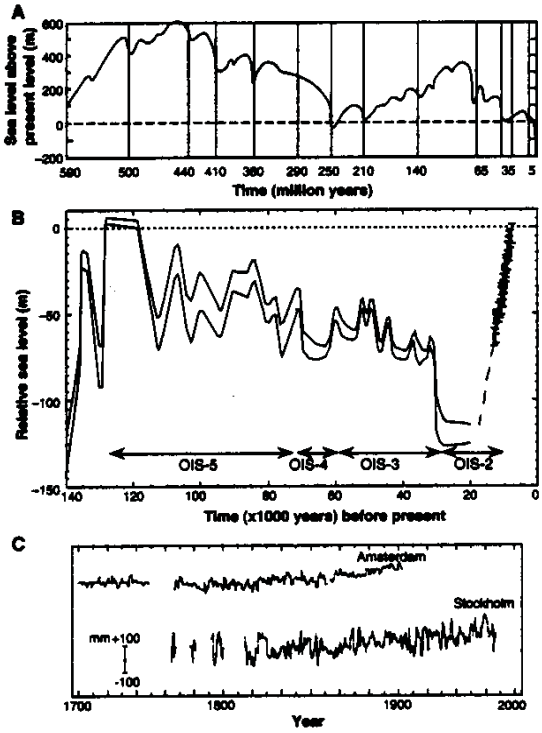


Figure 8. Sea level records illustrating change over multiple time scales: A) Tectonically induced changes affecting global ocean volume over millions of years as inferred from seismic sequence stratigraphy data. B) Glacially induced changes spanning tens of thousands of years as inferred from coral reef data from the Huon Peninsula, Papua New Guinea and northwestern Australia. C) Climatically and tidally induced changes operating on a decadal scale as inferred from historical tide gauge data from Amsterdam and Stockholm. (from Lambeck & Chappell, 2001:680)

Finally over periods of tens of years or less, oceanographic, meteorological and tidally induced changes such as air pressure, storms and water temperature become important (see Figure 8C; and Sections 2.2.3.2) (Long & Roberts, 1997; Lambeck & Chappell, 2001). These can lead to sea level changes of up to several metres in regional and local contexts.

In spatial terms, some factors are more dominant in particular areas than others. This can result in major discrepancies observable in the sea level records of different parts of the world, as Figure 9 clearly shows.

In all cases, short timescale and regional to local variations are superimposed on top of a longer term, global signature of sea level. The outcome of this variation is that an absolute global measurement of sea level change may not be applicable when studying specific regions or areas, and thus, where possible sea level change is discussed in terms of the relative sea level change affecting a particular area. In some instances though, a lack of relative sea level data means there is no recourse but to use global sea level to provide an approximation.

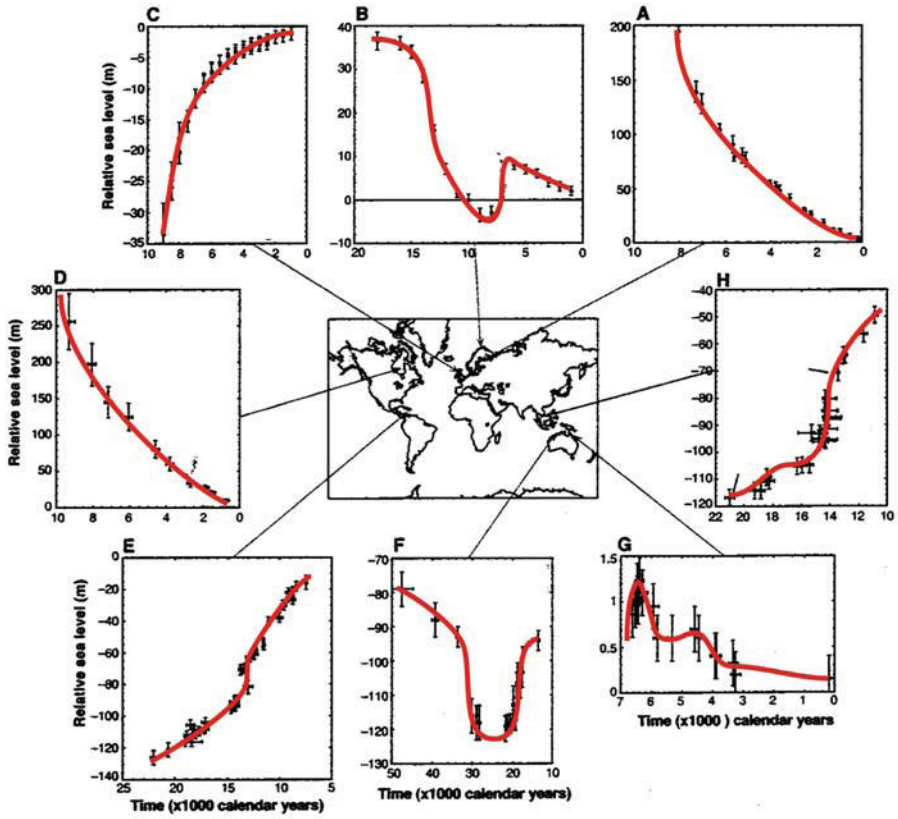


Figure 9 Relative sea level curves for a number of different areas: A) Ångerman, Gulf of Bothnia, Sweden; B) Andøya, Nordland, Norway; C) Bristol Channel, England; D) Hudson Bay, Canada; E) Barbados; F) Bonaparte Gulf, Northwest Australia; G) Orpheus Island, Queensland, Australia; H) Sunda Shelf, southeast Asia. These curves are not all drawn to the same scale. Note the range of variation in the vertical and horizontal axes (modified from Lambeck & Chappell, 2001:681)

Sea level change in relation to archaeology deals with the Pleistocene, and to some extent the end of the preceding Pliocene epoch, as it is during this period when the first artefacts and hominids appear (c. 2.5 million years ago: Klein, 1999). Throughout this time period, the interplay between variations in the Earth's orbit and axial tilt; ice sheet dynamics; and ocean circulation have resulted in a long-term (c. 41,000 years in the early Pleistocene and Pliocene and c. 100,000 years from c. 800 Kyr BP onwards) cycle of climate change consisting of alternating glacial and interglacial phases (Zachos et al, 2001; Lambeck et al, 2002a). This is observable in the records of oxygen isotope ratios of marine cores reaching far back into the Quaternary (e.g. Imbrie et al, 1984; Bassinot et al, 1994; and Figure 10). As the growth of major continental ice sheets requires significant quantities of water, evaporation from the oceans increases in glacial phases, resulting in the preferential removal of the light ^{16}O isotope, thus enriching the oceans with the heavier ^{18}O isotope. Conversely, this situation is reversed during the warming phases of interglacial phases resulting in the relative depletion of ^{18}O. This ratio can be measured by examining the remains of planktonic formanifera and calcareous nanofossils from deep-sea sediment cores as these record the chemical composition of the oceans at the time they were alive (Lambeck et al, 2002a).

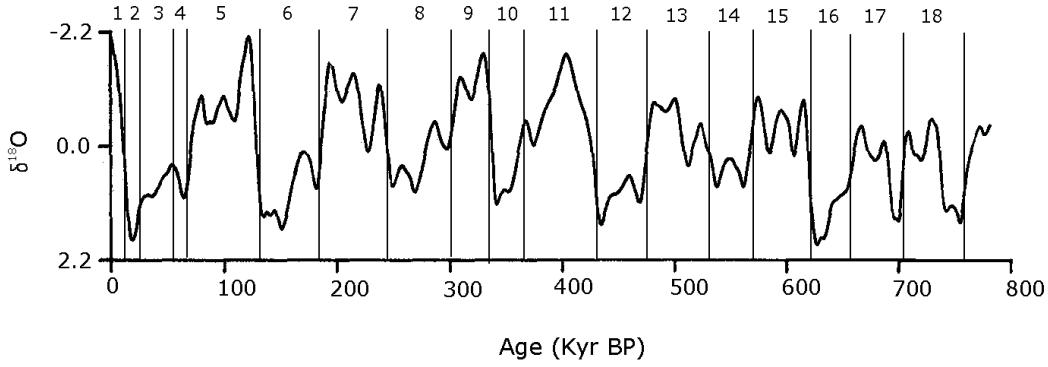


Figure 10. The oxygen isotope record from 800 Kyr BP till present. The greater the value of the $\delta^{18}O$ measurement, the greater the extent of the global ice sheets. The approximate timing of the oxygen isotope stages have been superimposed on the record. Even numbered stages represent relatively cold periods, while odd numbers represent warm stages (modified from Imbrie et al, 1984).

Factors operating as a result of these cycles lead to: variations in the volume of water in the oceans; the deformation of both ocean basins and continents; the density of ocean water; dynamic changes affecting water masses and modifications to the Earth's equipotential geoid. These variations have the potential to affect the movement of both the sea and land surfaces thus producing sea level fluctuations. Each of these factors will now be discussed individually, and the ways in which they interact to form shifting patterns of sea level change over space and time will be demonstrated in the synthesis at the end of this section (Section 2.2.6).

2.2.2 Tectonic Displacement of Oceanic and Continental Crust

Long-term tectonic movements of the portions of the crust that form the ocean basins will inevitably modify the distribution of oceanic water on the basis of the 'bathtub principle' (Leeder, 1999). Assuming that the total volume of ocean water is conserved, if the containing ocean basins grow larger or smaller due to tectonic processes, the distribution of ocean volume, and hence the height of the ocean surface, will be modified (Figure 11).

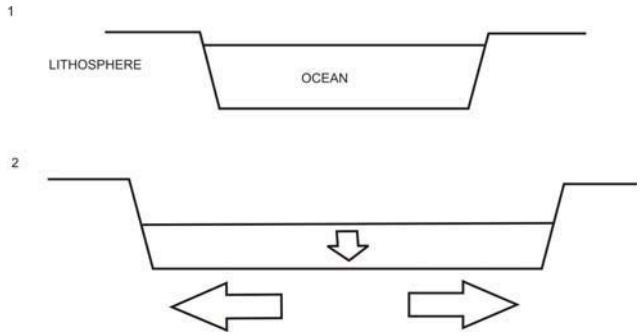


Figure 11 Schematic representation of tectono-eustasy. If water volume is conserved, the expansion of the ocean basin by tectonic movement will lead to a reduction in the height of the ocean surface as the volume of water is spread over a greater area.

Over a timescale of hundreds of millions of years, these '*tectono-eustatic'* processes create and shape the earth's ocean basins and thus represent the primary control on sea level (see Figure 8A).

Tectonic movements also have the potential to alter the height of the land surface. Localized tectonic activity can be caused by factors such as crustal plates faulting, folding or tilting thereby resulting in vertical movements of the land surface. While these vertical movements are usually gradual and continuous (a consequence of the viscous flow of mantle material over tens of thousands of years), they do sometimes take the form of rapid spasmodic events such as earthquakes, which can result in the uplift or downwarping of the land surface to a significant degree in a very short space of time (Pirazzoli, 1996). These '*coseismic'* (rapid crustal displacement occurring at the time of an earthquake) events are caused by earth movements, usually associated with crustal plate boundaries, as built up stress is rapidly released (Goudie, 2001).

In coastal areas these movements will create localized relative sea level fluctuations. For example, evidence from the John's River in Washington State (Northwest USA) points to periods of relative sea level fall punctuated by sudden relative sea level rises ranging from 0.15 to 1.5m. These have been interpreted as coseismic subsidence events. That these displacements are highly regionally variable is illustrated by evidence from Netart's Bay in the neighbouring state of Oregon which displays contemporaneous coseismic displacement of a maximum of 0.5m (Long & Shennan, 1998). More dramatic examples of coseismic uplift are observable in a number of areas, notably Crete, where an earthquake occurring at c. 450 AD resulted in uplift of 9 metres in the western part of the island (Goudie, 2001).

Long term tectonic movements are observable in a number of areas, such as the Huon Peninsula of New Guinea where flights of coral terraces dating back to some 300,000 years have been raised up to 1000 metres above present sea level (Chappell & Shackleton, 1986; Lambeck & Chappell, 2001). The speed of long term tectonic uplift varies considerably with the causes of the movement. In ancient mountain belts associated with early phases of plate collision it can be up to 5mm/yr, while in zones of current mountain building it can be up to 20mm/yr (Goudie, 2001). In fact, these movements are far more common than often realized. For example, Southern England might be assumed to be tectonically stable area. Nevertheless it has still experienced uplift of between ~0.1 and ~0.9mm/year for the past 3 million years (Maddy et al, 2001; Westaway et al, 2003). Over the course of 100,000 years this results in uplift of between 10 to 90m, clearly not an insignificant amount. In fact areas regarded as tectonically stable tend to exhibit displacements ranging from 1.1mm/year to –4.2mm/year. As a point of comparison, areas affected by isostatic factors (Section 2.2.4) move at a rate of between 15 and –7.5mm/year (Pirazzoli, 1993).

2.2.3 Eustatic Controls on Sea Level

Before going straight into a discussion of the factors that result in the addition or subtraction of water to the oceans, it is worth clarifying some of the terminology and concepts that surround this aspect of sea level change.

Addition or subtraction of water is simply one way in which ocean volume can be modified. Other factors are the distribution of water about the planet's ocean basins and the thermal expansion of water. These factors all serve to create sea level changes by raising or lowering the vertical height of the ocean surface.

Traditionally, changes in the level of the ocean surface were thought to be uniform across the globe. This was based on the ancient Greek geographer Strabo's idea that no slope could exist on the ocean surface and thus the ocean surface had to remain at the same level above the centre of the earth. Subsequently, in 1885 this hypothesis was formalised in the concept of *'eustatic'* - vertical displacements of the ocean surface occurring uniformly across the globe - sea level changes by the Austrian geologist Suess (Mörner, 1987; Pirazzoli, 1991; 1996; Douglas, 2001a).

However, in recent years the validity of this concept has been questioned. For a start, given the variation inherent in regional sea levels highlighted in Section 2.2.1, the concept of a uniform global sea level change may not be applicable to geographically restricted situations. Indeed, some researchers have argued that the very concept of a uniform global change in sea level is of rather limited value because of this variation (e.g. Lambeck, 1996) and, with respect to archaeology, because human societies live by reference to a perceivable local sea level rather than an abstract global one (e.g. Van Andel, 1989). In addition, methodological difficulties tend to prevent the elucidation of a truly global sea level record. As Stanley points out,

"[The] quest for a single stable position on the world's surface to serve as a standard from which to derive a reliable sea level curve for the late Quaternary till present remains frustrating as such a point may not exist." (Stanley, 1995:3)

In addition, satellite altimetry has revealed that the ocean surface is not in fact a level surface and does not change uniformly over time (Mörner, 1980, 1987). Some researchers therefore now see eustatic variations simply as changes in ocean level (e.g. Mörner, 1976; 1980; 1987), or effectively a change in ocean volume by addition or removal of water (e.g. Lambeck, 1995; 1996; Long & Roberts, 1997). Note the differences in the quotations below:

"The best definition of eustasy is simply 'ocean level changes' regardless of its causation and implying vertical movements of the ocean surface (Mörner, 1976:125)

"The oceanic (or eustatic) variables control the global volume of water in the ocean basins" (Long & Roberts, 1997:25)

"'Eustatic sea-level change' ... is the spatially uniform signal produced by direct mass exchange between the ice sheets and the oceans" (Milne et al, 2002:364)

"The concept of eustasy…a uniform change of sea level occurring everywhere from addition or thermal expansion of water" (Douglas, 2001a:8)

Indeed, short term dynamic sea level changes though, such as those induced by meteorological or oceanographic causes, tend to be excluded from definitions of eustatic change, despite the fact they do alter the height of the ocean surface in regional and local contexts (see section 2.2.3.2) and can be thought of as oceanic variables. This division is more the result of tradition, in that eustasy was once seen as globally uniform, rather than logic (Mörner, 1987).

As there is no strict consensus on the meanings and usage of the various terms, this paper will therefore adopt the position that eustatic changes represent variations in global ocean volume and its distribution across the earth's surface. Therefore Section 2.2.2.1 will deal with eustatic shifts induced by volume changes caused by the addition or removal of water, whilst Section 2.2.2.2 will address those caused by redistribution of ocean volume, including oceanic variables.

2.2.3.1 Controls on Volume Change in the Oceans

Minor potential inputs to ocean volume come from atmospheric water, rivers, lakes e.t.c. as part of the hydrological cycle. Table 1 gives an impression of the volumes of water stored during various stages of the present day hydrological cycle.

Water Source	Present volume (km^3)	Equivalent water depth
Biological water	700	2 mm
Rivers and channels	1700	5 mm
Swamps	3600	10 mm
Atmospheric water	13,000	36 mm
Moisture in soils and the unsaturated zone	65,000	18 cm
Lakes and reservoirs	125,000	35 cm
Ground water	$4x10^6$ to $60x10^6$	11 to 166 m
Frozen water	$32.5x10^6$	90 m
Oceans and seas	$1370x10^6$	3.8 km

Table 2. Estimates of the present day volumes of water stored in different parts of the hydrological cycle. Equivalent water depth, represents the total potential contribution to ocean volume of such sources, and is calculated using the present day ocean surface area (361.3×10^6 km^2) and the assumption that the water is evenly distributed across the globe (modified from Pirazzoli, 1996)

It is clear from this that atmospheric water, lakes, rivers, marshes, peat bogs and so on could potentially contribute relatively little to present ocean volume – a layer of water some 58.3 cm thick evenly distributed worldwide. In contrast, the potential impacts of groundwater and frozen water are in the order of tens of metres (Pirazzoli, 1996).

In warm epochs, which lack evidence for large-scale continental glaciation, such as the late Triassic, changes in lake and ground water storage in response to monsoonal fluctuations are believed to have resulted in fluctuations in global sea level of the order of several meters and thus dominate the observed eustatic sea level signal. However, in geological epochs subject to the cyclical growth and decay of massive ice sheets, such as the Pleistocene, lacustrine and groundwater fluctuations represent only a small component of the volumetric changes (Jacobs & Sahagian, 1993).

Therefore, with respect to these later periods, the single most important volumetric input is related to the cyclical growth and decay of the Earth's ice sheets (Pirazzoli, 1996; Lambeck et al, 2002a). Essentially, the growth of the ice sheets remove water from the oceans and locks it up in glaciers, thus decreasing global ocean volume. However, as the glaciers retreat, glacial meltwater enters the oceans, thus increasing their volume. For instance, the most commonly used estimation for the volume of water removed from the oceans at the Last Glacial Maximum (LGM: c. 22 Kyr BP), is a layer around 121 +/- 5 metres thick evenly distributed across the world's oceans, a volume of 437.2×10^8 km^3 (Fairbanks, 1989; Bard et al, 1990). Although as with most sea-level data this figure is currently under debate with current estimates ranging

from 116 m to 140 m (Grosswald & Hughes, 2002; Huhyrechts, 2002; Lambeck et al., 2002; Peltier, 2002; Yokoyama et al, 2000), whilst Siddall et al. (in prep) provide compelling evidence for a LGM lowstand of 125 m with uncertainty bounds between 120 and 126 m. In terms of total global ocean volume this represents a potential volume difference of 5% but as will be demonstrated in Section 2.5 this can represent a variation in palaeo-coastline position of several 10's if not 100's of kilometres. If other factors contributing to sea level change are left aside for the moment, and if meltwater volume is assumed to be evenly distributed, these '*glacio-eustatic*' or '*ice volume equivalent*' changes are not spatially variable, but are solely a function of time (Lambeck, 1995; 1996; Lambeck & Chappell, 2001). As a result, glacio-eustasy is frequently regarded as having a 'global' influence.

At this point, it should be noted that in some instances the terms glacio-eustasy and eustasy are conflated and taken to mean the same thing. Examples of this include many of the numerical models of sea level change (see section 2.4 for more detail). Note for instance the statement by Milne and co-authors at the start of Section 2.2.2 , and the quote below:

"We adhere to the conventional meaning and use the term 'eustatic' synonymously with 'meltwater' (Milne et al, 2002:364).

In addition, to those contributions from the hydrological cycle, and its variability with time, we also have to consider a number of factors which may modify the volume of water in the ocean basins and hence the overall water level. Changes in atmospheric pressure, temperature and salinity of seawater can have an impact on sea level – from millimetres up to several metres. This results from the fact that these changes alter the density of water, and thus its volume. Denser water is principally the product of lower temperatures and increased salinity and as such occupies a smaller volume of space, thus resulting in a sea level fall. The reverse is true of warmer, less saline water, which occupies a larger space. It has been calculated that a temperature variation of 1°C over a 4000m thick layer of water, or alternatively a salinity change of 4%, will produce a change in volume equivalent to 0.6m of vertical height. At present the contribution of thermal expansion to 20th century global eustatic sea level is estimated to be of the order of 2 to 7 cm (Douglas, 2001b).

Until 1960 the Mediterranean, for example, exhibited a trend of rising relative sea level of around 1 to 2 mm yr^{-1}. This trend reversed sign until 1994 when relative sea level began rising again, this time at a rate of c. 20 mm yr^{-1}. Increases in atmospheric pressure have been interpreted as causing the reversal of the trend from the 1960s to the 1990s, while the mid 1990s increase has been correlated with temperature changes in the upper waters (surface to 200m depth) of the Mediterranean (Tsimplis & Rixen, 2002). These changes though represent averages over the whole of the Mediterranean Basin, even within this relatively restricted area, local variations in sea level are apparent. Over a period of 6 years (1993-1998) satellite altimetry has revealed that relative sea levels in the Ionian Basin (east of Sicily) have been falling at a rate of 15–20mm yr^{-1}, while the Levantine Basin (south of Crete) has exhibited a rise of 25-30mm yr^{-1}. Spatial variability in sea surface temperatures is not sufficient to suggest that it is the sole driving force behind these variations, additional influences are believed to include decadal scale fluctuations of the general circulation patterns of the Mediterranean, as well as local changes in salinity (Cazanave et al, 2001; Tsimplis & Rixen, 2002).

Over the Quaternary timescale, the storage of large amounts of freshwater during glacial stages in the form of ice sheets would have increased the salinity of the oceans. During the last glaciation for instance ocean salinity increased by around 3.5%. This, in conjunction with lower temperatures, would have resulted in a minor sea level fall, probably of the order of centimetres to a few meters (Pirazzoli, 1996).

2.2.3.2 Mechanisms for the Re-distribution of Ocean Volume

On a global scale the principal mechanism by which the volume of ocean water is re-distributed is via modifications to the Earth's gravitational field. The ocean surface approximately correlates to the geoid, an equipotential surface of the Earth's gravitational field. The correlation is only approximate because influences such as waves and tides cause ocean level to deviate from the geoid on orders of up to several meters. However, if these influences were removed, the two surfaces would correspond.

The term equipotential surface can be taken to mean one in which all energy is evenly distributed. This can be thought of as a level surface, or alternatively one in which no work is required to move about it. The broad structure of the geoid follows the rough shape of the Earth; an ellipsoid of radius 6378.137 km at the equator and 6356.753 km at the poles. This is a product of the Earth's rotation, which gives the planet a slight equatorial bulge and minor flattening at the poles (PSMSL, 2003).

The geoid itself however is not a completely level ellipsoidal surface; its surface does in fact undulate, roughly correlating with the earth's topography. Therefore it follows that if the ocean surface corresponds roughly to this equipotential surface, then it too exhibits topographical variations in the form of bumps and depressions. For example, bulges in the ocean surface develop around seamounts as these large masses produce a gravitational attraction towards themselves (Mörner, 1976; Douglas, 2001a; PSMSL, 2003).

Assuming for the moment that the planet was rigid, resulting in the geoid remaining stable over time, then the concept of eustasy as a uniform global phenomenon would be valid (Figure 12A). The geoid though is a function of the Earth's gravity, which is itself a function of the planet's structure, rheology, density, rotation and astronomical gravity. Consequently, any temporal changes to these will modify the geoid and thus the distribution of ocean water rendering any concept of a globally uniform sea level invalid (Figure 12B).

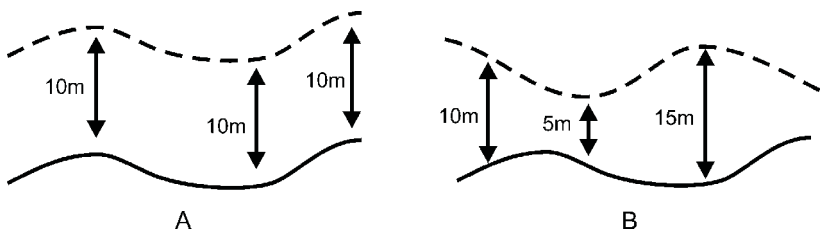


Figure 12. The solid line is the position of the geoid at time *t*, the dotted line is the position of the geoid at time t_1. A) The geoid is uneven, but stable over time, so the concept of a global uniform sea level change is valid. B) The geoid fluctuates over time, rendering the concept of a global uniform sea level change invalid. (After Mörner, 1980)

Factors that modify the geoid, and hence sea level, can be divided into two broad categories – those that affect the Earth's gravity field by modifying the planet's rotational pattern, and those that do so by modifying the gravitational attraction of materials on and within the Earth.

Modifications to the planet's rotational pattern tend to result in the movement of the entire geoidal ellipsoid. These changes are caused by large scale mass redistributions on and within the planet, such as the movement of mantle material by convection currents over the long term.

Similarly, the transfer of water between oceans and ice sheets over the glacial-interglacial cycles of the Quaternary can also have an effect on geoidal variation. The magnitude of the masses involved in these redistributions is such that they can alter the planet's rate of rotation and axial tilt (the angle of its axis of rotation). Changes to the planet's rate of rotation, will lead to geoid height varying in opposite sign about the equator and the poles. For instance, an increase in rotational rate will result in the geoid (and hence sea level) rising around the equator, and falling at the poles, while a decrease will result in the reverse. Movement of the planet's rotational axis in a lateral direction (commonly known as polar drift), modifies the geoid such that sea level at the equator remains stable, while highstands or lowstands are experienced in the northern and southern hemispheres, depending on the direction of drift (Mörner, 1980; 1986; 1987; Pirazzoli, 1996; Milne et al, 2002).

Since variations in the gravitational attractions of different materials on, and in the planet also serve to give the geoid its irregular relief, changes in their distribution of these materials will result in corresponding movements of the geoid. This can be illustrated by looking at the modification of the geoid by the changing ice masses of the Quaternary. This is observable in that during glacial phases, the gravitational attraction of the accreting ice sheets increases, thus pulling water towards them. This is observable as a relative sea level increase close to the ice sheets and a fall further away (Clark, 1980; Fjeldskaar & Kannestrom, 1980; Mörner, 1980; Lambeck, 1996; Milne et al, 2002).

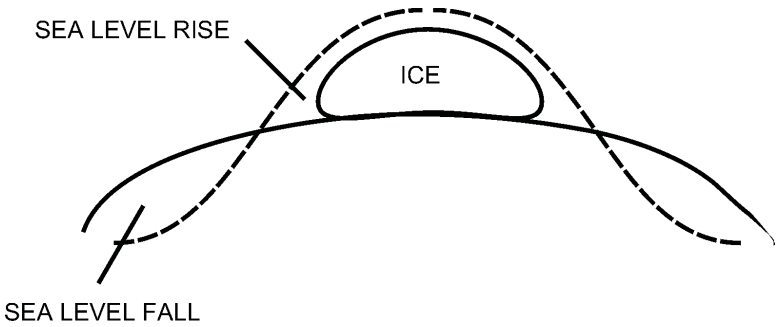


Figure 13. A schematic representation of the modification of the geoid by the gravitational attraction of an ice sheet. The solid line is the undeformed (pre-ice sheet) geoid and the dotted line represents the geoid after modification by the gravitational attraction of the ice (After Fjeldskaar & Kannestrom, 1980)

As the ice sheets melt, the reverse process occurs as their gravitational attraction decreases. Although the effects of this tend to be most significant close to the ice sheets, they should not be underestimated as it has been demonstrated that during the last glacial phase the Scandinavian ice sheet had such an effect on the waters of the

Aegean Sea, some 2000-2500km from the centre of the ice mass. The vertical displacement of the ocean surface in this instance was of the order of several metres (Lambeck, 1996).

In addition, the redistribution of materials of different densities within the planet's interior as a result of mantle flow and crustal displacement serves to modify the geoid further. This is a feedback process in that the changing distributions of ocean volume will in turn lead to differential loading of the crust and thus additional geoid deformation until equilibrium is reached (Clark, 1980). Deformations of geoid relief resulting from processes operating within the Earth can be rapid; up to 10-30mm/year (Mörner, 1986). Therefore, with respect to sea level change in relation to the glacial cycles of the Quaternary, this is a major factor that needs to be considered.

The advent of satellite altimetry measurements have provided a quantitative global perspective on geoidal variations in sea-level, with fluctuations in relief of nearly 200m and wavelength undulations of the order of several thousand kilometres having been identified (Donovan, 1979; Pirazzoli, 1996). For example, a geoid hump of c. 76m (measured with reference to the best fitting geoidal ellipsoid) exists over New Guinea, while a depression of c. -104m exists over the Maldive Islands, representing a total difference of 180 metres (Mörner, 1976).

The effect of geoidal changes on sea level, or '*geoidal-eustasy*' (Morner, 1976; 1987; Long & Roberts, 1997) is another of the ambiguous terms encountered in sea level research. Morner (1976) and Long & Roberts (1997) firmly classify it as a eustatic factor, due its influence on the ocean, while Lambeck (e.g. 1993a, b; 1995; 1996) distinguishes it from the eustatic change (which in his context of use relates solely to glacio-eustasy) and uses it in conjunction with isostatic processes on the basis that their effects vary regionally, and that it is in some measure caused by the shifting distributions of ice, water and mantle material. In this document the former position will be adopted, given the influence of the geoid on the distribution of global ocean volume.

Tides also influence the distribution of the ocean water volume and hence relative sea level on a local to regional scale (Shennan et al, 2000a, b). This is highlighted by the fact that on very small temporal and spatial scales (e.g. one beach over a period of one day) the position of the sea surface is not stable, but varies over a vertical range of several tens of centimetres to several metres. Globally, tidal ranges vary from a maximum of 18 metres in the Bay of Fundy (Canada) to less than 0.5 metres in the near tideless Mediterranean. Even within localized regions a significant degree of variation is possible, for instance Arklow, in southeast Ireland has a tidal range of 2m, while Avonmouth (west of England), on the other side of the Irish Sea has a tidal range of 14 metres (Plag et al, 1996).

Globally, tides are governed by gravitational forces in that the gravitational attraction of the Moon and the Sun pull ocean water towards them resulting in the creation of a tidal bulge. This is manifested as high tide in some areas, while the redistribution of ocean water results in a low tide in other areas. Variations in tide level may therefore take place if these gravitational forces are modified. For instance, when the position of the Sun and Moon are such that their attractive forces operate in conjunction with each other higher, spring, tidal levels will result. Conversely, when their forces operate against each other, they create lower, neap, tides. These combinations of orbital geometry occur regularly to the point where neap and spring tides occur approximately every 14 days (Davis Jr., & Fitzgerald, 2004).

On regional scales basin dimensions represent a major influence on the oceanic tidal wave as it crosses the shelf. The oceanic tidal wave in fact rotates around amphidromic points (nodes where tidal range is zero) as a result of forces generated by the Earth's rotation (the Coriolis effect). Rotation is anticlockwise in the northern hemisphere and clockwise in the southern hemisphere. On shallow shelves and in enclosed basins, such, as the North Sea, tidal waves reflect off the coastline and encounter subsequent incoming tidal waves. The waves interact and are then deflected by the Coriolis force to one side of the basin. This deflection results in water piling up on one side of the basin, thus amplifying the tidal range in that area, while on the other side of the basin, tidal ranges decreases. This can be seen in the North Sea where tidal ranges on the east coast of Britain are of the order of 3-4m, while on the Norwegian and Danish coasts have ranges of less than 1m (Open University, 1989; Davis Jr. & Fitzgerald, 2004)

Finally, on a local scale, the configuration of basin morphology will also influence tide levels to a great extent. For instance, the water level in a lagoon depends to some extent on the ratio between its surface area and the cross sectional area of its inlets. In certain situations, this ratio may lead to a lagoon experiencing a different water level to that of the open coast (Open University, 1989; Pirazzoli, 1996). This has been highlighted by investigations of tidal distortion in a salt marsh on the east coast of North America. In this instance it was found that local Mean High Water (MHW) varied within the salt marsh by up to 55cm and also potentially differed from oceanic Mean Sea Level by a similar magnitude. In addition spatial variations were also noted with respect to the flood frequency and inundation duration of the local tides. These modifications to the oceanic tidal regime have been interpreted as a consequence of local morphology, for example; the greatest tidal distortions were noted at furthest away from the inlet (Van der Molen, 1997).

Tides vary over time as well as space. In relation to sea level fluctuations, major tidal parameters such as elevation amplitude may be altered as a result of the modification of the local coastal configuration and bathymetry by transgressive or regressive events (Scourse & Austin, 1995; Shennan et al, 2000a; Shennan & Horton, 2002). For example, tidal modelling suggests that the most rapid increases in Holocene tidal range that took place in the western North Sea region were coincident with the most rapid palaeogeographical shifts resulting from sea level rise. High tide at Flamborough Head is estimated at 1.6m above mean tide level at 8 Kyr BP, 1.9m at 7 Kyr BP and 2.1m at 6 Kyr BP (Shennan et al, 2000b). In comparison, present day Mean High Water Spring Tide (MHWST) at Bridlington (Flamborough Head) is 2.55m above mean sea level. More dramatic changes have been modelled further south with the Holocene breaching of the Straits of Dover by relative sea level rise transforming the tidal regimes of both the southern North Sea and eastern English Channel from ones of low tidal amplitude (<0.5m) to states approaching the current level of tidal action; around 2m (Scourse & Austin, 1995).

Finally, dynamic factors, such as winds or currents have the potential to alter local sea levels by up to several metres by holding back or pushing forward tides. In exceptional circumstances, such as a combination of high tide, high winds and low atmospheric pressure, anomalously high sea levels or storm surges may result. For example, the North Sea storm surge of 1953 resulted in a local sea level rise of up to 3m above normal and flooded low lying areas of Britain and the Netherlands. Storm surges of up to 5m have been observed in the North Sea and the Bay of Bengal, while satellite altimetry has shown that current forces of the Gulf Stream can create sea

surface bulges of similar magnitude above the geoid. Conversely, combinations of high pressure and offshore winds may create negative storm surges, or unusually low sea levels (Mörner, 1987; Pirazzoli, 1991; 1996; Open University, 1989). Collectively, sea level changes induced by dynamic factors and changes in water density (see section 2.2.3.1) are known as 'steric' variations (Pirazzoli, 1996).

The above examples are all from very recent periods, and may not be exact representations of the situation happening in the past given the climatic variations taking place over the Quaternary. As with tidal variations, changes in overall bathymetry with changing sea level will result in a changing wave climate. Van der Molen & de Swart (2001) have modelled changes in wave climate for the Holocene, their results suggesting mean wave height has steadily increased since 7.5 ka BP (although this did assume a modern wind climate).

2.2.4 Isostatic Controls on Sea Level

Changes in both basin volume and ocean volume represent only first order approximations of the changing sea level situation. Additional, and more localized, changes are caused by the Earth's response to shifting loads – such as ice sheets and oceans – which is governed by the principle of isostasy. According to this theory, all parts of the Earth's crust float on a denser underlying layer, and are in balance, or gravitational equilibrium, with each other. Consequently, any changes in the thickness or density of the crust will alter the system such that it will try to return to equilibrium through flowage of the underlying layer, and movement of the crustal elements.

Geophysical and geological evidence suggests that this is indeed the case with the lithosphere (the outermost shell of the Earth's structure, some 100 km thick and encompassing the crust and the uppermost portion of the mantle), floating on top of a denser underlying layer – the mantle – which itself is divided into upper and lower sections on the basis of density, the lower mantle being more dense. The placement of loads on the Earth's surface therefore results in the elastic deformation of the lithosphere and flowage of the viscous mantle material beneath (Figure 14).

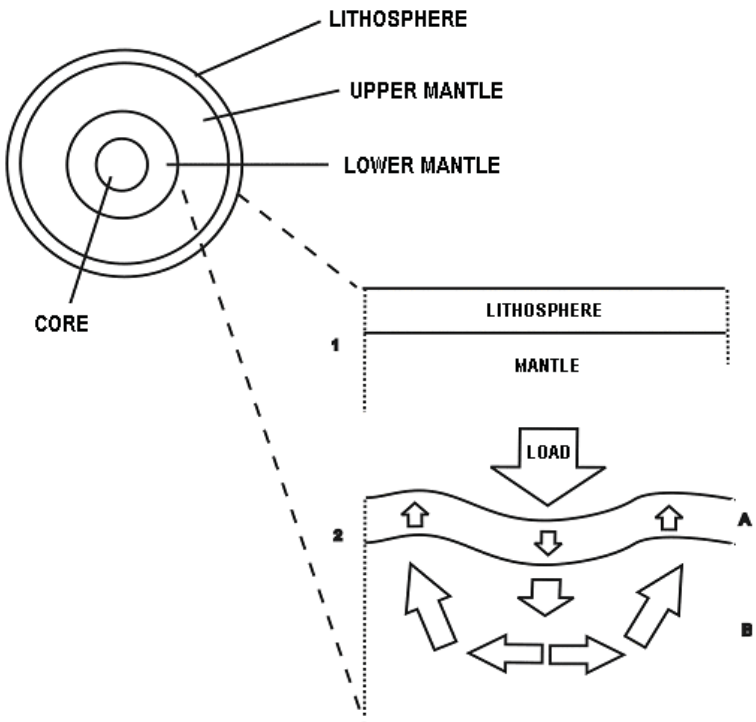


Figure 14. A schematic representation of the control of earth rheology on isostatic adjustment (not to scale). Stage 1 – Lithosphere and mantle are in a state of isostatic equilibrium. Stage 2 – Lithosphere and mantle respond to loading through A) elastic deformation of lithosphere and B) viscous flow of mantle

The lithosphere responds elastically to changing loads, such as ice sheets and oceans, on timescales of hundreds to thousands of years. This deformation is occurring almost instantaneously by comparison with the changes taking place in the denser material below it. The movement of the lithosphere displaces the mantle material below, but since this is denser than the layer above, its response takes the form of a viscous flow over a long period of time. The upshot is that deformation of the Earth's surface occurs not just at the point in time at which the loads vary, but also for thousands of years afterwards. For example, the isostatic uplift of North America and Scandinavia is only around half complete despite the fact that the ice sheets that constituted the loading factor had all but disappeared by 8 ka BP (Van Andel, 1989; Pirazzoli, 1996; Johnston, 1995; Lambeck, 1995, 1996; Lambeck & Chappell, 2001; Milne et al, 2002).

2.2.4.1 Glacio-isostasy

In relation to the glacial cycles of the Quaternary, the placement of a large ice mass on the Earth's surface leads to the subsidence of the crust directly beneath it. In tandem around the margins of the ice sheet a raised marginal rim or forebulge develops (Figure 15).

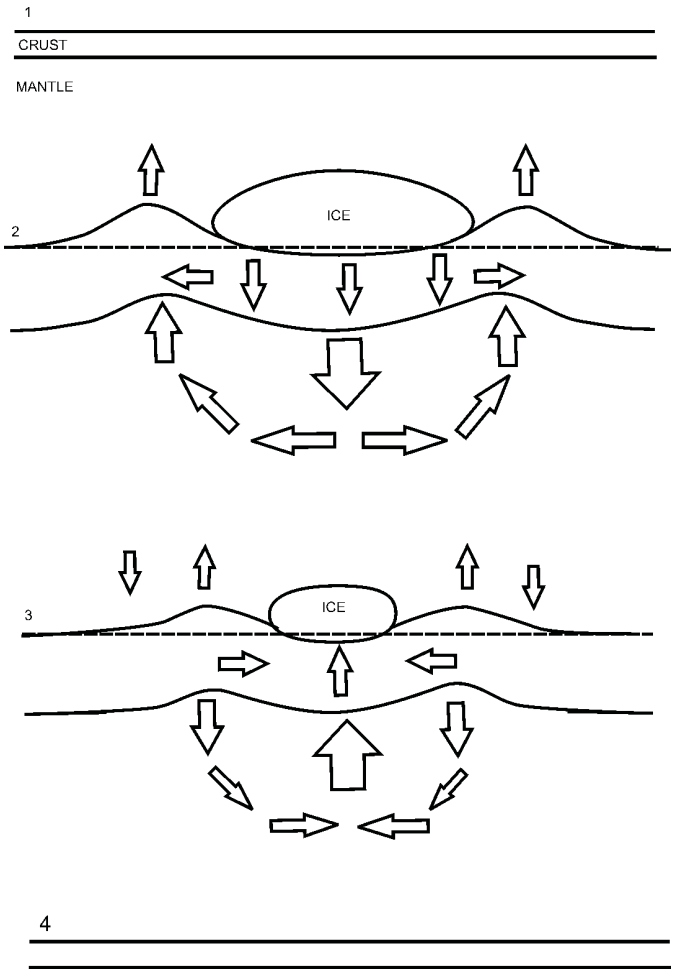


Figure 15. Schematic representation of the Earth's isostatic response to changing ice loads: 1) No ice sheet. The crust is in a state of isostatic equilibrium. 2) Growth of ice sheet. Weight of ice sheet depresses crust beneath it and leads to uplift around the ice margins through elastic deformation of the lithosphere and viscous flow of the mantle. 3) Decay of ice sheet. Reduction of load leads to crust uplifting beneath it and forebulge migrating back to centre of glaciation. 4) No ice sheet. Crust returns to a state of isostatic equilibrium. (After Pirazzoli, 1996).

The removal or retreat of an ice sheet in turn removes the load from the crust, thus resulting in the land below it uplifting and the forebulge dissipating (Figure 14). Simple models of this process predict that the forebulge should only comprise around 10 percent of the maximum rebound since mass must be conserved during and after deglaciation and the lateral extent of the forebulge is greater than that of the central rebound area (Johnston, 1995).

Isostatic movements stemming from ice sheet growth and retreat are known as '*glacio-isostatic'*. Glacio-isostatic changes vary in both time and space as a consequence of the spatial, as well as temporal distribution of the ice sheets. This, in conjunction with the other factors outlined in this document, can lead to significant differences in relative sea level change between different regions in the same time period (see Figure 9). Should land uplift exceed the glacio-eustatic rise, sea level will appear to fall and vice versa.

2.2.4.2 Hydro-isostasy

Variation in volume and distribution of water masses, for instance during cycles of glaciation or deglaciation, can also lead to uplift or subsidence of the oceanic crust, thus adding a further isostatic contribution to sea level change. As meltwater is added to the oceans, the increased weight of water will lead to the subsidence of the ocean floor in relation to the land surface. If the rate of subsidence exceeds the glacio-eustatic increase, then a regional sea level fall will take place (Johnston, 1995). Like glacio-isostasy, these '*hydro-isostatic*' processes continue to operate even when the meltwater input into the oceans has ceased due the long term viscous flow of mantle material. This is observable in Figure 9G, which illustrates that the continental shelf off Queensland, Australia has been slowly subsiding since the glacio-eustatically induced highstand of 6000 BP (Lambeck & Chappell, 2001).

Like glacio-isostasy, hydro-isostasy is spatially and temporally variable, since it depends on the fluctuating distribution of large masses, in this case, ocean water. Consequently, the hydro-isostatic effect tends to increase as one moves seaward, as the size of the meltwater load varies with topography in that shallow continental shelves will be loaded with less water than deep ocean basins and will therefore be subject to a lesser degree of subsidence (Pirazzoli, 1996) (Figure 16).

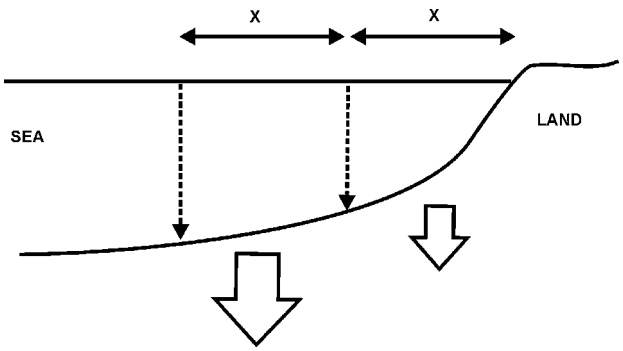


Figure 16. A schematic representation of the topographical variations in water loading effects. Close to the shoreline, water volume (and hence mass) is reduced due to the shallow topography. Further away, depth increases and therefore so does the mass of water in the water column. The increased load leads to a higher rate of subsidence in the deeper parts of the shelf (after Johnston, 1995; Pirazolli, 1996).

2.2.4.3 Other isostatic influences

Ice sheets and oceans are not the only loads that result in the isostatic adjustment of the planet. The accumulation of significant quantities of sediment over short timescales (i.e. up to hundreds of years) such as near major river deltas may lead to isostatic subsidence of the order of a few millimetres a year (*sediment-isostasy*). For example, the Netherlands is estimated to be subsiding at a rate of 2.5mm/yr as a result of the weight of sediment deposited in its delta basin (Goudie, 2001). Similar loading effects take place over longer timescales (i.e up to tens of thousands of years) as well, notably as a consequence of the deposition of sediment on exposed continental shelves during sea level lowstands (Reynolds et al, 1991; Pirazzoli, 1996)

Similarly, the extrusion of large quantities of lava by volcanoes also has the potential to load the crust to the extent that isostatic equilibrium is disrupted (*volcano-isostasy*). Hawaii, for instance, is estimated to be subsiding at a rate of 4.8mm/year.

Finally, the activation of isostatic processes does not necessarily have to involve the addition of a load onto the surface of the lithosphere; changes in the density of the crust itself will have the same effect. As the oceanic crust emerges and spreads away from submarine ridges, it cools and thickens, thus increasing in density, and begins to subside (*thermo-isostasy*). This process can also be reversed if the moving crust encounters a hot spot such as a lava source (Pirazzoli, 1991; 1996).

2.2.5 Sedimentary Controls on Sea-Level

In addition to being a driver of sediment-isostasy, sedimentary processes also have the potential to alter sea level on a local to regional scale. Assuming that the volume of water within a given basin remains constant, the deposition of significant quantities of sediment on the seabed will result in a sea level rise as the added sediment alters the geometry of the basin holding the seawater by reducing its volume (decreasing accommodation space), while erosion leads to the reverse process (increasing accommodation space: see Section 4). It has been calculated, that the amount of sediment annually removed from the United States has the potential to raise sea level by 1.5 cm every thousand years (assuming that no other influences on sea level were operating). This process is sometimes termed "*sedimento-eustasy*" (Donovan & Jones, 1979; Mörner, 1986). In reality, the effects of sedimentation on sea level tend to be mitigated by sediment-isostasy and compaction induced subsidence created by fluid escape. The former has been described in section 2.2.3.3, while the latter involves a reduction in volume of the sediment as its water content reduces due to increasing weight of the sedimentary overburden. The magnitude of compaction is significant, with a thickness reduction of an individual sedimentary unit of up to 70% being possible (Donovan & Jones, 1979; Reynolds et al, 1991; Swift & Thorn, 1991). Both these processes serve to increase the available accommodation space; or available volume within which sediment may be subaqueously deposited, which would otherwise have been decreased by sediment deposition (Leeder, 1999). In doing so, they also cause the vertical displacement of the ocean surface (assuming that water volume has remained constant).

In addition, sedimentary processes have the potential to instigate local transgressions and regressions without the need for a change in sea level. At this point it is worth stating that regressions can be defined as the seaward movement of the shoreline, and transgressions as its landward movement. These processes may be brought about by relative sea level fluctuations initiated by the factors described in the previous sections. However, the deposition of significant quantities of sediment on the continental shelf, such as at the mouth of a river delta, may result in the expansion of the coastal plain, and thus the shoreline, seawards without sea level change actually taking place. Alternatively, the erosion of shoreline sediment may result in the shoreline moving landwards. These processes are known respectively as '*progradational regression'* and *'erosional transgression'* and are distinguished from shoreline movement induced by purely sea level change (be it eustatically driven, isostatically driven, or a combination of the two) which are termed *'forced regression'* and *'forced transgression'* (Leeder, 1999). It is worth keeping in mind therefore that marine transgression and regression may involve combinations of all the aforementioned processes (See Discussion in Chapter 4).

During the Holocene in North Germany, for instance, regressive layers of peat developed where bog growth resulting from a water table rise (itself induced by the slowly rising sea level) was sufficient to compensate, or even exceed, the local rise in sea level and hence result in a temporary reversal of the movement of the shoreline (Streif, 2004 and Figure 17).

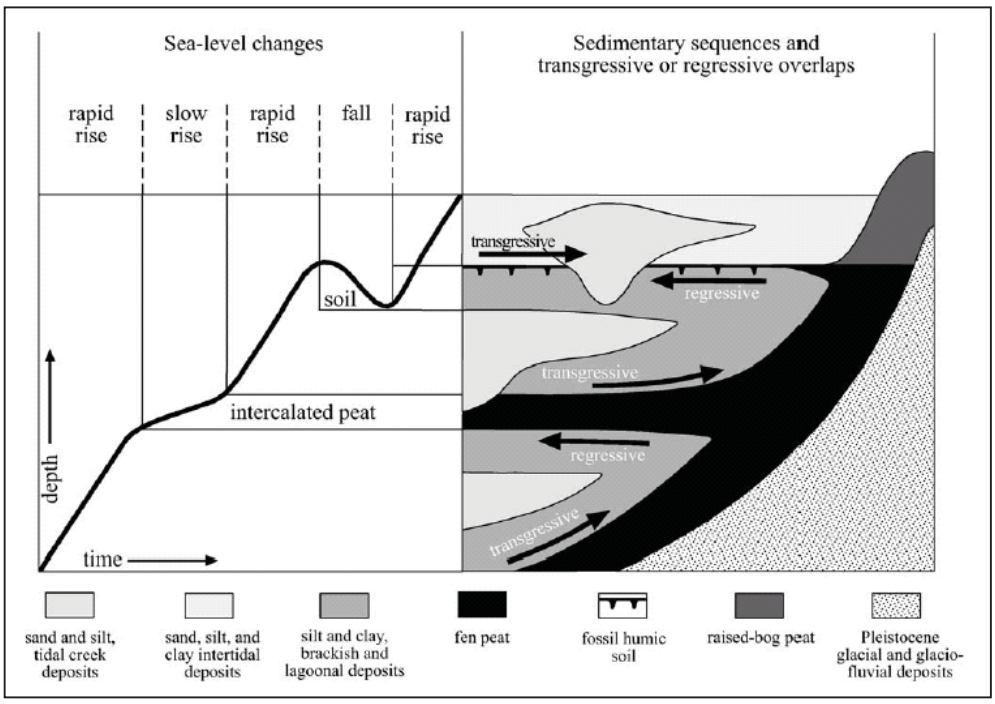


Figure 17. Diagram illustrating how progradational regression may come about under sea level rise. Note the regressive intercalated peat layer that forms during a slow rise in sea level (From Streif, 2004).

Coastal progradation takes place when the quantity of sediment supplied exceeds that removed by marine transport. This is particular common in areas receiving large quantities of fluvial sediment, such as deltas. Rates and magnitudes of delta progradation therefore vary spatially and temporally depending on the rates of the above two variables. For example, it has been calculated on the basis of present day measurements that it would take between 11,600 and 12,700 years for the Amazon River to prograde across the adjacent continental shelf (a distance of 320 km and a depth of 90m at the shelf break, creating a shelf volume of 4.75 X 10^{12} m^3). In contrast, similar calculations for the Nile reveal that it would take between 28,000 and 140,000 years to prograde across 50 km of shelf (depth of shelf break here is 250m, creating a volume of 1.25 X 10^{12} m^3: Burgess and Hovius, 1998). These are estimates, and in reality would be mitigated by sediment compaction, sediment isostasy and changes in both the marine and fluvial processes by a number of processes including sea level change.

2.2.6 A Synthesis of Sea Level Change

Most considerations of Quaternary sea level tend to ignore the smaller scale components such as steric changes, tidal changes and sedimentary processes. As a result, in areas not suffering from rapid coseismic events, the relative sea level changes for a particular place and time should at least take into account glacio-eustatic change, geoidal fluctuations, local glacio-hydro-isostatic change and local tectonics.

An example of the amalgamation of these processes can be seen in the Holocene sea level curves from the North Sea region. This region has been chosen because it has provided a great deal of sea level data and it illustrates well the range of variation possible in sea levels within a relatively restricted region (see Figure 18).

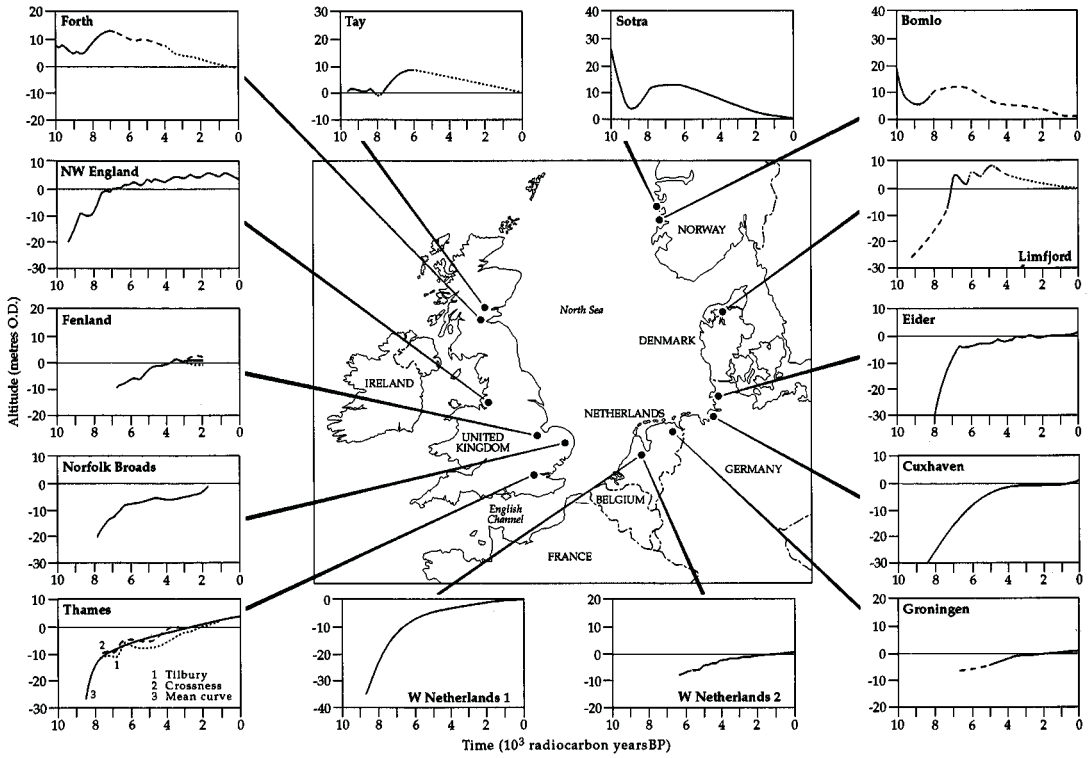


Figure 18. Holocene relative sea level curves from around the North Sea. Note the range of variation between even relatively close locations (Shennan, 1987).

The complexity in these sea level curves is principally a result of the proximity of the region to the British and Scandinavian ice sheets during glacial phases, which resulted in complex isostatic, eustatic and geoidal fluctuations (Figure 19). Although it should be noted that there is still significant debate on the dimensions and glacial/deglacial history of these ice sheets (Sejrup et al, 1998 vs Bowen et al, 2002).

Figure 19. One representation of the extent of the British and Scandinavian ice sheets at the beginning and end of the Last Glacial Maximum. Ice limits are based on multiple sources. See Lambeck (1993b) for summary. The map on the left depicts the situation at 22000 (C^{14}) yr BP, when the British ice sheet was at its maximum extent, and the map on the right depicts the onset of deglaciation at c. 18000 (C^{14}) yr BP (modified from Lambeck, 1995)

In areas under the greatest weight of ice, i.e. the central part of the glacier, the dissipation of the ice sheets tends to result in the domination of the glacio-isostatic contribution such that the land uplifts at a higher rate than the glacio-eustatically induced increase in ocean volume, resulting in a local sea level fall (Lambeck & Chappell, 2001) as exemplified by the curves in Figure 20.

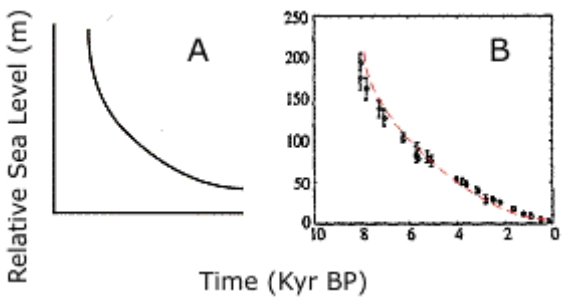


Figure 20. Sea level change near the centre of glaciation. A) Generalized curve B) Ångermanälven, Gulf of Bothnia, Sweden (modified from Lambeck et al, 1998).

Towards the margins of the ice sheets, given that the weight of ice is somewhat less than at the glacial centre, land rebound initially dominates before being overtaken by glacio-eustatic increases. Once glacial melting ceases however, the continuing process of isostatic uplift once again becomes the dominant process. The sea level curve therefore shows an initial relative sea level fall, then a sharp rise and finally a gradual fall (Figure 21).

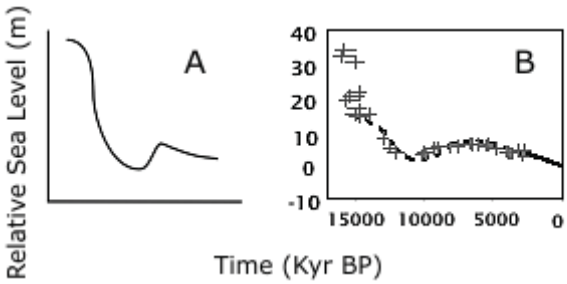


Figure 21. Sea level changes at the ice margins. A) Generalized curve. B) Arisaig, NW Scotland. Crosses represent observed sea level indicators (see section 3), line represents predicted sea level (see Section 2.4. Modified from Shennan & Horton, 2002)

Just beyond the margins of the ice sheets, bulges in the land surface develop as a result of the weight of the ice mass displacing mantle material outwards, away from the centre of glaciation (see Figure 15). Sea level change in these areas therefore takes the form of a very rapid rise, resulting from the combination of the subsidence of the forebulge due to the reduction in the weight of ice, and the glacio-eustatic increase. Once melting ceases, sea level rise continues due to the continuation of forebulge subsidence, albeit at a reduced rate (Figure 22).

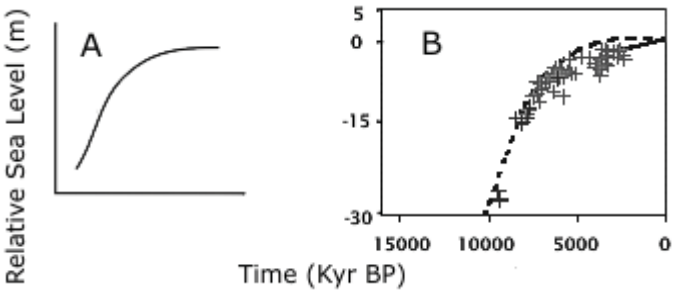


Figure 22. Sea level changes beyond the ice sheets A) Generalized curve B) Bristol Channel, SW England. Crosses represent observed sea level indicators (see section 3), line represents predicted sea level (see Section 2.4. Modified from Shennan & Horton, 2002)

Further away from the ice sheets, beyond the area of the forebulge, glacio-eustatic and hydro-isostatic changes dominate, hence the initial sea level rise tends to be rapid, but once meltwater influxes into the ocean cease, the load of water on the ocean floor pushes it down, thus leading to a very slow sea level fall (see Figure 23: Lambeck, 1996; Lambeck & Chappell, 2001).

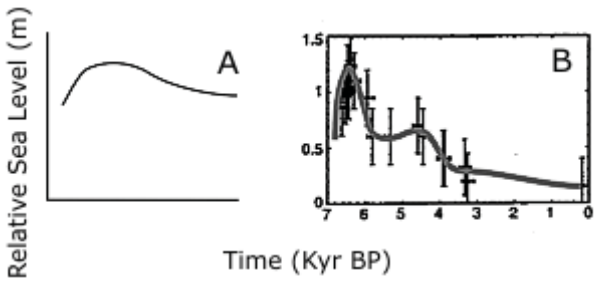


Figure 23. Sea level change far from the ice sheets. Given the distance of the North Sea region from the European ice sheets, no curves of this sort are found in this area. A) Generalized curve B) Sea level change at Orpheus Island, Queensland, Australia (modified from Lambeck & Chappell, 2001)

2.3 Documenting Past Sea Level Change

2.3.1 Introduction

A number of lines of evidence exist which make possible the documentation of past sea level changes. The most complete and highest resolution, records tend to provide estimates of local relative sea level trends since the start of the Holocene, and to some extent the post-LGM Pleistocene. This is a consequence of the fact that a large proportion of the earlier record has either been destroyed by advancing ice sheets or obscured by postglacial sea level rises (Lambeck & Chappell, 2001). To quote Pirazolli (1996:83):

"There is no way of accurately measuring sea level changes and vertical isostatic movements prior to deglaciation; in these areas estimates can only be based on assumptions and modeling or on proxy data, rather than on direct observation."

However, even the post-deglaciation sea level record suffers from a number of gaps. For instance, it is often assumed that sea level indicators are rapidly submerged or exposed by transgressive and regressive events, respectively. In reality, even when the rate of relative sea level rise was much greater than at present, such as during the Late Pleistocene, features stayed in the intertidal zone for periods ranging from decades to thousands of years. To illustrate this point, it has been calculated that assuming a rate of relative sea level rise of 160mm/yr (compared to the estimated present day rate of 1.5mm/yr) and a tidal range of 2m, a feature will remain in the intertidal zone for 13 years. At the other end of the scale, assuming a rate of 1mm/yr and a tidal range of 14m, the time spent in the intertidal zone increases to 14,000 years (Plag et al, 1996). Given that wave effects may operate between -15m below mean low water and 10m above high water during fair weather, and can extend as deep as 200m during storms, this means that the potential for erosion and destruction of the sea level record is very high indeed (Open University, 1989; Plag et al, 1996).

There are a number of exceptions to this rule, notably the coral reefs of the Huon Peninsula in Papua New Guinea, which have been tectonically uplifted beyond the reach of destructive processes associated with the onset of the last glacial cycle and consequently provide a semi continuous record of sea level change over the past 300,000 years. These and some other types of sea level indicators permit the reconstruction of broad trends in sea level far back into the Quaternary. Conversely, detailed (centimetric) analysis of a range of organic and inorganic indicators can be used to give high resolution records of sea level change on a local scale. These curves although do not necessarily provide information on process do provide a very accurate indication of the magnitude and sense of change particularly during the mid- to late-Holocene in many parts of the world.

Even when sea level indicators are available, care must be taken over their interpretation as they rarely give an exact position of sea level. Instead they tend to provide an estimated vertical range, which can vary from anywhere between several centimetres to several metres, depending on the indicator in question. Although these figures may not seem to be of particularly large magnitude, they can have significant implications for the reconstruction of past landscapes (see Section 2.5). What follows now is a synopsis of the most commonly used indicators of past sea level change, and the errors margins inherent in their use (Section 2.3.2), the critical issue of dating indicators (Section 2.3.3) and finally the most common methods of displaying sea level data (Section 2.3.4).

2.3.2. Indicators of Past Sea Level Change

2.3.2.1 Erosional geomorphological indicators

A number of geomorphic and geological features can be related to past sea level. These can take the form of erosional features created by wave, tidal and surf action. If they can be identified and dated, they will indicate the position of the palaeo-shoreline. Erosional indicators are only preserved in competent rocks and include features such as notches, benches, platforms, pools, potholes and sea caves (Plag et al, 1996; Pirazzoli, 1996). A distinction should also be made between features that have been created due to the differential erosion of weaker rock layers rather than sea level position (*structural notches*), and those that are a direct result of wave and surf action (*abrasion notches*). The accuracy of the various indicators depends on the exposure of the site, and also the type of the indicator. Tidal notches for example, are relatively precise indicators; their bottoms are situated near the lowest tide level, their vertexes near MSL and their tops where waves regularly splash at high tide. Wave abrasion of an erosional bench or platform however, may occur anywhere between the highest level reached by storm waves and the maximum depth at which wave action shifts sediment (Pirazzoli, 1996).

2.3.2.2 Coral Reefs

Coral reefs also provide a measure of past sea level fluctuations. Because of light requirements, most corals are restricted to the sub-littoral zone, with their upper limit close to mean low water spring tide. If a particular species of coral can be identified and dated, and its habitat preferences are known, then raised or submerged coral reefs can provide an estimate of past sea level that is accurate usually to within a few metres (Pirazzoli, 1996). The Caribbean species *Acropora palmata* for example, tends to be restricted to the upper 5 metres of water. Dated examples of this species, and the species *Porites asteroides*, off Barbados have provided a sea level curve going back to the Last Glacial Maximum (LGM). When corrected for local tectonic uplift, the data provided an estimate of a glacio-eustatic rise of 121 +/- 5 metres since the LGM (Fairbanks, 1989 and Figure 24). The correction for tectonic uplift is obtained by identifying a past shoreline for which mean sea level was believed to be similar to that of the present day, usually the Oxygen Isotope Stage 5e or mid-Holocene highstand, and measuring how high above the present shoreline it is. Assuming that uplift has been consistent over the intervening time period, this method provides an estimate of the rate of crustal movement. The need to include a tectonic uplift correction was necessitated in this case, because the researchers were attempting to isolate and ascertain the glacio-eustatic component of the post glacial sea level rise. Had this correction not been made, the Barbadian coral record would have only provided a measure of relative sea level change in this localized area.

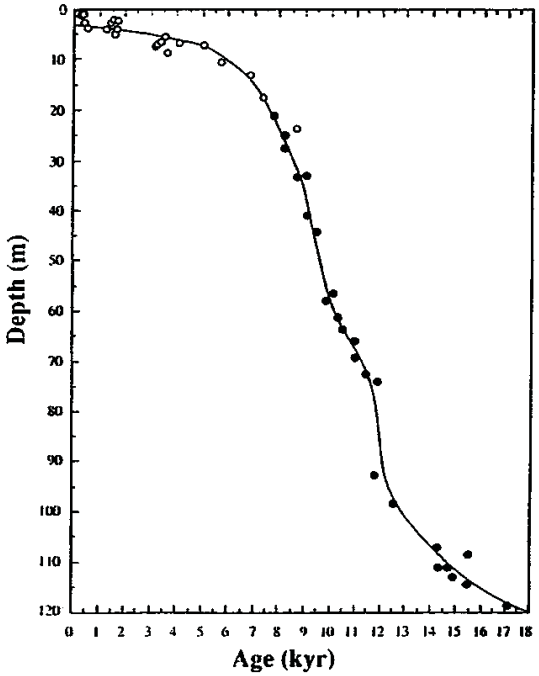


Figure 24. Sea level curve obtained from Caribbean coral reefs. Timescale is in uncalibrated ^{14}C years. The filled circles represent dated *A. palmata* samples from Barbados, and the open circles *A. palmata* from four other Caribbean islands. The data has been corrected for an estimated mean tectonic uplift of 34 cm kyr^{-1} (modified from Fairbanks, 1989).

Tectonically uplifted coral reefs on the Huon peninsula of New Guinea provide a record of sea level highstands going back some 300,000 years. The estimate of the glacio-eustatic sea level change since the LGM, based on information for this area is 130m below present sea level (Chappell & Shackleton, 1986). The 9 m discrepancy between these two, coral reef based, estimates for the LGM lowstand may represent the problems of the eustatic concept (Section 2.2.3) or the inherent errors of the method (e.g. environmental range of habitat and dating).

2.3.2.3 Biological indicators - Macrofossils

One means of obtaining a record of sea level change involves looking at the remains of fossilised marine organisms. Marine organisms and biotic communities tend to be arranged in horizontal bands along the slope of the continental shelf that correspond to certain water depths (*biological zonation*). Therefore, if the habitat preferences of a group of fossils are known, and they can be dated; their distribution provides an indication of sea level at that point in time.

For instance, on rocky shores the littoral fringe, or supra-littoral zone (an area that is never submerged but is wetted by surf) tends to be dominated by lichen and Cyanobacteria; the mid-littoral zone (submerged by tides and waves) is home to algae and fauna such as barnacles, limpets and mussels, while the infra-littoral, or sub-littoral, zone (mean sea level to 25–50m) is densely populated by brown algae, and various species of coral, sponges and molluscs (Laborel & Laborel-Deguen, 1995; Pirazzoli, 1996). In addition to the fossils themselves, certain species may leave behind distinctive traces of their activity. The grazing habits of limpets, for example, create distinct erosional patterns. If these patterns can be identified and dated, they too can indicate past sea level (Laborel & Laborel-Deguen, 1995; Pirazzoli, 1996).

Recent studies in the ancient harbour of Marseilles have utilized the presence of dated fossils of the barnacle *Balanus amphitrite* to produce a sea level record for the past 4000 years with a precision of +/- 10cm. In terms of biological zonation, the distribution of this particular species stops abruptly at mean sea level, thus making it a particularly accurate sea level indicator. A similar sea level record had already been obtained from La Ciotat (35 km to the east), this time using a species of calcareous algae - *Lithophyllum lichenoides* (Laborel & Laborel-Deguen, 1995; Morhange et al, 2000).

It must be pointed out that not all biological indicators, have habitat ranges as narrow as *B. amphitrite*. The mussel species *Mytilus edulis* for example has been recorded at depths ranges of less than 5 m down to 20m in the North Sea (Kearney, 2001). Furthermore, significant regional differences may occur in the distribution of individuals of the same species. The polar species *Portlandia arctica*, for example, is found off east Greenland at depths of between 10 and 60 metres. However, it is restricted to a depth of 6m off northeast Greenland. The differences result from the environmental contexts of the two regions. In the latter, the dominance of the polar current in shallow water permits the survival of the species at a lesser depth (Plag et al, 1996). This highlights the need to consider the local environmental conditions when using fauna as sea level indicators.

Plant macrofossils such as tree stumps or vegetation, can provide evidence of sea level fluctuations when found in submerged contexts (Kearney, 2001). These however, tend only to provide broad qualitative indications, i.e – that sea level has risen, or limiting values of sea level change rather than quantitative estimates accurate to within less than several metres. For example, in the case of submerged forests, sea level must have been lower than the tree roots at their time of growth; however, the determination of numerical extent of this change requires analysis of other sea level indicators. This is necessitated by the fact that many coastal or riparian species are not exclusively restricted to the shoreline, but may also occur inland and in upland zones. An example of this is *Pinus taeda*, which, though common in shoreline areas along the US Atlantic coast also occurs inland in the piedmont zone. Exceptions to this consist of species which survive almost exclusively in the intertidal zone, such as mangrove trees, or salt marsh environments such as the *Spartina* species of grass (Long et al, 1999; Pirazzoli, 1996; Kearney, 2001).This is especially true of the latter environment, in that vegetation zonation is such that divisions within a salt marsh can be recognised. For example, in most New England salt marshes the low marsh (up to Mean High Water (MHW)) is dominated by *Spartina alterniflora* (tall) while the high marsh (up to highest spring tide level) is characterised by *Spartina patens*, *Distichlis spicata* and *S. alterniflora* (stunted). Should sufficient freshwater enter the marsh, the high marsh plants will be replaced down to the level of the MHW spring level, by species such as *Phragmites australis* and *Scirpus robustus*. This final zone is termed the upper marsh (Van de Plassche et al, 1998).

The remains of both faunal and floral sea level indicators can also be moved about after they die, and thus the elevation at which they are found may not be an accurate representation of their position in the past. Sea level reconstructions that use biological indicators must therefore ensure that they come from in situ and not derived contexts.

2.3.2.4 Biological indicators – Microfossils

Investigations of sea level using foraminiferal data can be conducted on two temporal scales: one utilises microfossils to look at long term geological scale changes, the other to provide unparalleled high resolution data over the Holocene.

The classic method of using micro-fossils is the extraction of oxygen-isotope records from benthic foraminifera (Shackleton, 1987). As described in Section 2.2.1 the ratio of ^{18}O to ^{16}O in the oceans acts as a proxy indicator of past glacio-eustatic sea level change (Chappell & Shackleton, 1986). This ratio is recorded in the calcareous skeletons of benthic foraminifera and hence can be extracted from the fossil record. However, temperature variations in the abyssal ocean also affect the oxygen isotope ratio of benthic foraminifera. Sea level curves derived from this data are therefore a combination of global (continental ice volume) and local (ocean temperature) components (Figure 25). Therefore a measurement of the oxygen isotope ratio from deep sea foraminifera therefore provides only a first approximation of continental ice volume and hence glacio-eustatic sea level (Shackleton, 1987).

A more accurate second approximation can be obtained by combining planktonic foraminiferal records with their benthic counterparts, as the temperature effect in the former is judged to be minimal (Shackleton, 1987; Chappell et al, 1996). Even so, they are still regarded to have realistic uncertainties of up to ± 20 m (Rohling et al., 1998). Oxygen isotope records go back as far as 5 million years ago (Lambeck et al, 2002a), however, though they only provide a continuous sea level record for the past 140,000 years (Rohling et al, 1998) and in the absence of detailed relative sea level records, have often represented the only recourse when looking at past sea level change over long timescales.

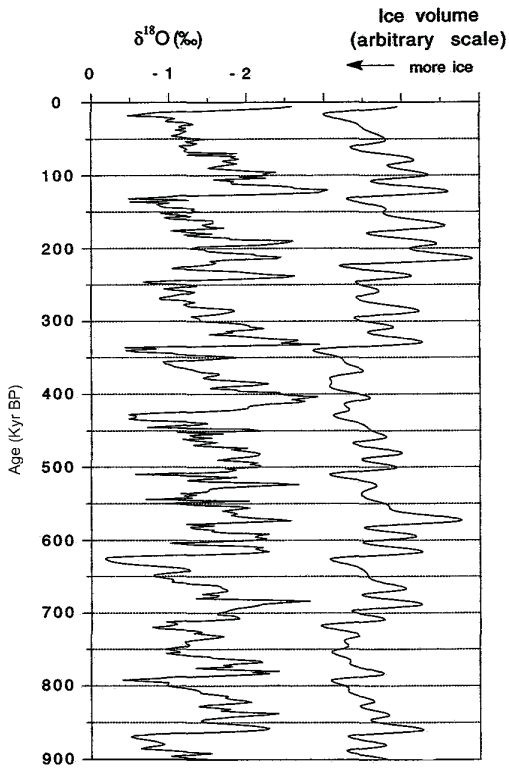


Figure 25. Oxygen isotope record from ocean core MD900063 compared to modelled global ice volumes over the Pleistocene (modified from Bassinot et al, 1994).

A method has recently been developed that allows estimations of past glacio-eustatic sea level lowstands up till 500,000 years ago to be made (Figure 26: Rohling et al, 1998). It involves examining evidence of salinity conditions in the Red Sea during glacial phases in conjunction with a model of water flow between the Red Sea and the open ocean. The basic premise is that at times of lower ocean volumes resulting from the growth of ice sheets, water flow between the Red Sea and the ocean will be reduced, thus leading to increasingly saline conditions which in turn affect the composition of the local benthic and planktonic foraminiferal communities (Sirocko, 2003).

When corrected for local tectonic uplift this technique has yielded depths for the major lowstand events of the last 500,000 years: OIS 6: 131 +/- 6m bpsl; OIS 8: 120 +/-8m bpsl; OIS 10: 122-134 +/-9m bpsl; OIS 12: 139 +/-11m bpsl (all measurements are in metres below present sea level - bpsl).

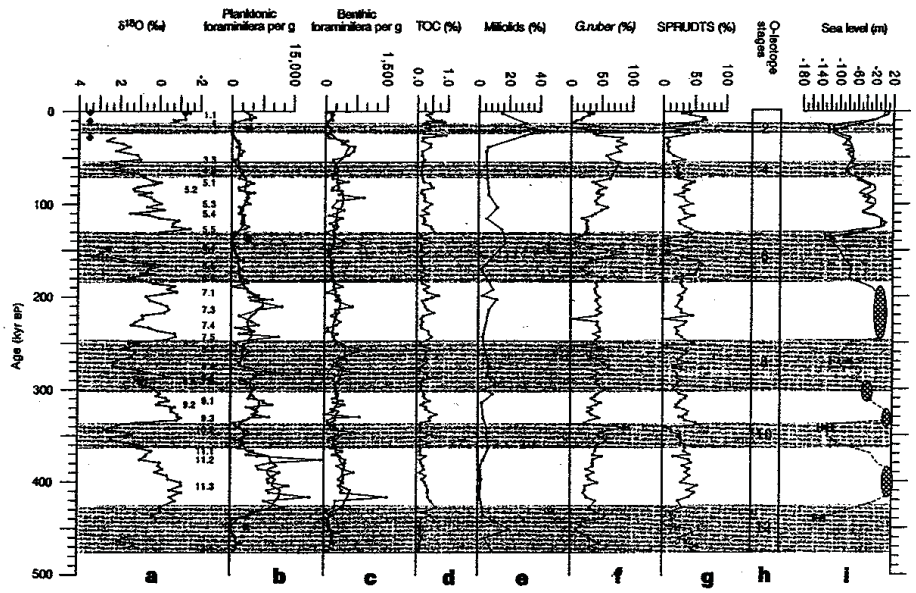


Figure 26. The results of Rohling et al's (1998) inference of sea level lowstands for the past half million years. Column A represents OIS data, columns B to G represent foraminiferal information, column H denotes the OIS stage and column I represents the inferred sea level. The shaded areas in column I represent the error margins involved in the estimation of past highstands. Highstands have been deduced from coral terrace and OIS data, while lowstands have been inferred from Red Sea salinity data.

More recently, this method has been extended to allow a continuous approximation of global eustatic sea level from 470 Ka to present to be made (Figure 27: Siddall et al, 2003). This reconstruction is claimed to be accurate to within +/- 12m and to provide centennial scale resolution for the period between 70 and 25Ka, the highest so far for this period (Figure 28) (Siddall et al, 2003).

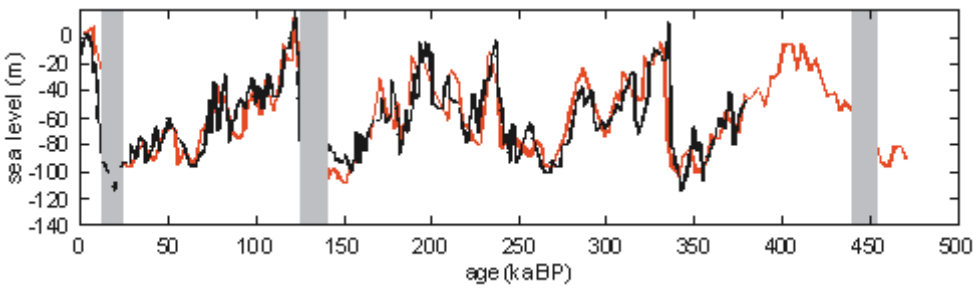


Figure 27 Sea level change over the past 470 Ka obtained from salinity conditions in the Red Sea. The grey bands represent gaps in the record resulting from aplanktonic conditions. The red and black lines represent data from two different Red Sea cores (from Siddall et al, 2003).

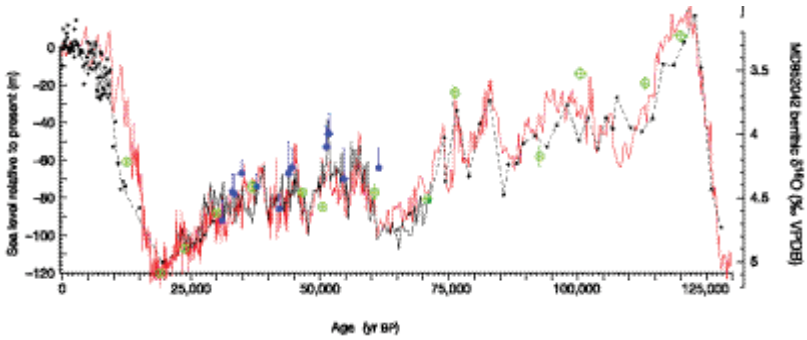


Figure 28 High-resolution sea level record from 125 Kyr BP till the present as obtained from Red Sea data. Red line = benthic oxygen isotope record, black line = sea level record obtained from core KL11, broken black line = centennial scale resolution record from core KL11, blue and green circles = dated coral records (modified from Siddall et al, 2003).

High resolution temporal studies can also be undertaken and again this method utilises the habitat preferences of certain species of diatoms and foraminifera. The vertical distributions of foraminifera in particular, are closely related to tide levels and thus have the potential to provide high resolution and detailed records of relative sea level change (Van de Plassche et al, 1998). This can be done using foraminiferal transfer functions, or statistical methods that extract the ecological data from modern distributions of salt marsh foraminifera, and which can then be applied to fossil foraminiferal distributions to obtain palaeo-ecological data. (Edwards & Horton, 2000). For example, low marsh and mudflat environments in the UK tend to be dominated by *Ammonia beccarii*, *Haynesina germanica* or the *Elphidium* species. Subtidal environments though contain a high proportion of the *Reophax* species (Edwards & Horton, 2000).

Analysis of AMS dated foraminiferal assemblages from sediment cores taken from Poole Harbour have identified four phases of sea level change in this area; rising relative sea level from 4700 till 2400 cal BP, stable to falling from 2400 till 1200 cal BP, a brief rise between 1200 and 900 cal BP and then stable to falling until a renewed rise between 400 and 200 cal BP. This contrasts somewhat with the earlier record of sea level change in this area, which could only indicate a constant rise over the past 5500 years (Edwards & Horton, 2000; Edwards, 2001).

Microfloral evidence, in the form of pollen, can also be used as a sea level indicator, in that it can be used to reconstruct past coastal environments on the same principle as plant macrofossils (see section 2.3.2.3). If the elevations and dates of these pollen remains are known, then their relationship to past sea level can be inferred. Pollen distributions however are influenced by a number of taphonomic factors such as the nature of local air and water transport pathways, the size and strength of the pollen grain and the preservational conditions of the deposition site. In addition, pollen deposited in the intertidal zone can also suffer extensive reworking as a result of tidal forces (Long et al, 1999).

2.3.2.5 Sedimentary and stratigraphic indicators

Analysis of sedimentary sequences can also provide clues as to sea level change in the past. This arises from the fact that the types of sediment deposited under marine conditions are different to those laid down under terrestrial conditions. For example, muds and clays are usually deposited in calm water or low energy environments such as sheltered coastal basins. Changes between layers can be determined by biostratigraphical, granulometric and physiochemical analysis. In all cases the occurrence of marine sequences above terrestrial ones implies a transgressive event, while a terrestrial layer overlying a marine one implies a regressive event (Pirazzoli, 1996). Examples of commonly used sedimentary indicators include beachrock – beach sediments deposited in the intertidal zone and cemented by calcium carbonate. These deposits are generally regarded as providing good estimates of past mean sea level. Their accuracy varies within a vertical limit depending on the local tidal range. Near tideless areas, such as the Mediterranean, are characterised by thin layers of beachrock, while in macrotidal areas they can be greater than 3 metres in thickness (Pirazzoli, 1996).

Dated transitions between peats formed under freshwater and brackish conditions also allow sea level inferences to be made. Freshwater peats only provide an upper limit for past sea levels as they can form at any altitude given the presence of stagnant water. Brackish peats however, can provide a better estimation of the past shoreline, as they are generally located near mean sea level, though their altitude can vary as a result of local topography and tidal range (Pirazzoli, 1996). Stratigraphic analysis of sediment from the Pacific North West coast of North America has revealed sequences of marsh peats, with well preserved vegetation, overlain by intertidal mud. This has been interpreted as rapid coseismic subsidence and transgression of the land (Long & Shennan, 1998). In all cases, the appearance of freshwater peats above brackish ones deposits indicates regression, while brackish deposits above freshwater ones are a sign of transgression (Lambeck & Chappell, 2001). The accuracy of peat deposits however is relative and in many instances the changes cannot be quantified (Van der Molen, 1997). More accurate sea level estimates though can be made on the basis of combining sedimentological information with the biological they contain (see Sections 2.3.2.3 and 2.3.2.4). Notable examples of these sorts of studies include biostratigraphical investigations of foraminifera and vegetation from salt marsh sediments on the east coast of the USA and the south coast of Britain (e.g. Edwards, 2001; Van de Plassche et al, 1998) that have provided a number of high resolution Holocene sea level curves.

However, sediment compaction may result in errors when sequence stratigraphy is used as a means of providing a measure of sea level change. As described in section 2.3.4 the weight of overburden may lead to fluid loss induced compaction. In

sediments with a high water content this can lead to a significant decrease in the thickness of the layer, up to 90% in the case of some peat layers. As a result the apparent elevations of the sea level indicators may be rather less than they really were at the time of deposition (Pirazzoli, 1996).

Over very long timescales (i.e. hundreds of thousands to millions of years), stratigraphy on continental shelves can provide an image of the pattern of sea level change over time. The basic premise is that each large-scale sea level fluctuation removes sediment from the continental margins and re-deposits it in recognisable patterns that can then be used to interpret palaeo-sea level trends.

For example, during lowstands large portions of the continental shelf are exposed to the atmosphere and channel cutting by rivers draining the land right down to the lowstand shoreline occurs. As marine transgression takes place, these terrestrial deposits will be overlain by a marine sequence. If the cycle continues, these in turn will be overlain by a successive layer of terrestrial lowstand deposits and so on. This technique has provided relatively coarse sea level curves going back hundreds of millions of years (see Figure 8A). On a shorter timescale though, such as the period since the LGM, sequence stratigraphy data can provide preliminary information as to the location of past shorelines, and thus a qualitative rather that quantitative estimate of past sea level that can then be narrowed down through use of other sea level indicators (Leeder, 1999; Pirazolli, 1996). The types of environments typically encountered on the continental shelf are described in more detail in Theme 3 (Section 4).

Caution must always be taken with stratigraphic interpretations and in particular care should be taken over the dating and general temporal interpretation of sequences. This can be best illustrated through an example of extreme event, storm surges or tsunamis, sedimentation rather than that associated with stable sea levels. This is the case in western and northern Scotland, where a layer of marine sand deposited around 7100 to 7200 yr (C^{14}) BP has been interpreted as the result of a tsunami initiated by the Storegga slide, an underwater landslide that took place off western Norway. This layer is currently located between –1.1 and 8.9m above current Mean High Water Spring tide depending on the extent of local uplift (Dawson & Smith, 2000; Smith et al, 2000). In western Norway, the effects of this tsunami and the main Holocene transgression can be distinguished on the basis that the former deposited beds of marine sand, rip up clasts of peat, marine silt and gyttja, and coarse plant material, while the latter laid down sequences of homogenous gyttja with some silt and fine sand (Bondevik et al, 1998).

2.3.2.6 Archaeological Indicators

Archaeological evidence can also be used to provide indications of sea level change. Submerged terrestrial artefacts or structures represent convincing evidence of sea level rise. Examples of these are diverse and range from Roman harbour constructions to Neolithic monuments to Palaeolithic implements (e.g. Flemming, 1998; Scarre, 1984; Smith & Bonsall, 1991 respectively). These are somewhat limited in that this sort of data can only provide a qualitative estimate of the sea level trend (i.e rising or falling sea level).

More accurate archaeological indicators of sea level change consist of dated examples of artefacts or structures designed specifically for use on the foreshore or the intertidal zone. Slipways, harbour constructions, fish traps, fish tanks or salt

extraction sites can be, and have been, used to provide estimates as to the position of the past shorelines (Blackman, 1973; Pirazzoli, 1996; Scarre, 1994; Flemming, 1998).

For example, a series of submerged slipways in the port of Apollonia has been used to estimate a local relative sea level rise of 2m. This measurement has been inferred on the basis that any less depth at the base of the slips would have restricted access to all but very small ships, while any greater depth would have reduced the dry length of the slips to too great an extent (Blackman, 1973).

In terms of accuracy, the estimates of past mean sea level as inferred from these indicators tend to be of the order of several metres. This is primarily a consequence of the fact that their relationship to the past shoreline is rarely exactly known, but can be estimated on the basis of the function of the particular indicator. However, it must be remembered though that many of these sites need not necessarily lie on the shoreline proper, but could be situated in estuaries or inlets, or could even be linked to the sea by canals (Scarre, 1984). Furthermore, it should also be kept in mind that harbour structures may not be complete. For example, the present day submergence of a quay or breakwater could be as much due to the loss of the upper parts of the structure rather than a local change in sea level (Blackman, 1973). Exceptions, to these do occur and in particular in areas with very low "micro-" tidal regimes such as the Mediterranean. Here tidal erosion notches (see Section 2.3.2.1) can be cut in specific maritime features (e.g. quay and harbour walls) providing a sea level at an inferred date (Blackman, 1973). However, this further highlights one of the key issues of sea level indicators i.e. the ability to date them.

2.3.3 Dating sea level indicators

The construction of a record of sea level change over time, whichever of the indicators described in Section 2.3.2 are used, requires accurate dating. There are a number of available methods that can be brought to bear, depending on the nature of the indicator in question. Frequently used methods include radiocarbon dating; both conventional ^{14}C and the newer AMS (Accelerator Mass Spectrometry) method; Uranium series dating, tephrachronology and dendrochronology.

However, dating the sea level indicators may also present a number of interpretative difficulties as each method has its own inherent errors margins. Radiocarbon dates for example deviate from calendar years as a result of temporal variations in atmospheric carbon. This deviation increases over time by the last glacial maximum can be as much as 3000 years (Bard et al, 1990). In addition, radiocarbon dates do not provide an exact figure, but an estimate to within 2 standard deviations. Similarly, although Uranium series dating does not need calibration in the same way as radiocarbon dates, uncertainties do creep in the further back in time one goes (Chappell & Shackleton, 1986; Pirazzoli, 1996; Kearney, 2001).

It is beyond the scope of this review to demonstrate multiple dating techniques, and cover their respective advantages and disadvantages in detail. More detail on the relevant dating methods can be found in a number of texts, such as Aitken (1990) and Taylor & Aitken (1997). What should be taken away from this brief discussion is that there are errors inherent in most dating methods and these will have an impact on the resulting sea level record, as will be demonstrated in the following section.

2.3.4 Displaying past sea level change

The creation of a sea level record, or sea level curve, requires a sequence of index points (i.e. sea level indicators) whose age, location, altitude (relative to a modern datum) and indicative meaning (a quantified vertical relationship within a former tidal frame – i.e. their relation to former sea levels) is known (Edwards & Horton, 2000). These can then be plotted such that the elevation or depth of the indicator is on the vertical axis, and its date on the horizontal axis. However, sections 2.3.2 and 2.3.3 have highlighted the fact that significant variations may exist with respect to the position of the indicators relative to sea level, and the dating techniques. Collectively, these can be termed '*age-height errors*'. Curves should therefore be drawn with this in mind, and error bars or error bands should be used to represent both the height range and the date estimate of each index point (Shennan & Tooley, 1987; Pirazzoli, 1996).

As the creation of a sea level curve is a subjective interpretation on the part of an author of a number of observed index points, simply depicting sea level with a single line represents a poor summary of the situation and represents the interpretation of the observations by the curve's author for their own purposes. This does not provide the reader with the information that will allow them to draw their own conclusions out of the original data, or indeed gain a sense of the uncertainties involved (Pirazzoli, 1991). Figure 29 represents an example of this. Although the data in question is the same for both curves, with the exception that the Bard et al (1990) curve is calibrated, the potential age height errors are hidden in one curve but not the other.

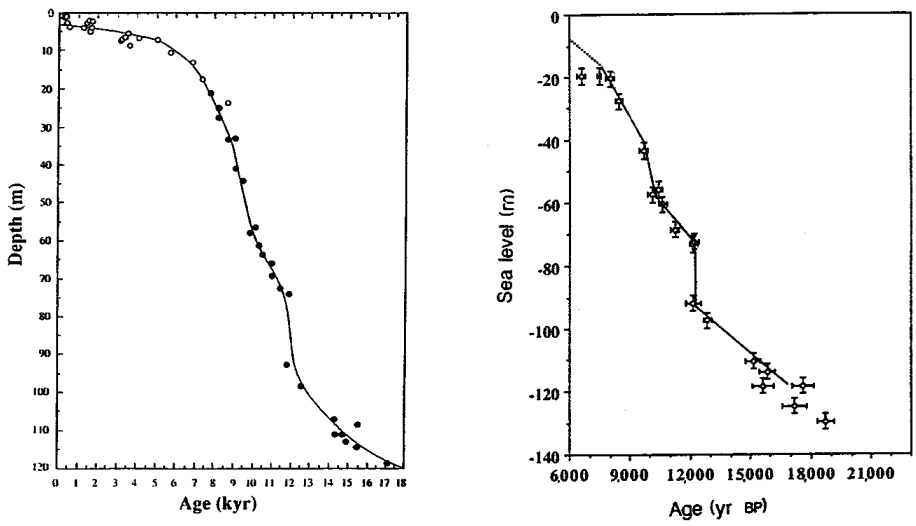


Figure 29. Glacio-eustatic sea level inferred from Barbados coral reefs. In the curve on the left the potential age height errors are hidden while in the curve on the right they are illustrated through the use of error bars (modified from Fairbanks, 1989; modified from Bard et al, 1990).

The age-height errors within an individual sea level curve can have a significant impact on any resulting palaeogeographic reconstructions in that the reconstructed shoreline position can vary from anywhere between several tens of metres to tens of kilometres depending on the gradient of the coastal area under reconstructed, and the magnitude of the errors within the sea level data (see Section 2.5).

2.4 Numerical Models and Predicting Past Sea Level Change

2.4.1 Background

Ideally, all investigations of past sea level change would be based on observations of relative sea level indicators. However, as Section 2.3.1 has indicated, these are not always available, especially with respect to the earlier phases of the Quaternary. Further, as Sections 2.2.3 and 2.3.2.4 illustrate, global glacio-eustatic changes inferred from proxies such as the oxygen isotope record represent only an approximation of past sea level change. A number of Glacial Isostatic Adjustment (GIA) models have therefore been developed to fill in the gaps in the record, by providing predictions of past sea level change and the position of palaeo-shorelines by taking into account both the key eustatic and isostatic variables. The construction of these models therefore requires the following information:

- A model of the glacio-eustatic term.
- A model of the Earth's rheology and its response to changing surface loads, both through isostasy and geoidal fluctuations.
- A detailed description of the growth and decay of the ice sheets.

Not all the processes discussed in Section 2.2.2 to 2.2.5 are included in these models. The reason for this is that the complexity of the sea level change process is such that some degree of simplification is necessary in order to allow its modelling. The scale of the models (e.g. Lambeck, 1993a, b; 1995; Peltier, 1998) also tends to be regional and global rather than local and as a result the factors modelled tend to be those that operate over larger spatial and temporal scales. It has in fact been explicitly stated that the basic models assume that the exchange of surface loads takes place entirely between ice sheets and oceans, that ice and water density are constant and that the effect of other contributions to sea level change, such as thermal expansion and tectonics are ignored (Lambeck et al, 2003).

The two main research groups involved in this field are based at the Australian National University (ANU) and the University of Toronto, with the former tackling mainly regional situations (e.g. the British Isles) and the latter approaching the problem from a more global perspective. Key figures in these groups include Kurt Lambeck (ANU) and Richard Peltier (Toronto). The basis of both of their approaches is the sea level equation formulated by Farrell and Clark (1976), the solution of which makes possible the computing of sea level given a known ocean and ice load (Mitrovica, 2003). However, differences do exist in the ways this solution is accomplished, for instance over the use of different algorithms in analysing the migration of shorelines over time (Mitrovica, 2003). Several of the differences between the two groups have been highlighted in a recent debate sparked off by Peltier's (2002b) comment that the ANU group's approach was based on 'faulty logic' and 'invalid notions', a claim which has inevitably been refuted by the ANU and independent authors (Lambeck et al, 2002c and Mitrovica, 2003 respectively).

In terms of comparing the work of the two research groups, analysis of the theories and methods of the ANU group by independent researchers has lead to the claim that their approach is significantly more accurate than that used by Peltier (e.g. 1994; 1998) and his co-workers (Lambeck et al, 2003). At the time of writing however, a reply from the Toronto group has not been published.

2.4.2 Basic Principle

The basic principle governing the GIA models is as follows. As relative sea level change is an outcome of the interplay between the eustatic and isostatic processes described in Sections 2.2.3 and 2.2.4, the observed sea level record has the potential to provide information on each of them. This stems from the feedback process implicit in Section 2.2.3.2 – sea level change is dependent on the Earth's responses to variations in surface load, which in turn are partly a function of changes in ocean volume (Clark, 1980; Lambeck et al, 2003; Mitrovica, 2003). More specifically the observed sea level record provides information which constrains the ice and earth models used in the interpolation process (Johnston & Lambeck, 2000). The fact that sea level records are available from a number of different regions and thus may reflect the dominance of different isostatic or eustatic influences (see Figure 9) means that it is possible to separate out the above parameters to some extent and examine their effects separately (Nakada & Lambeck, 1988). Numerical modelling of ice sheets, water distribution and isostatic rebound then provides a picture of changing land surfaces over time onto which the glacio-eustatic function can be applied.

Each of the above parameters will now be examined in greater detail.

2.4.3 The Glacio-eustatic Term

Glacio-eustasy is the only ocean volume altering input included in these models, for the reason that other inputs tend to add relatively little to volumetric changes (see section 2.2.3.1). It is worth pointing out, that for this reason much of the literature concerning numerical modelling tends to use the term 'eustatic' synonymously with 'glacio-eustasy' (see Milne et al's (2002) comments in section 2.2.3.1).

Published estimates of the ice volume equivalent sea level change since the last glacial maximum range from as high as 163 metres to as low as 102 metres, though the most common adopted measurements fall between 116 and 140 metres (Chappell & Shackleton, 1986; Fairbanks, 1989; Clark & Mix, 2002; Lambeck et al, 2002b). In particular, the most recent published estimates of the ice volume equivalent sea level change since the LGM are 130 to 135m (Yokoyama et al, 2000; Lambeck et al 2002c, Mitrovica, 2003), although Siddall et al (in prep) will again push this value back down to 125 m.

This variation is the result of many different studies operating in different areas where varying isostatic, tectonic and geoidal contributions or differing interpretations of the sea level indicators may have skewed the samples (Lambeck, 1996; Pirazzoli, 1996). Even regions far from the centres of glaciation will be affected to some extent by isostatic and geoidal fluctuations. The best estimates of glacio-eustatic change are therefore based on regions where local tectonic and isostatic movements are believed to be minimal or insignificant, or where corrections can be made for these movements (Lambeck, 1996; Lambeck & Chappell, 2001). Commonly used examples include records from Barbados (Fairbanks, 1989), the Huon Peninsula (Chappell & Shackleton, 1986), the Bonaparte Gulf (Yokoyama et al, 2000) and the oxygen isotope (OI) record (Shackleton, 1987; van Andel & Tzedakis, 1996, e.g. Figure 30).

While these eustatic sea level measurements are not truly globally applicable (see section 2.2.3.1), they are critical for constraining the glacio-eustatic volumes input into the modelling process (Scourse & Austin, 2002). Care should also be taken with the OI record as evidence from different cores may be affected by local variations in temperature and salinity, which will in turn affect the ratio of O^{18} to O^{16}.

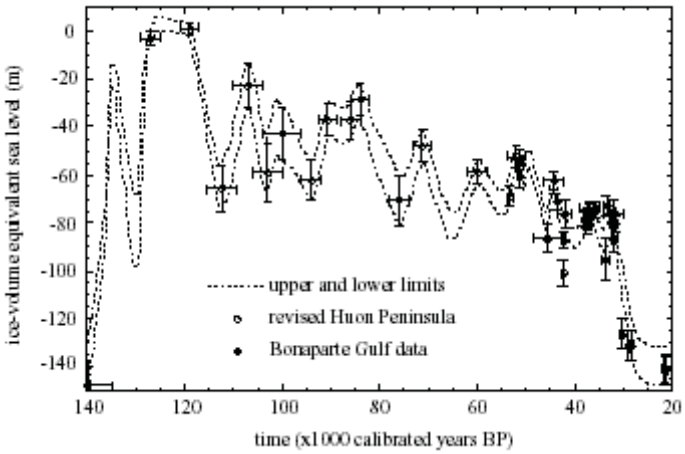


Figure 30 Ice volume equivalent sea level change for the last 140,000 years obtained from raised coral reefs on the Huon Peninsula and sediments from the Bonaparte Gulf. Note the variation between upper and lower limits of the sea level observations (from Lambeck et al, 2002b)

2.4.4 Earth rheology

As explained in section 2.4.1, models of the Earth's response to surface loading revolve around the idea of an elastically deforming lithosphere floating on top of a viscous mantle. The main questions in relation to interpolation models focus on the depth dependency of the mantle viscosity and whether this viscosity is laterally uniform. Seismic studies and analysis of glacial rebound from observed sea level records have tended to indicate that viscosity is indeed depth dependent, with an average lower mantle (below 670km) viscosity that is between 50 and 100 times greater than that of the upper mantle. The significance of this is that ice sheets tend to only stress the upper mantle, however, meltwater loading, given the size of the oceans, is much greater in extent and thus stresses the lower mantle as well. Both layers therefore need to be taken into account. Simple models of glacial rebound therefore tend to use a three-layer model with a lithosphere of between 60-100 km thick, an upper mantle of viscosity of between 2 to 5 x 10^{20} Pa s, and a lower mantle of viscosity 2 to 50 x 10^{21} Pa s (Johnston, 1995; Lambeck, 1996; Peltier, 1998; Lambeck & Chappell, 2001; Milne et al, 2002). In some instances, models with up to five layers (lithosphere, upper mantle 1, upper mantle 2, transition zone, lower mantle) have been used (e.g. Lambeck, 1995). However, in most cases the 3 layer model appears to be adequate. The choice of mantle parameters is obtained by iterating, or searching for solutions whereby predicted and observed sea level changes qualitatively agree with each other. This choice of parameters is important as different lithospheric thickness and mantle viscosities will respond differently to changing loads, thus leading to different predictions of isostatic rebound. Individual modellers do however use different parameters in their models, and as a consequence, differences can appear in their predictions. For example, recent modelling of the isostatic adjustment of the British Isles has made use of a modified version of Peltier's ICE 4G (VM2) model with a lithosphere of 90km and mantle parameters of 4 x 10^{20} Pa s (upper mantle) and 2 x 10^{21} Pa s (lower mantle) (Peltier et al, 2002; Shennan et al, 2002). This contrasts somewhat with the parameters adopted by Lambeck as set

out in Table 3 below, which are similar for the upper mantle, but include a thinner lithosphere and more viscous lower mantle.

Recent studies have indicated that lateral variations in mantle viscosity and lithospheric thickness do exist (see Table 3). These general variations can be translated to regions so that, for instance, mantle viscosity beneath continental margin Australia will be less than that of northern Europe (Lambeck et al, 2002b).

Model	H_1 (km)	Upper Mantle η_{um} (x 10^{20} Pa s)	Lower Mantle η_{lm} (x 10^{21} Pa s)
Continental mantle	65-85	3-5	5 - 30
Continental margin mantle	65-80	1.5-2.5	5 - 30
Oceanic mantle	~50	~1	(10)
Values for specific regions as estimated from sea level observations			
Australia	70-80	2-3	5 - 30
Australia	75-90	1.5-2.5	(10)
Scandinavia	65-85	3-4	6 - 13
British Isles	65-70	4-5	7 - 13
Northwest Europe	(65)	2-5	10 - 30
South Pacific	~50	1	(10)

Table 3. Lateral variations in lithospheric thickness (H_1), upper mantle viscosity (η_{um}) and lower mantle viscosity (η_{lm}). Estimates have been obtained by different authors at different times, hence the multiple Australian results (after Lambeck & Chappell, 2001; Lambeck et al, 2002b).

Until very recently though, no models took account of this spatial variation and as a result, rheological parameters were back calculated from the available sea level data in each region under study (Lambeck, 1996; Lambeck & Chappell, 2001). The latest approaches however, model isostatic predictions for a range of Earth models that have been obtained from regions where mantle parameters have been estimated from observed sea level change (Lambeck et al, 2002b).

2.4.5 Surface Load (ice sheets and oceans) Models

Tied into these rheological models is information concerning the size and distributions of the Earth's ice sheets and oceans as these determine the extent and position of the isostatic movements. Analysis of glaciological and geological features such as moraines, drumlins and erratics allow estimates of ice sheet limits and the pattern of retreat since the last glacial maximum to be made (Lambeck, 1993a; 1993b; Lambeck & Chappell, 2001). Only the pattern of retreat is known since the retreating ice will have destroyed most evidence of the ice advance. Analysis of oxygen isotope samples does however provide some indication as to the broad trend of advance as well as retreat. The extent of the ice sheets at the LGM is relatively well understood (e.g. Bowen et al, 2002), with the major contributions to sea level change coming from the American ice sheets - the Laurentide, Cordeilleran and Innuitian, the European ones – the British and Scandinavian, and the Antarctic. Evidence also points to the existence of an ice sheet over the Barents Sea and possibly the Kara Sea (Clark & Mix, 2002).

Uncertainties though, arise over the thickness of the ice sheets. In some areas, it can be measured off nunataks, however, where these are not present estimates must be made based on snow supply, ablation and the nature of the rock-ice interface (Nakada & Lambeck, 1988; Lambeck & Chappell, 2001). In some instances, a constant scale factor may have to be applied to the estimated ice sheet thickness and iterated until a solution is found that is in good agreement with the observed evidence of glacial limits (Lambeck 1993a). Alternatively, assuming that mantle parameters and Earth rheology are known, it is possible to numerically calculate ice sheet extent and size from observed sea level data (Johnston & Lambeck, 2000).

Ice sheets are not solely restricted to land; floating and grounded marine based sheets also exist. Contributions from floating ice to the hydro-isostatic term do not need to be included, as they have already contributed their mass to the volume of ocean water. Grounded marine ice sheets however, will exert pressure directly on the lithosphere and upper mantle, and should be treated in the same way as land based ones. However, ice sheets do experience stages of floating and grounding, and thus it is necessary to calculate the amount of floating ice at each epoch (Lambeck et al, 2003).

In all cases, the effects of ice sheets far from the study area, as well as those close to and within it must be taken into account. While near field ice sheets are likely to contribute to sea level change through glacio-isostasy, ice sheets further away will contribute to the glacio-eustatic change in sea level, and hence, hydro-isostatically as well. In the case of the North Sea basin, the general scheme is uplift and falling sea levels in the northern sector (Scotland, Sweden, Norway and North Denmark) with subsidence and rising sea levels in the southern sector (England, Belgium, the Netherlands, South Denmark) (Figure 18). Large areas of the North Sea floor are

likely to be subsiding as well due to the collapse of the glacial forebulge (Lambeck, 1993a, b, 1995; Lambeck et al, 1998; Kiden et al, 2002; Shennan & Horton, 2002).

Furthermore, rebound models must take into account the effects of glacial loading prior to the onset of deglaciation. With respect to the Late Pleistocene this means knowledge of the size and distribution of the ice sheets before the LGM. The starting point for the ice sheet history is therefore usually taken to be the Last Interglacial Stage (OIS 5e – c. 128 to 118 Kyr BP), and it is assumed that at this point in time the planet was at isostatic equilibrium (Lambeck et al, 2003). Given the lack of glaciological evidence for this, oxygen isotope data (e.g. Chappell & Shackleton, 1986; Shackleton, 1987) must be used to provide an approximation of the situation. This is necessary because the maximum extent of the ice sheets may not have persisted long enough for the mantle to reach a state of hydrostatic equilibrium. Consequently models which assume isostatic rebound beginning from a hydrostatically stable position may lead to overestimations of uplift and thus inaccurate sea level predictions (Lambeck, 1993a).

With respect to the distribution of the meltwater load over time, care must be taken to ensure that the migration of shorelines due to relative sea level rise is included, as this will affect the extent of the hydro-isostatic contribution through the distribution of meltwater. Fixed coastline models will tend to underestimate this contribution since a point close to the coastline will remain so throughout the deglaciation period. In reality, the movement of shorelines could result in it being ten or hundreds of kilometres from shore for much of the deglaciation and therefore subject to a greater water load and therefore a larger hydro-isostatic effect (see section 2.2.4.2: Johnston, 1995). Shoreline changes due to the position of grounded marine ice also need to be taken into account. Ice, being more massive, will displace water, hence there will be areas that might otherwise be inundated (i.e they lie below local relative sea level), but are covered by ice and thus should not be included in the water load term (Lambeck et al, 2003).

Recent studies also take into account the effects of what Milne has described as 'water dumping' (Mitrovica, 2003) and Peltier as 'implicit ice' (Peltier, 1998, 2002a). Essentially, the disappearance of a marine ice sheet leaves behind a 'hole' which serves as accommodation space for meltwater. Therefore, additional ice melting has to be included in the models to reconcile this fact with the above constraints on glacio-eustatic change (Mitrovica, 2003).

2.4.6 Solution of the models

The solution of these models takes the form of an iterative procedure whereby a range of earth parameters and ice and meltwater models are modelled in search of an optimum solution (Lambeck, 1993a, b; 1995; Lambeck et al, 2003). As stated earlier, glaciological and observed sea level evidence are used to constrain the parameters of the ice and earth models, and in addition, the observed sea level evidence may be compared to the predicted results. This allows iterative fine tuning of the parameters and thus ensures a better fit, and hence a more accurate model, between the observations and predictions. More detailed accounts of this process can be found in the literature produced by the various modellers (e.g. Lambeck, 1993a,b; Lambeck et al, 2003; Milne et al, 2002; Peltier, 1994; 1998). Figures 31, 32, 33 display different palaeo-shoreline predictions for North West Europe made using a variety of GIA models.

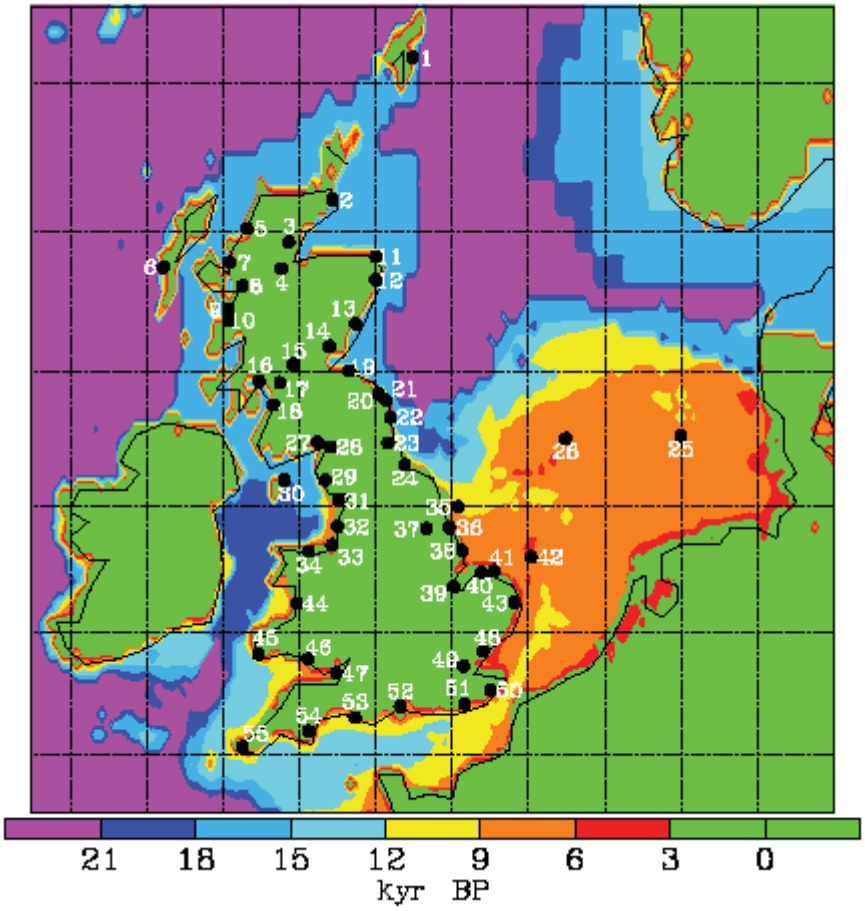


Figure 31. GIA model reconstruction of palaeo-shorelines (a.k.a. submergence history) and ice extents for North West Europe from 22 ka (C^{14}) BP till present (from Peltier et al., 2002).

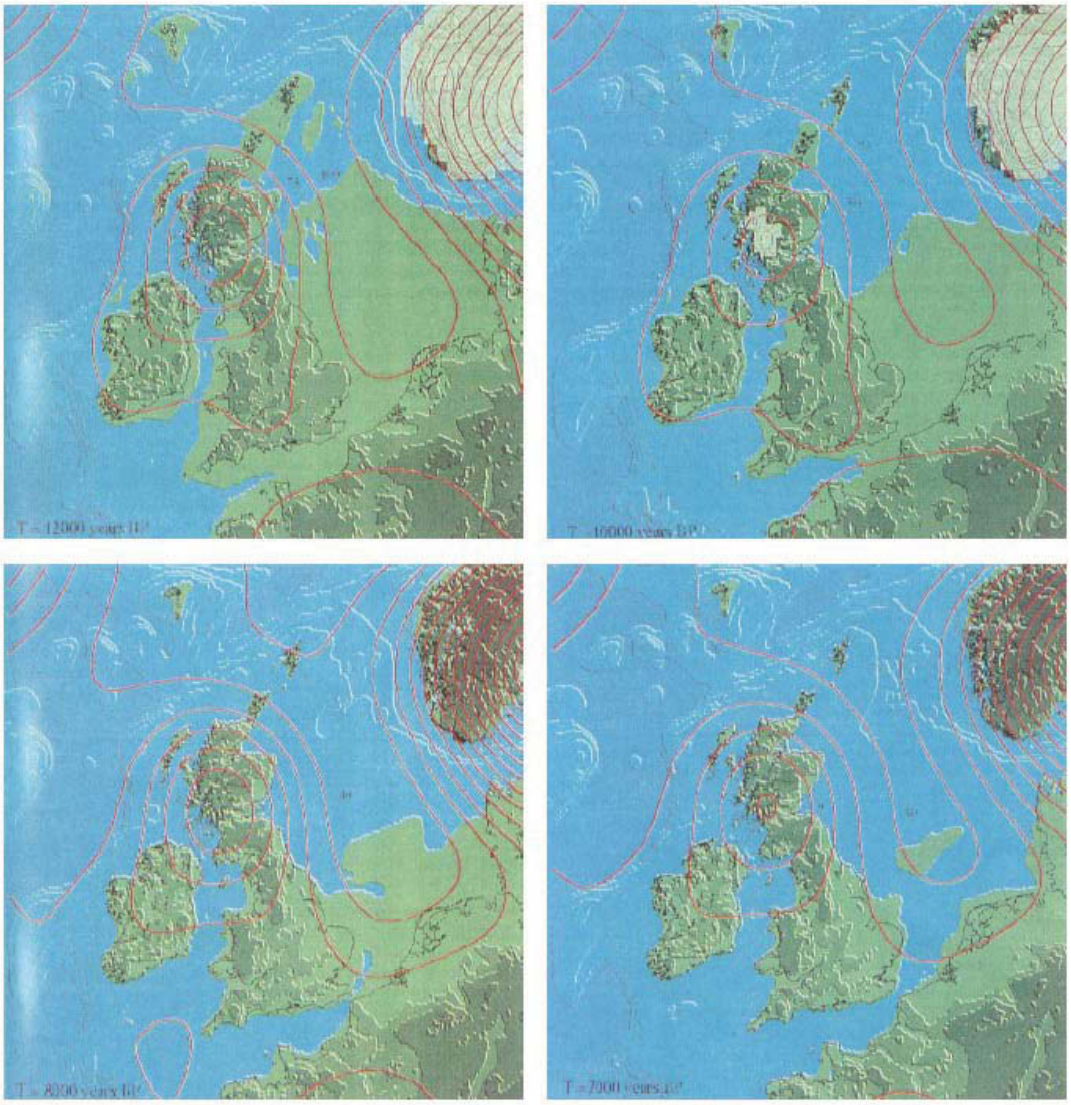


Figure 32. GIA model reconstruction of palaeo-shorelines and ice extents for North West Europe from 12 till 7 ka (C^{14}) BP. Red lines are isobases representing the degree of isostatic uplift and subsidence (from Lambeck, 1995).

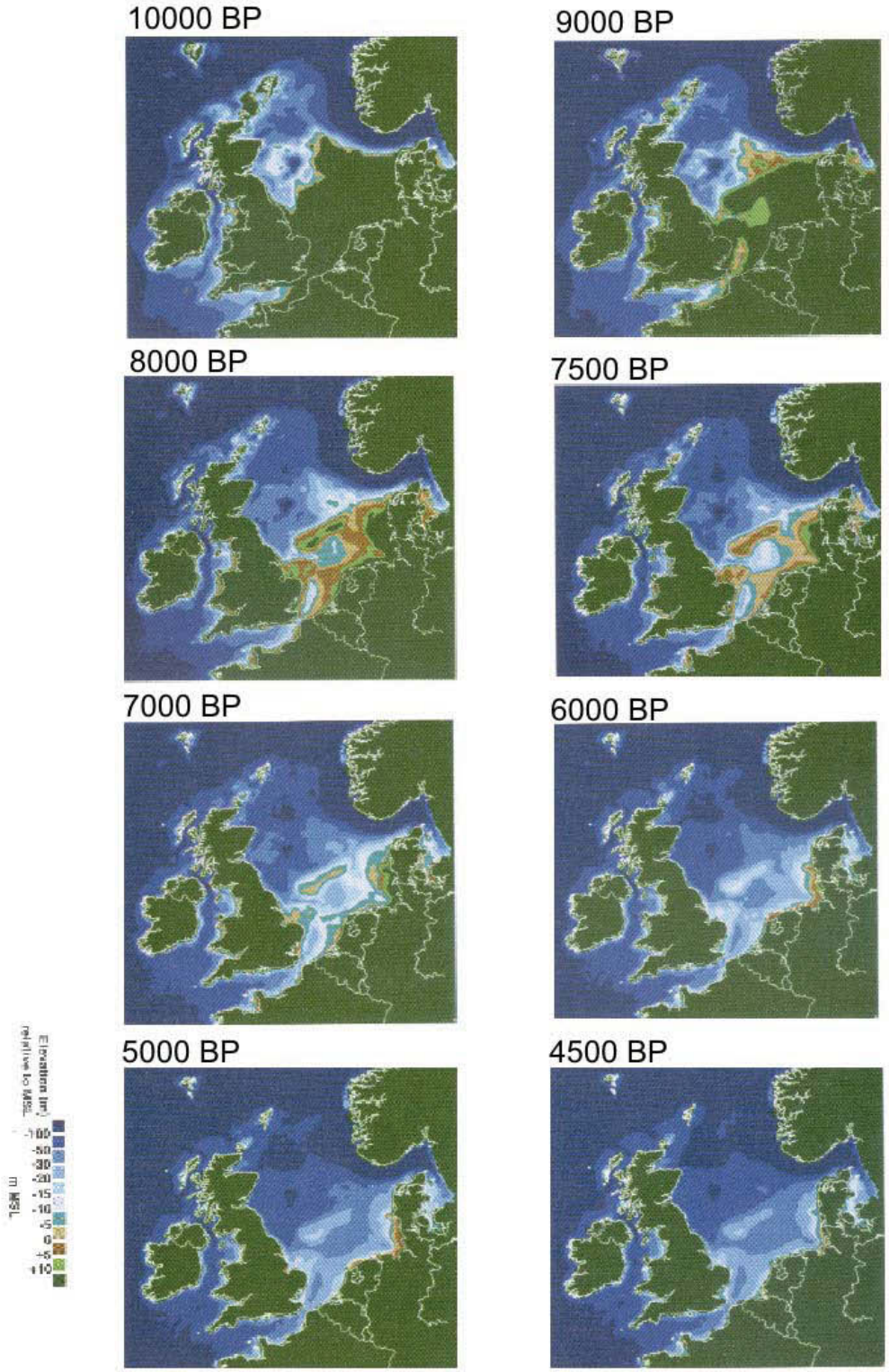


Figure 33. GIA model reconstruction of palaeo-shorelines and ice extents for North West Europe from 10 till 4 Kyr (C^{14}) BP. The colour coded contours represent elevation in metres relative to mean sea level (from Shennen et al, 2000b).

2.4.7 Effectiveness of the models

In plan view the difference between models can be clearly seen if one compares the 12 ka C14 BP model of Peltier et al. (2002 and see Figure 30) with Lambeck (1995 and see Figure 31). The palaeo-shorelines in the former (yellow – turquoise boundary) extend well in to the English Channel, extends as far North as the Firth of Forth in the North Sea and isolates the Shetland Isles. By comparison the Lambeck model has palaeo-shorelines that are located towards the entrance of the English Channel, extand as far north as the Moray Firth in the North Sea and encompass the "Shetland Isles" as part of a single land mass.

In terms of comparison with measured relative sea level curves obtained from UK shores these numerical models do provide good fits (see Figure 34), yet a unique solution has still to be developed that agrees with all the available sea level observations (Shennan & Horton, 2002).

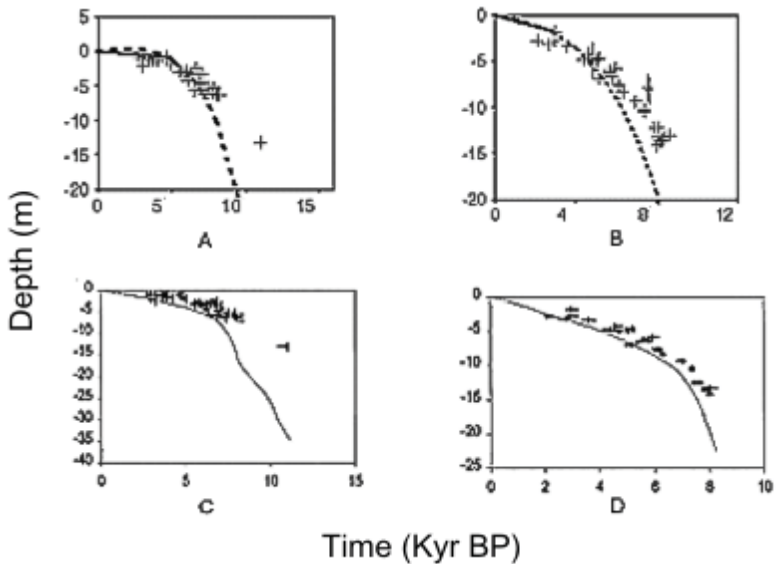


Figure 34 Comparisons of observations (index points) and predictions (lines) of sea level from the Tees estuary (A and C) and Lincolnshire (B and D). Predictions for A and B are based on Lambeck's (1995 model), and predictions for C and D are based on Peltier et al's (2002) modified ICE4G(VM2) model with a lithosphere of 90km (modified from Shennan et al, 2000a; Shennan & Horton, 2002).

McCabe (1997) for instance has pointed out that the numerical predictions for the Western Irish Sea basin do not match the sedimentological and glaciological evidence, while Plag et al (1996) draw attention to the existence of a number of unexpectedly low submerged shorelines, such as the –100m features on the Hebridean Shelf, the central Celtic Sea and the Viking Bank. These all date to c. 11,000 BP and exhibit crustal adjustment that is greater than predicted by the numerical models. In general, on a regional scale the models tend to underestimate sea level and discrepancies between observations tend to be greatest for areas under the thickest ice, or where ice limits or volumes are least well known (Shennan et al, 2000a; Shennan & Horton, 2002).

The problem has partly arisen as a result of the fact that modellers tend to pick certain factors to go into the models while ignoring others. As mentioned in section 2.4.1, only long term and large-scale influences on sea level tend to be examined.

The results of this are especially clear in that the models tend not to be able to predict short duration (c. 1000 to 2000 years), high amplitude events, such as rapidly migrating marginal bulges close to the former ice sheets (Plag et al, 1996). The situation is rendered even more problematic by the fact that certain parameters such as ice sheet thickness, lithospheric thickness and mantle viscosities are still not yet exactly known, as described in Section 2.4.4 and Table 3.

In defence of the models, they are still works in progress and some simplification of the situation is necessary to enable their construction. Once certain parameters are understood, other parameters can brought in. For instance, recent attempts have been made to integrate Lambeck's (1993a,b; 1995) models with new offshore sea level data and models of Holocene tidal regime changes in the western North Sea (Shennan et al, 2000b) while previously ignored factors such as lateral variation in the mantle are now being incorporated into the latest models (Lambeck et al, 2002b). In addition, the criticisms of McCabe (1997) can been countered by questioning his interpretation of sedimentological evidence rather than altering the GIA model (Lambeck & Purcell, 2001). In any case, once the basic structure of the models has been constructed, refining them is possible as more evidence becomes available. This can be seen with respect to the ice models (e.g. Johnston & Lambeck, 2000), and earth rheologies. Note the differences in mantle parameters between Lambeck's (1993a,b; 1995) earlier models and the more recently updated ones (e.g. Lambeck et al, 1998). The former make use of a lithosphere 65km thick, a upper mantle of $(4\text{-}5)\text{x}10^{20}$ Pa s and a lower mantle of one magnitude greater viscosity, while the revised parameters are respectively; 65-85km, $(3\text{-}4)\text{x x}10^{20}$ Pa s and one magnitude greater. Indeed, the need for refining and improving the models is recognised by the modellers themselves. Peltier, for example has mentioned that the latest fully developed numerical models (ICE-4G(VM2)) constructed by the Toronto group are not exact and incontestable representations of glacially induced sea level change (Peltier, 2002a).

However, in the absence of reliable relative sea level records going back to the LGM, these predictive models represent the only way of reconstructing formerly exposed continental shelves and thus placing the relevant archaeology within a reasonably accurate palaeogeographic context.

2.5 Comparing Sea Level Data and Palaeo-geographic Reconstructions

2.5.1 Introduction

The information presented in Sections 2.2 to 2.4 suggests that there is an intrinsic dilemma in the creation of regional scale palaeo-geographic maps. Fundamentally the best reconstructions would be based either on the combination of direct geological interpretation of coastal sedimentary facies and a local relative sea-level curve. However, inherent variability of sea-level change and subsequent sedimentological response makes the extrapolation of, or conflation of local sea-level curves to create a regional scale map difficult, if not currently impossible. By contrast GIA models give the regional scale but depending on the precise formulae used and the inevitably guestimated vales for rheological and loading factors result in a reduction in accuracy and a lack of agreement in palaeo-shoreline position. Further, both of these methods breakdown when looking at pre LGM events as the necessary lithological /glaciological records e.t.c. are just not available.

Section 2.5 therefore demonstrates how different sources of sea level data can result in the creation of varying palaeo-geographic reconstructions for the same times and places. Though the vertical differences in sea level between different datasets may seem small, around a few metres usually, if one takes into account the fact that a sea level change of this magnitude has the potential to flood or expose a very large area of low gradient land, the resulting landscape can look very different.

The majority of the discussion will focus on palaeo-shoreline reconstructions, however the impact of topographical and morphological changes on reconstruction will briefly be touched on. The main issue that will be considered is the magnitude of the error that exists within, and between, sources of information used in palaeo-geographic reconstructions. An appreciation of this should create a greater awareness of the accuracy and applicability of different sources of sea level data and different approaches to reconstruction. To this end, digital reconstructions of the palaeo-shorelines of North West Europe (based on a variety of data) are compared.

Highlighting the error margins inherent in a data set is not a new idea. This type of perspective has been advanced to an extent with respect to the construction of sea level curves from observed evidence. Given that sea level indicators tend to fall within a range of age and height estimates rather than a single definitive value (see Section 2.3.4), some uncertainties are likely to be involved in the creation of a sea level record for a particular time or place. The process of creating a sea level record from isolated pieces of evidence in turn is a subjective exercise of interpretation. Thus the depiction of sea level fluctuations with a single line rather than error bands or error bars provides a poor summary of the situation and represents the interpretation of the observations by the curve's author for their own purposes, rather than a completely objective picture of sea level change (See Figure 29: and Shennan & Tooley, 1987; Pirazzoli, 1991; 1996).

Previous work has also illustrated the difficulty of using inaccurate data on which to base reconstructions. Marcus & Newman (1983) outlined the problems associated with using glacio-eustatic data as the sole measure of palaeo-shoreline position, namely that it neglected the impact of other forcing factors on shoreline position, such as the tectonic and isostatic movements of the crust. This chapter will attempt to take

this sort of approach further by addressing other possible sources of inaccuracy and quantifying the margins of error.

2.5.2 Palaeogeographic reconstructions: Practical Issues

The most widely used method of palaeo-geographic reconstruction involves combining a record of sea level change - a sea level curve - with a topographic time horizon - typically present-day continental shelf bathymetry - to create a palaeocoastline map. This requires the consideration of three crucial factors:

- The choice of sea level curve used in the reconstruction
- The choice of topographic time horizon used in the reconstruction
- The resolution of the topographic time horizon data used in the reconstruction

2.5.2.1 The choice of sea level curve used in the reconstruction

There are three primary categories of sea level data that can be used in reconstruction of past landscapes: global glacio-eustatic curves; glacio-isostatic adjustment models and relative sea-level curves. As described in Section 2.2.3.1 glacio-eustatic curves provide a measure of global ocean volume change rather than global mean sea level change. Examples include the Fairbanks (1989) and Bard et al (1990) curves obtained from offshore Barbadian coral reefs (see Figure 29), oxygen isotope based curves, such as Shackleton (1987) and more recently curves based on Red Sea foraminifera (e.g. Rohling et al, 1998; Siddall et al, 2003 – see Figures 26 - 28). However, as relative sea level is controlled by a number of additional factors, such as glacio- and hydro-isostasy (Lambeck & Chappell, 2001; Pirazzoli, 1996), the use of glacio-eustatic curves tends to oversimplify palaeo-geographic reconstructions. This is especially pertinent in formerly glaciated regions such as North-West Europe, where fluctuations in ice sheet size and distribution have resulted in relative sea level varying significantly within a fairly restricted (i.e. several hundred km) area. Furthermore, the differential impact of sea level modifiers in different areas means that a degree of variation exists between eustatic curves obtained from different regions, and hence each curve will provide a slightly different approximation of global change (e.g. compare Figure 28 and Figure 30).

As we have seen in Section 2.4 glacio-isostatic adjustment models infer continental shelf exposure on the basis of mathematical models of the Earth's crustal response to shifting ice and meltwater loads (*glacio-* and *hydro-isostasy*) in conjunction with glacio-eustatic sea level change. The development of these models is inevitably an iterative process and at present there is no definitive model. The GIA models are again intrinsically limited as they do not extend beyond the Last Glacial Maximum due to a lack of data on ice sheet extents. Recently however, palaeoclimate simulations which make use of GIA model predicted shorelines have been constructed for Oxygen Isotope Stage 3 (60-24kaBP), just prior to the LGM (Barron et al, 2003, Lambeck et al, 2002b). These shoreline models were based on the Lambeck models from the LGM for north-west Europe with supplementary data for the Mediterranean (van Andel & Shackleton, 1982; Shackleton et al, 1984). However, the lack of relative sea-level curves for this period makes it difficult to assess the efficacy of such an approach. It is worth noting that Flemming (2002) has suggested that GIA reconstructions for the post-LGM and Holocene may be applicable to the final phases of each of the last major glaciations on the basis that the patterns of isostatic

depression and uplift in conjunction with glacio-eustatic sea level change at the end of each long cycle of glaciation (c. 100,000 years) and deglaciation (c. 20,000) would be similar. This suggestion has yet to be investigated in detail, yet if correct; it would prove to be of great assistance in reconstructing pre-LGM shorelines.

Finally, relative sea level curves, obtained directly from past sea level indicators (e.g. dated corals, foraminifera or archaeological material: Section 2.3.2) represent the most accurate way of reconstructing past coastlines for a particular region (e.g. Edwards, 2001; Edwards & Horton, 2000). This is because they reflect the local impact of eustatic isostatic and tectonic variables. In particular the impact of tectonic uplifts, which appears small on very short timescales (>1mm/yr), may be very significant over very long term (i.e. tens to hundreds of thousands of years) changes in palaeo-geography (Long 2003). Furthermore, when using relative sea-level curves one has to take into account the indicative meaning of the indicators. This is especially important in the case of sea level indicators which are sensitive to particular parts of the tidal cycle, such as foraminifera (Edwards, 2001; Edwards & Horton, 2000).

These curves tend to be spatially restrictive due to the unique local interactions of multiple sea level modifiers, and should therefore not be applied uncritically to large regions. Unfortunately, the vast majority of these curves do not extend back further than the Early Holocene, and even within the Holocene adverse preservational conditions mean that in many areas evidence of past sea levels does not exist locally. In these instances the only recourse is to use glacio-eustatic data or GIA models.

With respect to all the three data sources it should be remembered that all sea level data contains some inherent errors. These result from variations in the position of sea level indicators relative to sea level, and the dating techniques used on them (Shennan & Tooley, 1987; Pirazzoli, 1996). As will be demonstrated in sections 2.5.5.1 to 2.5.5.6 these '*age-height errors*' can potentially lead to significant variations in palaeo-coastline position.

2.5.2.2 The choice of topographic time horizon used in the reconstruction

All of the reconstructions reviewed as part of this exercise assumed that present day bathymetry is equivalent to the pre-transgression topography, although several did allude to the fallacy of this assumption (Coles, 1998; Shennan et al, 2000). In reality, syn-transgressive processes may have altered the coastal geomorphology, while post-transgressive processes of sedimentation and erosion could have further modified the submerged land surface in the interim (see Section 4). Conventional seismic techniques means that data can be relatively easily retrieved from sections that have been buried during the transgressive process, however, it will obviously never be possible to retrieve data on surfaces that have subsequently been eroded. Ideally, seismic investigations (in conjunction with borehole data) should be used to identify a chronostratigraphic time horizon appropriate to the period being reconstructed and this can be successfully done on local scales (e.g. reconstructions of the Arun and Solent rivers as part of PD3277, PD3543 and PD3364 respectively). However the feasibility of accomplishing this on a wide scale will require the integration of much more widespread (inevitably industry and military datasets). An interim alternative is to use incised bedrock surfaces to provide an effective maximum depth topographic time horizon. This could be used in conjunction with the bathymetric surface to provide a lower and medial limit on reconstructions (see Section 2.5.5.8). Although

these problems have been recognized before (e.g. Coles, 1998; Shennan et al, 2000), the scale of the inaccuracies and possible solutions have yet to be discussed to any significant degree. Therefore section 2.5.5.8 will attempt to rectify this situation to a limited extent.

2.5.2.3 The resolution of topographic time horizon data used in the reconstruction

The resolution of the input topographic time horizon data is also an important factor in determining what the final reconstruction actually looks like. This problem is often exacerbated when using Digital Elevation Models, as these tend to interpolate between data points on the basis of a particular mathematical or statistical algorithm (Hageman & Bennett, 2000). Lower resolution data will tend to smooth out relief, thus removing any topographic extremes. Other common problems include the creation of 'terraces' on contour lines used for interpolation and the creation of large areas of monotonous slope in low lying areas due to a lack of data points (Verhagen, 2000). The resolution of the first order topographic time horizon data (bathymetry) varies depending on the scale at which the area in question was surveyed. For example, until the advent of satellite altimetry of the ocean bathymetry, all bathymetric surveys had to be undertaken by 2D echosounder equipped survey ships. Their coverage of the seabed was uneven and surveyed areas could be as much as hundreds of kilometres apart. This data was synthesised to produce a global 5' data grid for the globe (ETOPO-5 data: Figure 35) which is freely available to all interested parties (Smith & Sandwell, 1997). However, the development, and use of satellite altimetry calibrated against the extant 2D echosounder survey data made possible the creation of a global digital bathymetric data source at 2' grid spacing (ETOPO-2 data: and Sandwell & Smith, 2003).

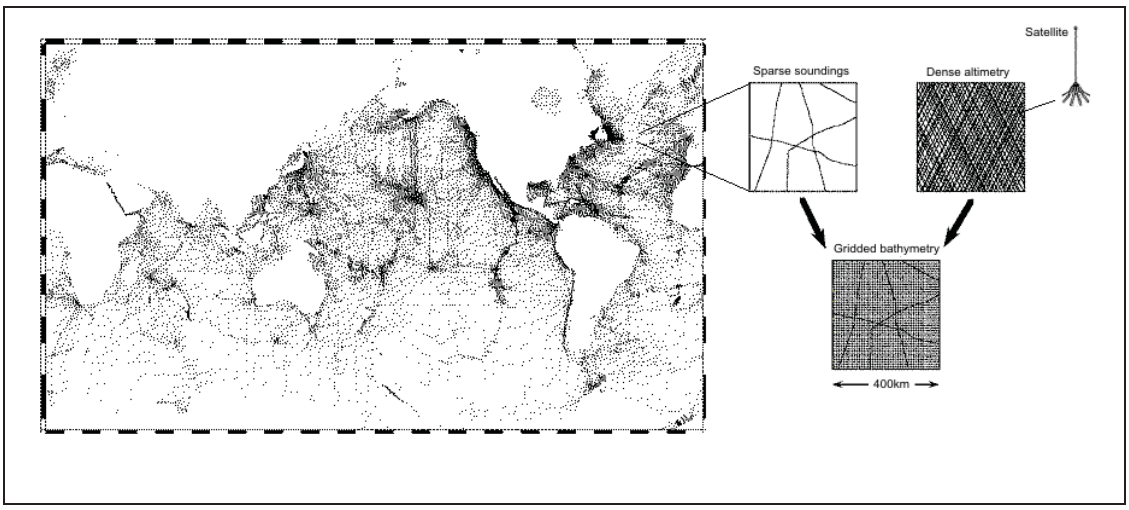


Figure 35. General principles underpinning the creation of digital bathymetric maps. Low resolution ship soundings are combined with high resolution satellite altimetry to produced a high resolution gridded bathymetric surface (modified from Sandwell & Smith, 2003)

On a regional level data sources such as the UKHO can provide bathymetric data on variable scales (from a very high resolution sub-metre bins swath bathymetry to course decametre 2D echosounder grids) depending on the location, ports and key navigational routes being generally surveyed at a much higher data density let alone more frequently. Current sub-bottom data will always be 2D in nature and thus will

tend to provide coarser spatial resolution of the bedrock horizon by comparison to swath bathymetry and in many cases the high density 2D survey data owned by the UKHO. For instance, the aggregate industry typically acquire regional sub-bottom data as part of the prospection process but rarely if ever at line spacings of less than 100 m and in places this can be as coarse as 250 to 500 m. Conversely, pipeline or installation site surveys may provide higher density data but over much more restricted spatial areas. Organisations such as the BGS do provide syntheses of sub-bottom work in the form of bedrock contour maps of certain sections of the UK continental shelf but frequently the original data density is still quite large (100's to 1000's metres).

Whatever the source, taking in to account data density will be essential for all palaeo-geographic reconstructions as low resolution data can result in misrepresentations of past topography and less obviously, differences in shoreline position (see Section 2.5.5.7).

2.5.3 Sources of Sea Level Data

For the assessment of palaeo-geographic reconstructions a wide variety of sea-level sources have been consulted. The sources can be divided into post-Last Glacial Maximum and pre-Last Glacial Maximum:

2.5.3.1 Post-Last Glacial Maximum

Bard et al (1990) is a sea level curve derived from the dating of offshore coral reefs in Barbados. It extends from the LGM till the mid-Holocene (see Figure 36). It is essentially a version of the Fairbanks (1989) curve calibrated using the Uranium-thorium (U-Th) dating method rather than conventional radiocarbon dating. The Fairbanks curve was claimed to provide a measure of global glacio-eustatic change as Barbados was deemed sufficiently far from the continental ice sheets to be minimally affected by isostatic factors, and a correction was applied to account for local tectonic uplift of c.34cm kyr^{-1} (Fairbanks, 1989). The inherent height error within this curve is estimated to be +/- 2.5m and results from the habitat range of the corals (*Acropora palmata*) used as sea level indicators (Bard et al, 1990).

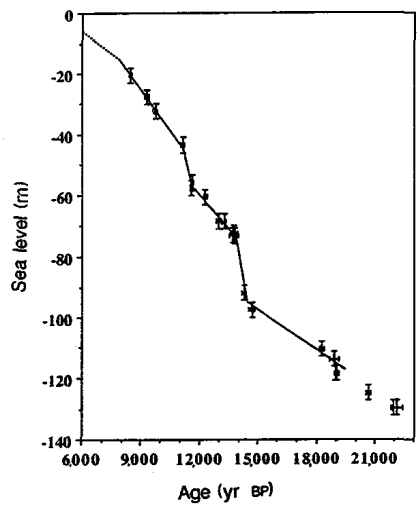


Figure 36. Global glacio-eustatic change since the LGM (modified from Bard et al, 1990).

Lambeck et al (2002b) can be considered to be a more reasonable approximation of global glacio-eustatic change than the Barbadian coral reef record. Rather than simply being based on the information from one area, it is based on 6 areas, including the Barbados reefs and Huon Peninsula terrace records, and has been corrected for isostasy. This curve also extends from the LGM to the mid-Holocene (Figure 37). To avoid confusion with another sea level curve also obtained from this article (see below), this data will be referred to in the rest of this document as Lambeck et al (2002b:Post-LGM).

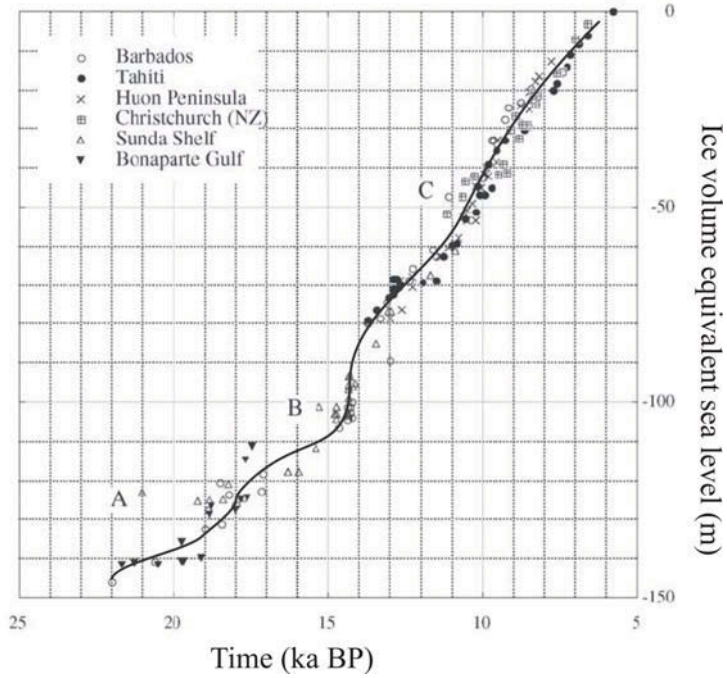


Figure 37. Post-LGM global sea level record obtained from 6 different regions and isostatically corrected (modified from Lambeck et al, 2002b)

Waller & Long (2003) is a regional relative sea level curve obtained using bio- and lithostratigraphical data for the Solent region in southern England. It is solely Holocene in scope (Figure 38).

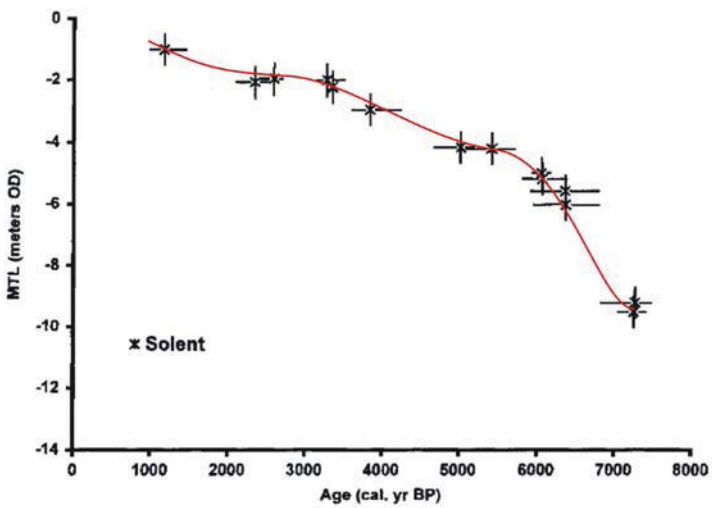


Figure 38. Holocene regional relative sea level curve for the Solent (modified from Waller & Long, 2003)

2.5.3.2 Pre-Last Glacial Maximum

Chappell & Shackleton (1986) is a sea level curve obtained from the Huon Peninsula coral terraces (Figure 39). It has been corrected for local tectonic uplift and scaled to an orbitally tuned timescale to correct uncertainties in the original U-Th dates. It represents an approximation of global glacio-eustatic change. In the original article (Chappell & Shackleton, 1986) it is referred to as HP2.

Figure 39. Pre-LGM sea level record from the Huon Peninsula (modified from Chappell and Shackleton, 1986)

Another curve obtained from Lambeck et al (2002b) is also based on relative sea level data from the Huon Peninsula coral terraces and sediment cores from the Bonaparte Gulf (Australia). For the construction of this curve, the Huon Peninsula data was re-evaluated to identify and resolve any inconsistencies in the original terrace data (Lambeck et al, 2002b). This curve has not been corrected for isostatic influences and extends from 140 kBP to the LGM (see Figure 40). To avoid confusion with the other sea level curve also obtained from this article (see above), this data will be referred to in the rest of this document as Lambeck et al (2002b: Pre-LGM).

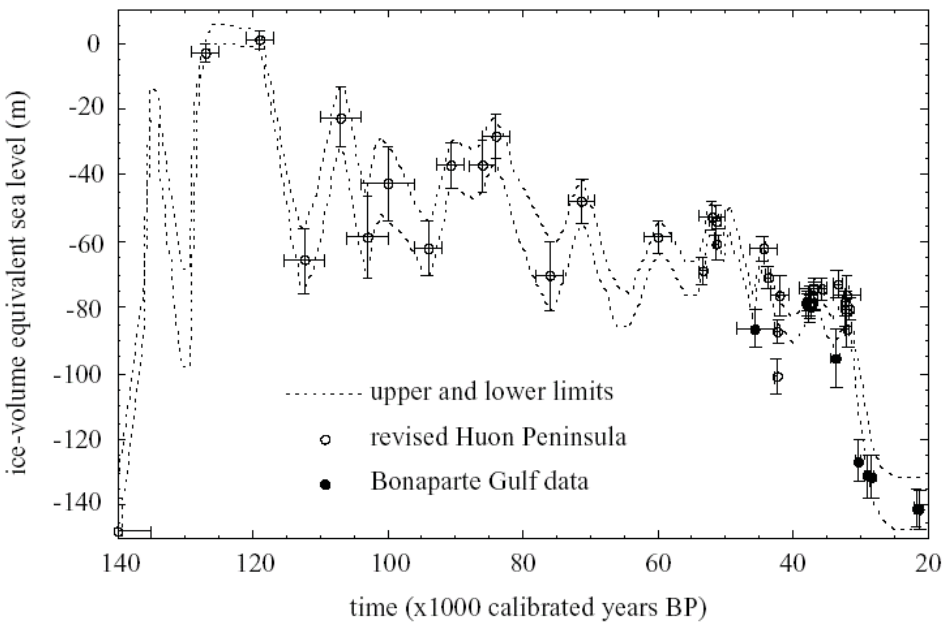


Figure 40. Pre-LGM sea level record from the Huon Peninsula and Bonaparte Gulf. Note the range of variability that exists between the upper and lower limits of the curve (from Lambeck et al, 2002b)

Chappell et al (1996) consists of a re-sampling and re-dating of the Huon peninsula coral terraces (Figure 41) to provide a closer correlation with the glacio-eustatic record derived from the oceanic foraminiferal record (Shackleton, 1987). This was undertaken after it was noted that large (up to 20-40m) discrepancies in sea level existed between the oceanic foraminiferal record of Shackleton (1987), and the HP2 curve of Chappell and Shackleton (1986).

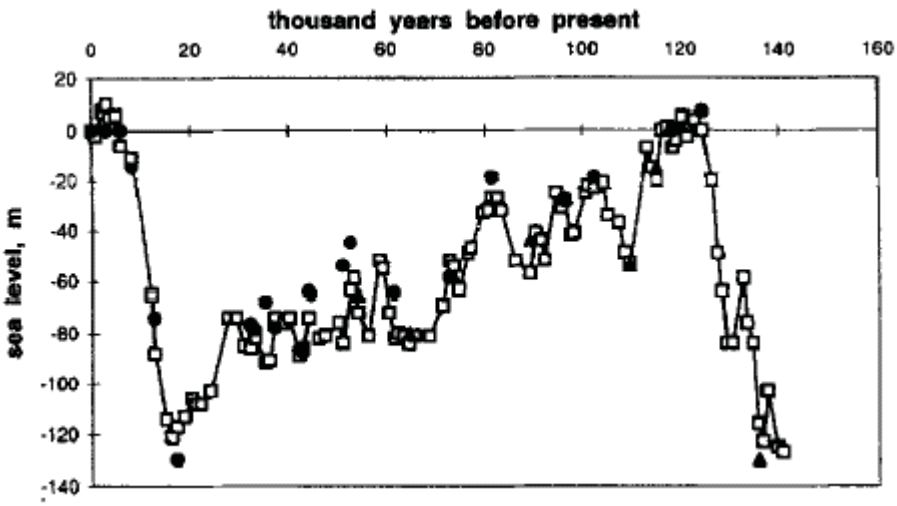


Figure 41. Pre-LGM sea level record from the Huon Peninsula after re-sampling and re-dating. Black dots refer to terrace data. White squares refer to the oceanic isotope record of Shackleton (1987) (from Chappell et al, 1996).

Rohling et al (1998) is a glacio-eustatic sea level curve derived from Red Sea salinity data (see section 2.3.2.4). The technique was used only to derive estimates of global glacio-eustatic lowstands (Figure 42). The highstands and sea level curve till 200 ka were derived from available coral reef terrace and oxygen isotope data (e.g. Pirazzoli et al, 1993; Bard et al, 1996).

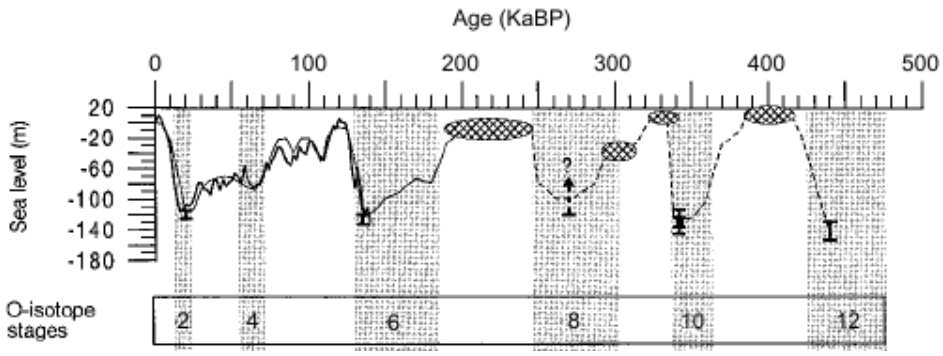


Figure 42. Pre-LGM glacio-eustatic sea level change inferred from Red Sea salinity data. Error bars represent the range of estimates of lowstands estimated from the salinity data, hatched ovals are highstand ranges derived from coral terrace and isotope data (modified from Rohling et al, 1998).

Finally, Siddall et al (2003) is the latest version for the sea level curve based on the Red Sea salinity data (Figure 43). It claims to provide much greater accuracy (+/-12m) and greater resolution (centennial scale) between 25,000 and 70,000 years ago (Figure 44).

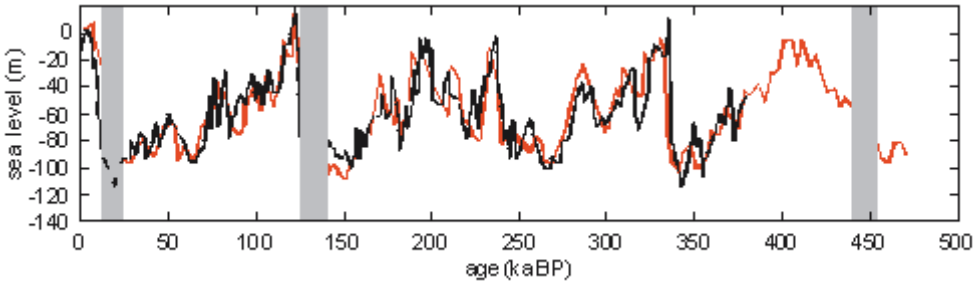


Figure 43. Pre-LGM glacio-eustatic sea level change inferred from Red Sea salinity data (from Siddall et al, 2003).

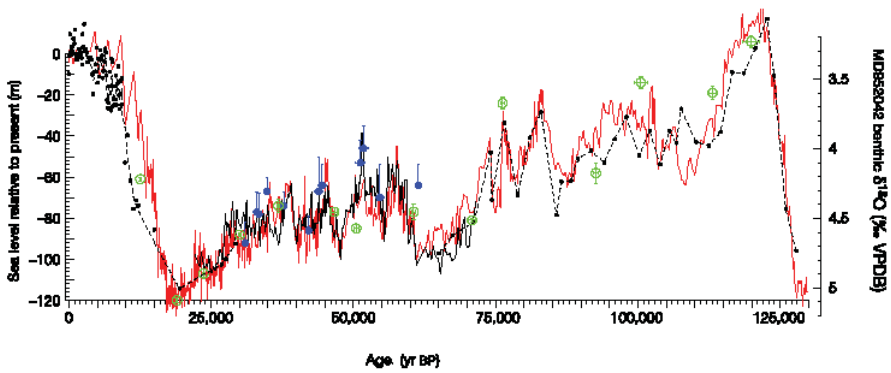


Figure 44. Reconstruction of sea level from Red Sea salinity data between 130 and 0 ka BP. The section of the curve between 25 and 70 ka BP is of centennial scale resolution. Green and blue dots represent coral terrace data included for comparison (modified from Siddall, 2003).

2.5.4 Methodology

Testing of the inherent errors involved with palaeo-geographic reconstructions was undertaken using the following methodology:

- Sea level positions were obtained from the data sets presented in Section 2.5.3.1 and 2.5.3.2 either directly from the text (e.g. Rohling et al, 1998), or tables accompanying the text (e.g. Chappell & Shackleton, 1986). If this was not possible, they were then read off the sea level curves that accompanied each article (e.g. Lambeck et al, 2002b).

- Sea level positions were then are applied to the present day bathymetry of North West Europe. The base dataset of present day bathymetry was obtained from the ETOPO-2 database of global elevations. This database was constructed using satellite altimetry and has a resolution of 2 minutes of latitude and longitude (1 minute of latitude = 1 nautical mile = 1.852km).

- The sea level positions and bathymetric data were inputted into Surfer 8, a contouring package. This allowed visual representations of approximate shoreline position to be calculated for particular times and sea level curves (Figure 45).

- Digitising the data also enabled calculations of the differences (in km) between particular shorelines to be made.

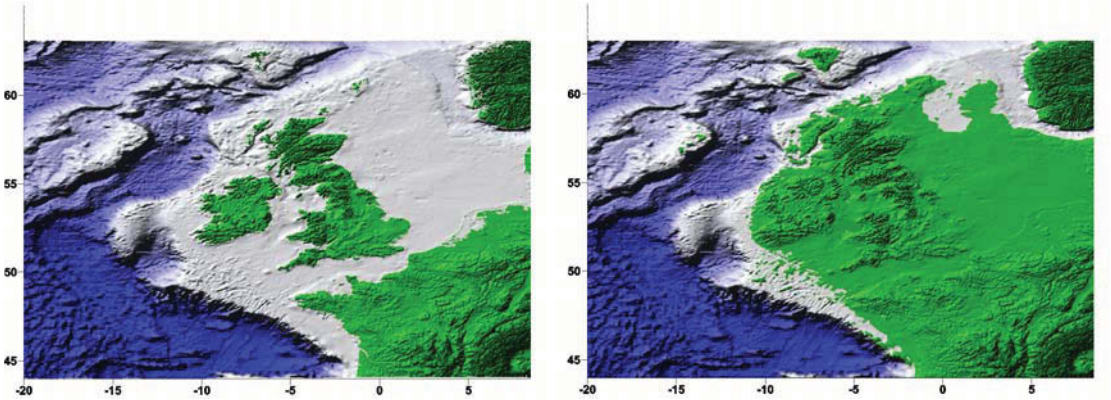


Figure 45. Example of digital reconstructions created using the Surfer 8 program. Image on left shows the present day configuration of North-West European shorelines. Image on right shows the shoreline position at –120m, the value often quoted for the global glacio-eustatic sea level fall at the LGM on the basis of Fairbanks (1989). Scales indicate latitude and longitude in decimal degrees, with negative values equating to degrees west of the meridian.

Following the above method it was then possible to make the following comparisons:

- Bard et al (1990) versus Lambeck et al (2002b: Post LGM). To provide an indication of the level of variability between two curves which purport to demonstrate global glacio-eustatic change, though it should be acknowledged that Lambeck et al (2002b: Post-LGM) provides a better approximation. It should also highlight the impact of isostatic variables, even in areas where they are considered minimal. A similar exercise will be performed for the pre-LGM data using Siddall et al (2003) and Lambeck et al (2002b: Pre-LGM). This will be demonstrated in Section 2.5.5.1.

- Rohling et al (1998) versus Bard et al (1990). To highlight the error margins within and between individual data set. – Section 2.5.5.2.

- Chappell & Shackleton (1986) versus Chappell et al (1996). To demonstrate how a seemingly secure sea level record can be modified significantly as additional data becomes available. – 2.5.5.3.

- The effect of tectonic influences on shoreline position is demonstrated by applying an estimate for long term regional (southern Britain) crustal uplift to Siddall et al (2003), and comparing it to a version of itself that has no uplift correction. This correction has been estimated at between 0.070 and 0.087 $mmyr^{-1}$ since OIS 12 (440 ka: Maddy et al, 2001), therefore a value of 0.0785 $mmyr^{-1}$ is used. – 2.5.5.4.

- The difference resulting from the use of glacio-isostatic-adjustment (GIA) models compared to eustatic curves is demonstrated by comparing palaeo-

geographic reconstructions from Lambeck & Purcell (2001) and Lambeck et al (2002b: Post LGM). – 2.5.5.5.

- The differences resulting from use of global glacio-eustatic curves and local regional sea level data is demonstrated by applying Lambeck et al (2002b: Post LGM) and a local curve (Waller & Long, 2003) to a local context: the West Solent (southern England). – 2.5.5.6.

- In situations where the palaeo-shoreline position fell within a range of values (e.g. Chappell et al, 1996), the average of the upper and lower limits was plotted.

Finally, the impact of maintaining a constant sea-level curve but a variable topographic time horizon is considered by comparing:

- Two different bathymetric surfaces; ETOPO-5 (resolution of 5 minutes of latitude/longitude) and ETOPO-2 (resolution of 2 minutes of latitude/longitude) will be applied to the same sea level data (Lambeck & Purcell, 2001: GIA model) to demonstrate the impact that resolution can have on shoreline variability. – 2.5.5.7

- Two different surfaces in the English Channel will be combined with the same sea level curve. These are the present-day bathymetric surface and the rockhead contours (base of the Quaternary sequence). This should provide an indication of the margin of error that can result from long term processes of sedimentation and erosion. - 2.5.5.8

- To facilitate comparison, the average of the upper and lower limits of the English Channel bathymetry and bedrock surfaces was also plotted. – 2.5.5.8

It must be stressed that the reconstructions produced by this exercise are approximate and should not be considered to as accurate enough to be applicable to underwater work. There are several reasons for this:

- These reconstructions primarily make use of present day bathymetry, which is itself only an approximation of the pre-transgression landsurface.

- In the instances where sea level data had to be obtained directly from sea level curves, variations of the order of a meter can result from interpretation by different individuals. In effect the scale of the curve and the thickness of the lines can lead to minor errors.

- The resolution of the bathymetric data (2 minutes of latitude or longitude, or c. 3.6km) is insufficient for more than general overviews at a continental scale.

However, the accuracy of the reconstructions is sufficient to enable comparison to be made between data sets, thus illustrating some of the practical issues involved in reconstructing submerged landscapes.

2.5.5 Results

2.5.5.1 Differences between glacio-eustatic approximations

Reconstructions for before and after the Last Glacial Maximum (LGM) will be examined separately. For the post-LGM comparison a series of c. 2000 year time steps was undertaken (Table 4):

Date (ka BP)	Sea level altitude (Bard et al, 1990)	Sea level altitude (Lambeck et al, 2002b: Post-LGM)	Maximum shoreline difference	Minimum shoreline difference
22	-130m	-146m	69.9km	0.2km
18	-112m	-125m	69.6km	1.2km
16	-102m	-113m	122km	0.6km
14	-81m	-87m	44.1km	0.3km
12	-60m	-66m	38.7km	0.6km
10	-36m	-45m	115.6km	0.3km
8	-16m	-18m	3.6km	0.2km
6	-6m	-2m	90.3km	0.1km

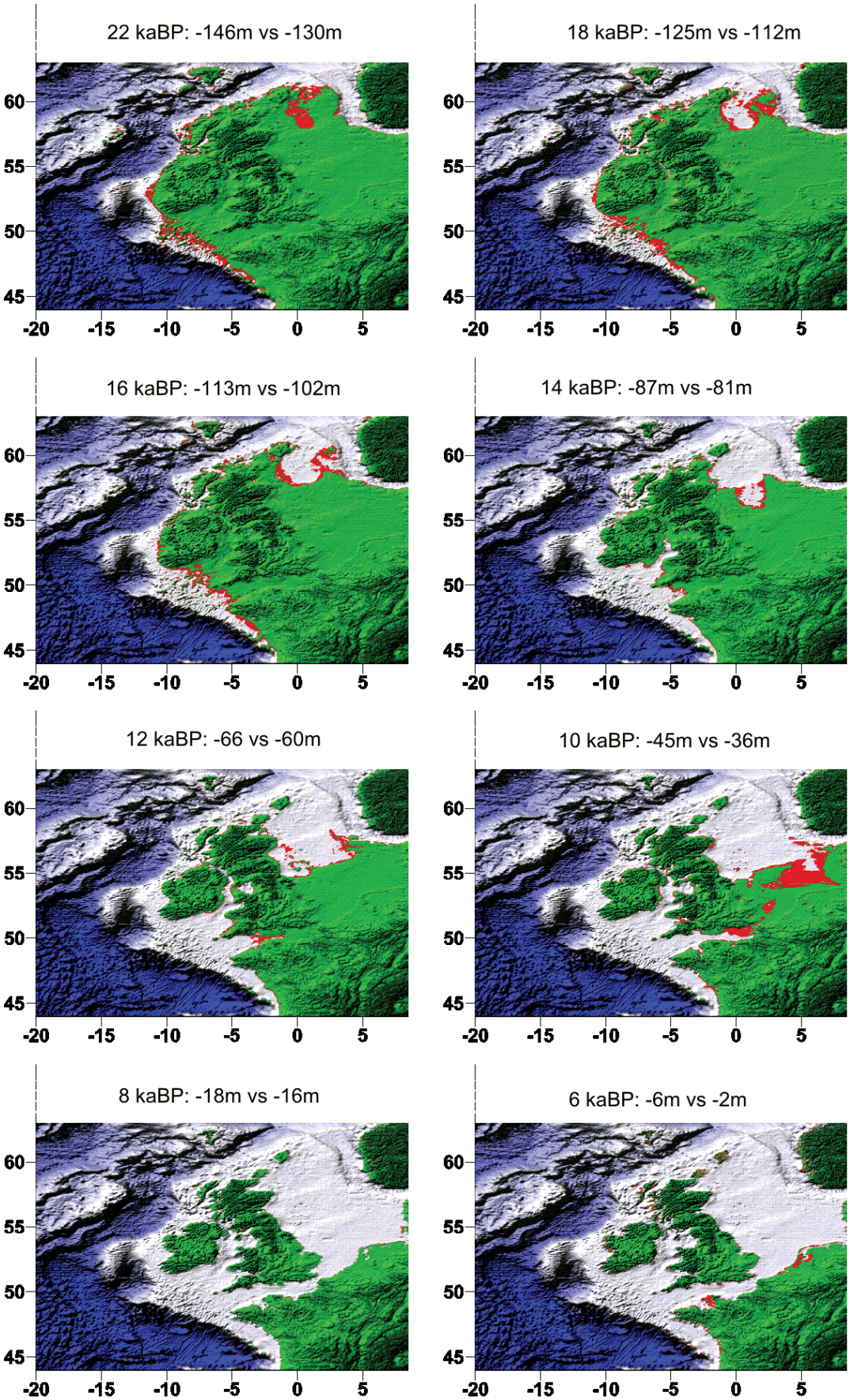
Table 4: Shoreline differences in kilometres between two different post-LGM eustatic curves applied to the bathymetry of the North-west European continental shelf

A brief glance at the maps overleaf (Figure 46) might give the impression that the differences in shoreline position are not that significant. However, quantification of the variability indicates that height differences of up to 16m can result in reconstructed palaeo-shorelines being positioned between 1 and 115 km apart.

In addition to the problem of different shoreline positions, application of different sea level curves can lead to the creation of very different coastal morphologies depending on the nature of the bathymetric surface being transgressed. Note for example the large (c. 100km by 250-300km) embayment which appears in the northern North Sea at 22ka BP if Bard et al (1990) is used, but which is not present in the same period if Lambeck et al (2002b: Post-LGM) is used. Another significant feature is the presence or absence of islands (note in particular the western edge of the shelf) depending on which curve is used. To some extent variability is reduced if the difference between shorelines is minimized, as demonstrated by the maximum value for 8 ka BP – 3.6 km. However, it also depends on the gradient of the bathymetric surface that the shorelines cross. For example, there is a maximum difference of 38 km between the shorelines for 12 ka BP, which differ by 6 metres of vertical height. However, there is a maximum difference of 90 km for 6 ka BP even though the shorelines are only 4 vertical metres apart. This is created by the existence of a large shallow water, low gradient surface extending west off the Cotentin Peninsula.

Overall, Lambeck et al (2002b: Post-LGM) consistently overestimates the palaeo-shoreline position relative to Bard et al (1990), except for the final reconstruction in which the reverse takes place.

Figure 46 (Overleaf). Comparison of palaeographic reconstructions created using different approximations of glacio-eustatic change. The red shoreline represents that of Lambeck et al (2002b: Post-LGM), and the green shoreline that of Bard et al (1990). The exception is the image for 6 ka BP, where the shoreline colours are reversed (i.e. red =Bard et al (1990); green = Lambeck et al (2002b: Post-LGM).


22 kaBP: -146m vs -130m
18 kaBP: -125m vs -112m
16 kaBP: -113m vs -102m
14 kaBP: -87m vs -81m
12 kaBP: -66 vs -60m
10 kaBP: -45m vs -36m
8 kaBP: -18m vs -16m
6 kaBP: -6m vs -2m
60
55
50
45
-20
-15
-10
-5
0
5

For the pre-LGM (65 to 33 ka BP) time steps of between 7000 and 2000 years between were digitally reconstructed and compared (Table 5):

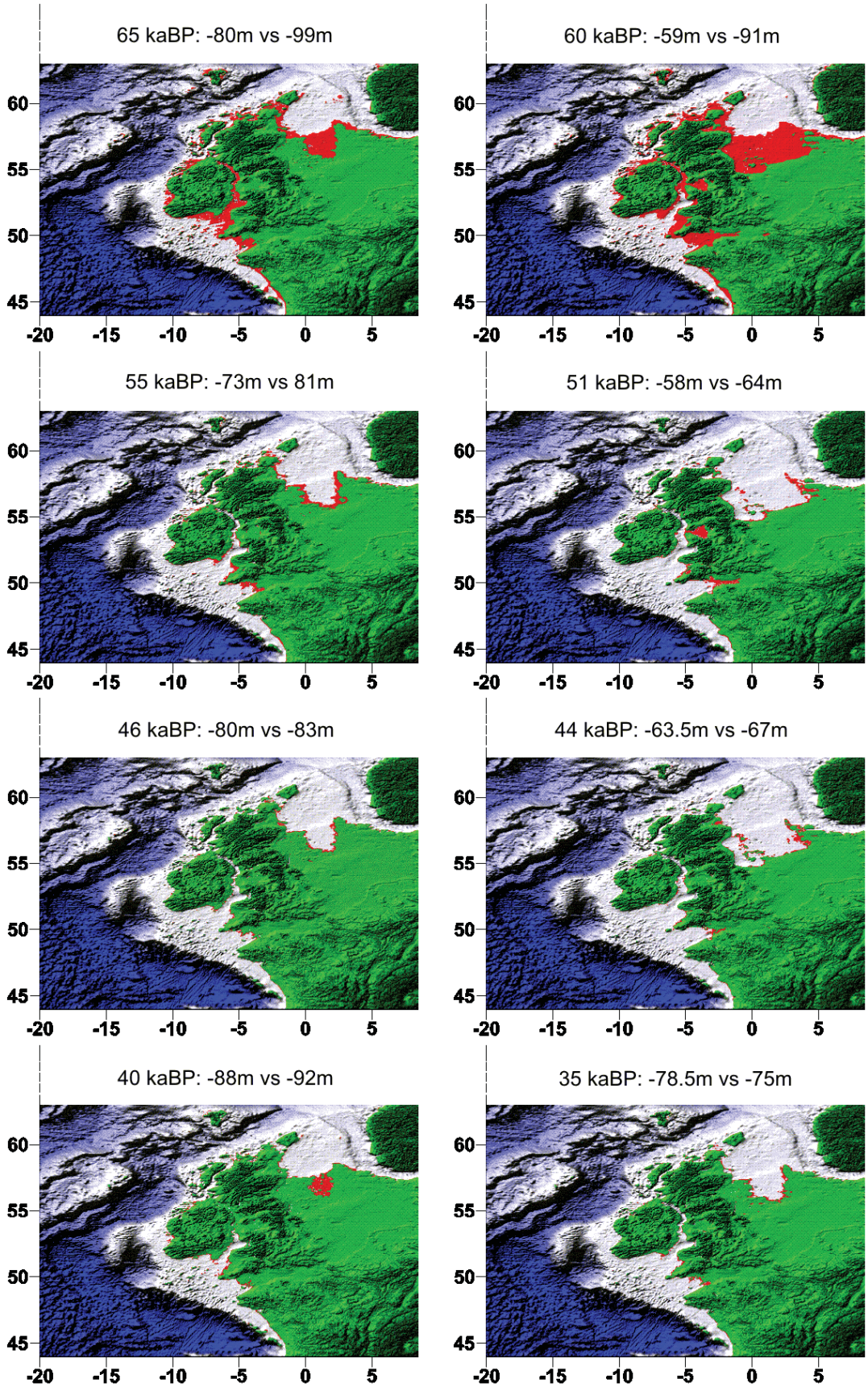
Date (kBP)	Sea level altitude (Siddall et al, 2003)	Sea level altitude (Lambeck et al, 2002b: Pre-LGM)	Maximum shoreline difference	Minimum shoreline difference
65	-99m	-80m	103km	1km
60	-91m	-59m	235km	1.9km
55	-81m	-73m	68.8km	8.3km
51	-64m	-58m	31.6km	0.3km
46	-83m	-80m	27.3km	0.4km
44	-67m	-63.5m	46.4km	0.2km
40	-92m	-88m	146km	0.4km
33	-75m	-78.5m	32.7km	0.2km

Table 5: Shoreline differences in kilometres between two different pre-LGM eustatic curves applied to the bathymetry of the North-west European continental shelf

As with the post-LGM reconstructions, major differences in shoreline position result from the application of different sea level data. This is most apparent in the reconstructions for 65 and 60 kaBP which offer radically different palaeo-coastlines. However, from then on there appears to be better overall correlation between the 2 curves, with shoreline differences ranging from 69 km to less than a kilometre. The large value of 146 km for 40 ka BP was created by the presence of a large embayment in the northern North Sea, when the –88m was used, but which was absent with the deeper shoreline (-92m). On the basis of this exercise, it seems that differences of this magnitude represent the best case scenario for this sort of data. As before, one curve (Siddall et al, (2003) in this case) consistently overestimates the other, except for one instance (35 kaBP) where the reverse situation occurs.

It is worth remembering that the data from the curves analysed above was an average of the inherent error margins. These mean that shoreline positions can potentially vary within a range. The magnitude of this range will now be examined.

Figure 47 Overleaf. Comparison of palaeographic reconstructions created using different approximations of glacio-eustatic change. The green shoreline represents that of Lambeck et al (2002b: Pre-LGM), and the red shoreline that of Siddall et al (2003). The exception is the image for 35 kaBP, where the shoreline colours are reversed (i.e. red =Lambeck et al (2002b: Pre-LGM); green = Siddall et al (2003).


65 kaBP: -80m vs -99m
60 kaBP: -59m vs -91m
55 kaBP: -73m vs 81m
51 kaBP: -58m vs -64m
46 kaBP: -80m vs -83m
44 kaBP: -63.5m vs -67m
40 kaBP: -88m vs -92m
35 kaBP: -78.5m vs -75m
60
55
50
45
-20
-15
-10
-5
0
5

2.5.5.2 Error margins within a data set

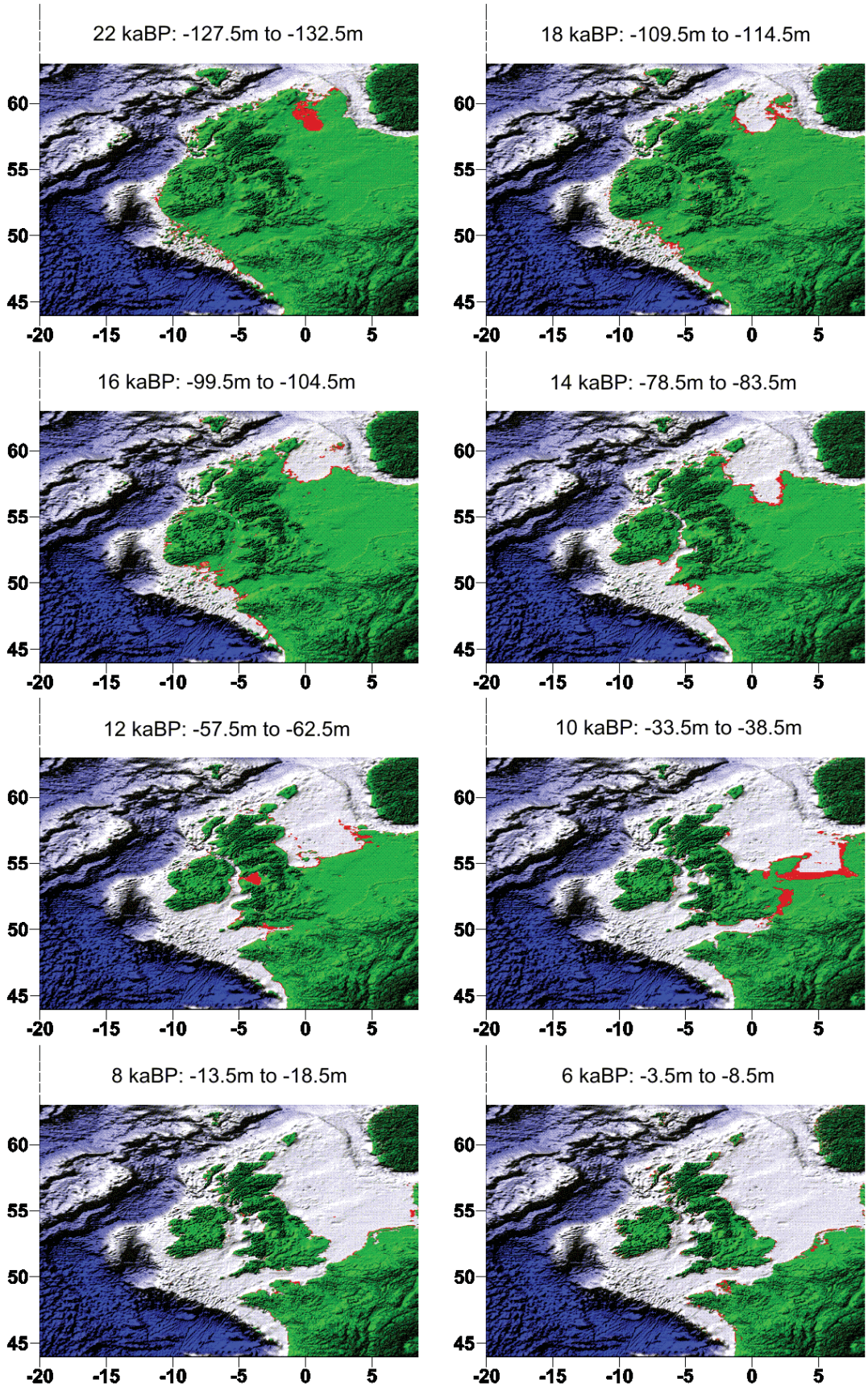
A number of the eustatic curves did provide error bars on their data. The following reconstructions aim to illustrate the impact such error bands on shoreline reconstruction for the NW European continental shelf. Again the post- and pre-Last Glacial Maximum (LGM) conditions will be examined separately. For the post-LGM comparison a series of c. 2000 year time steps was undertaken (Table 6):

Date (ka BP)	Sea level altitude range (Bard et al, 1990)	Maximum shoreline difference	Minimum shoreline difference
22	-127.5 to -132.5 m	29.5 km	0.1 km
18	-109.5 to –114.5 m	43.7 km	0.2 km
16	-99.5 to –104.5 m	59.6 km	0.2 km
14	-78.5 to -83.5 m	26.4 km	0.4 km
12	-57.5 to –62.5 m	38.1 km	0.3 km
10	-33.5 to -38.5 m	120.6 km	0.1 km
8	-13.5 to –18.5 m	12.9 km	0.1 km
6	-3.5 to –8.5 m	89 km	0.1 km

Table 6: Shoreline differences in kilometres between the maximum and minimum post-LGM eustatic sea level curve values of Bard et al (1990) applied to the bathymetry of the North-west European continental shelf

Error margins in the Bard et al (1990) sea level curve were +/-2.5 m, giving a total error range of 5 m. Overall, compared to the variations between different curves (see Section 2.5.5.1), the palaeo-shorelines, were generally closer to each other, though maximum differences were still of the order of several tens of kilometres (Figure 48). Nevertheless, significant shoreline variability was again possible in areas of low gradient local topography. Again, note the presence of the large bay at 22 ka BP when the minimum error value was used, and its absence when the maximum value was used. Further large differences are noticeable in the central North Sea at 10 ka BP and North-West England at 12 ka BP, and off the Cotentin Peninsula at 6 ka BP.

Figure 48 Overleaf. Palaeogeographic reconstructions demonstrating shoreline variability resulting from the error margins within a data set. Sea level information is from Bard et al (1990). The red shoreline represents the lowest end of the error margin (i.e. the greatest lowstand value), while the green shoreline is the highest end of the error margin (i.e. the minimum lowstand value).


22 kaBP: -127.5m to -132.5m
18 kaBP: -109.5m to -114.5m
16 kaBP: -99.5m to -104.5m
14 kaBP: -78.5m to -83.5m
12 kaBP: -57.5m to -62.5m
10 kaBP: -33.5m to -38.5m
8 kaBP: -13.5m to -18.5m
6 kaBP: -3.5m to -8.5m
60
55
50
45
-20
-15
-10
-5
0
5

For the pre-LGM single eustatic curve error comparison time steps of between 70 ka and 135 ka were taken, as these represent the major lowstand events between 440 ka BP and 135 ka BP – Table 7:

Date (ka BP)	Sea level altitude range (Rohling et al, 1998)	Maximum shoreline difference	Minimum shoreline difference
440	-150 to -128 m	74.2 km	2.2 km
340	-143 to -113 m	419.5 km	0.6 km
270	-128 to –112 m	74.6 km	1.6 km
135	-131 to –119 m	252.8 km	0.3 km

Table 7: Shoreline differences in kilometres between the maximum and minimum post-LGM eustatic sea level curve values of Rohling et al (1998) applied to the bathymetry of the North-west European continental shelf

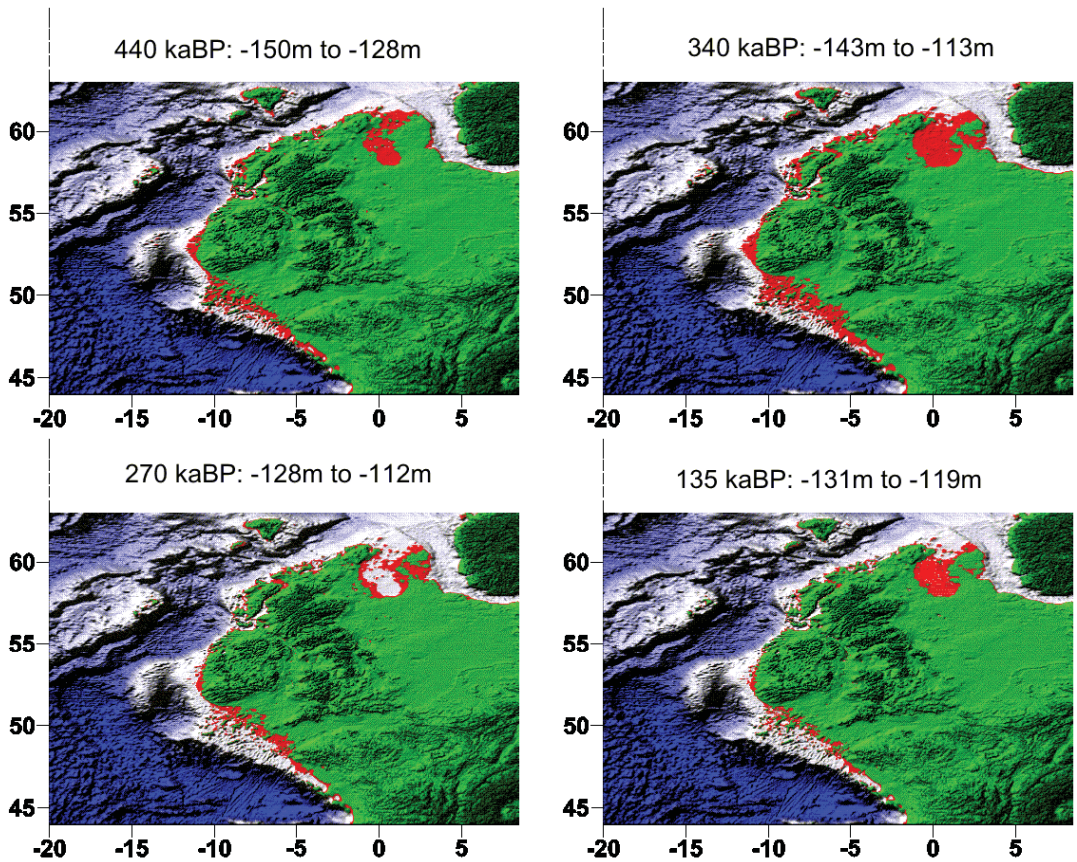


Figure 49. Palaeo-geographic reconstructions demonstrating shoreline variability resulting from the error margins within a data set. Sea level information is from Rohling et al (1998). The red shoreline represents the lowest end of the error margin (i.e. the greatest lowstand value), while the green shoreline is the highest end of the error margin (i.e. the minimum lowstand value).

In contrast to the post-LGM reconstructions, the error margins in this sea level curve are not constant, but vary from +/- 8 m to up to +/- 15 m. The fact that the error margins are, on the whole larger, means that the differences in shoreline position are generally more dramatic, up to several hundred km in some instances. These extreme results are created by the presence or absence of the northern North Sea embayment 340 and 135 ka BP (Figure 49). Also, despite the very large vertical difference between some of these shorelines, horizontal differences could still range within a few kilometres. These were created in areas where bathymetry was very steep. The more common shoreline differences tend to be of the order of several tens of kilometres. Once again, another major feature is the almost archipelago-like nature of the western shoreline when the upper end of the error margin is used. Even though the shoreline differences in this region may appear relatively small, especially in comparison to the northern North Sea situation, the differences are routinely of the order of several tens of kilometres up to over a hundred kilometres.

2.5.5.3 The effect of new data on a eustatic curve

Comparison of Chappell & Shackleton (1986) and Chappell et al (1996) show the difference in data as provided by the same authors a decade apart (Table 8):

Date (ka BP)	Sea level altitude (Chappell & Shackleton, 1986)	Sea level altitude (Chappell et al, 1996)	Maximum shoreline difference	Minimum shoreline difference
72	-36 m	-58.5 m	292.6 km	0.5 km
44	-45 m	-65 m	477.2 km	0.6 km
42	-52 m	-87.5 m	393.3 km	1.4 km
35	-65 m	-68 m	47.5 km	0.3 km

Table 8: Shoreline differences in kilometres between the two eustatic curves created by the same research but a decade apart, as applied to the bathymetry of the North-west European continental shelf

It is clear from these reconstructions (Figure 50) that significant changes in shoreline position and indeed the entire character of the coastline are possible in these circumstances. The greatest changes are once again visible in areas of low gradient, such as the central North Sea and English Channel. Shoreline variation again varies laterally but with maximum errors between c. 50 km and c. 300 km.

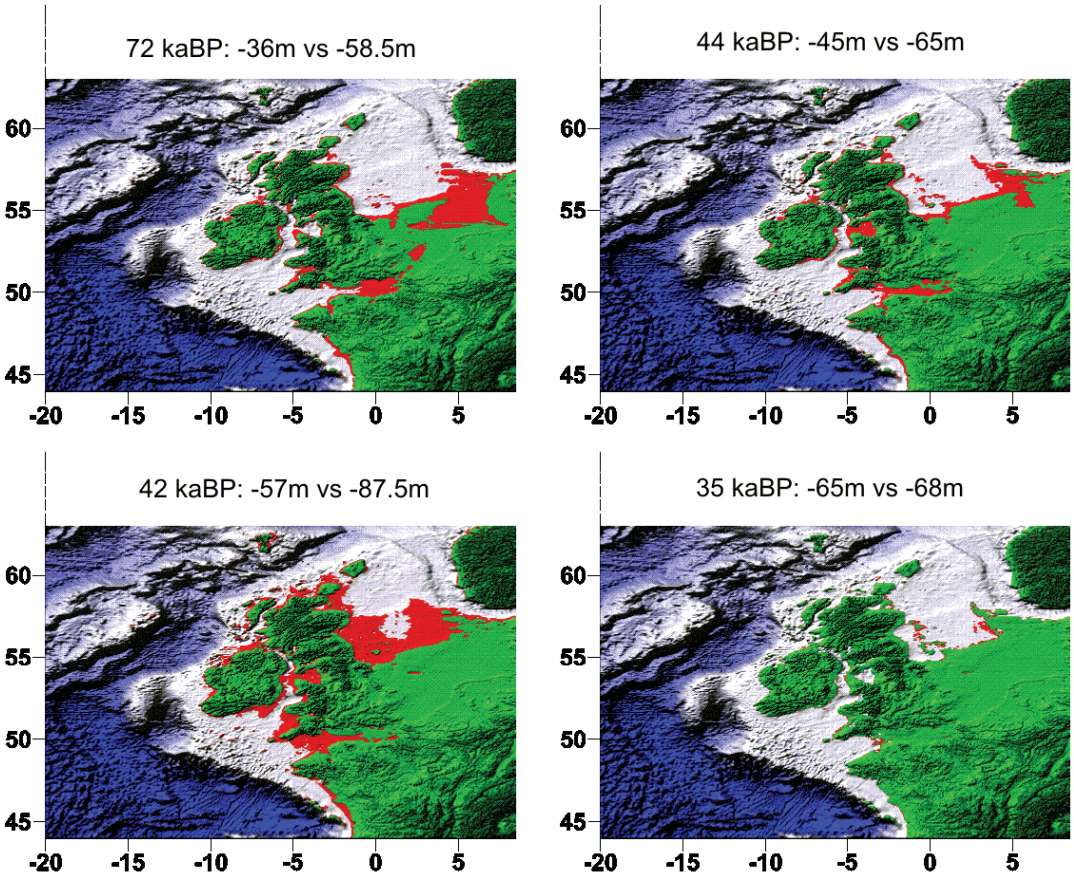


Figure 50. Differences in palaeo-geographic reconstruction resulting from the use of improved data. Green shorelines are from Chappell & Shackleton (1986), red shorelines are from Chappell et al (1996).

2.5.5.4 Long Term Tectonic influences

Tectonic influences were tested using a constant eustatic sea level curve, but with and without correction for long term uplift (Table 9):

Date (ka BP)	Sea level altitude (Siddall et al, 2003) (No uplift correction)	Sea level altitude (Siddall et al, 2003) (With uplift correction)	Degree of uplift (After Maddy et al, 2001)	Maximum shoreline difference	Minimum shoreline difference
350	-67 m	-94.48 m	27.475 m	225.7 km	1.9 km
300	-67.6 m	-91.15 m	23.55 m	178.3 km	1.3 km
250	-92.6 m	-112.23 m	19.625 m	400.1 km	1.4 km
200	-11 m	-26.7 m	15.7 m	109.6 km	0.3 km
150	-94 m	-105.78 m	11.775 m	306 km	0.7 km
100	-49.5 m	-57.35 m	7.85 m	152.5 km	0.4 km
50	-59.4 m	-63.33 m	3.926 m	187.3 km	0.6 km
22	-99.7 m	-101.43 m	1.727 m	66.3 km	0.1 km

Table 9: Shoreline differences in kilometres between the same eustatic curve (Siddall et al., 2003) but with variable long term uplift applied (no uplift vs uplift of 0.0785 $mmyr^{-1}$).

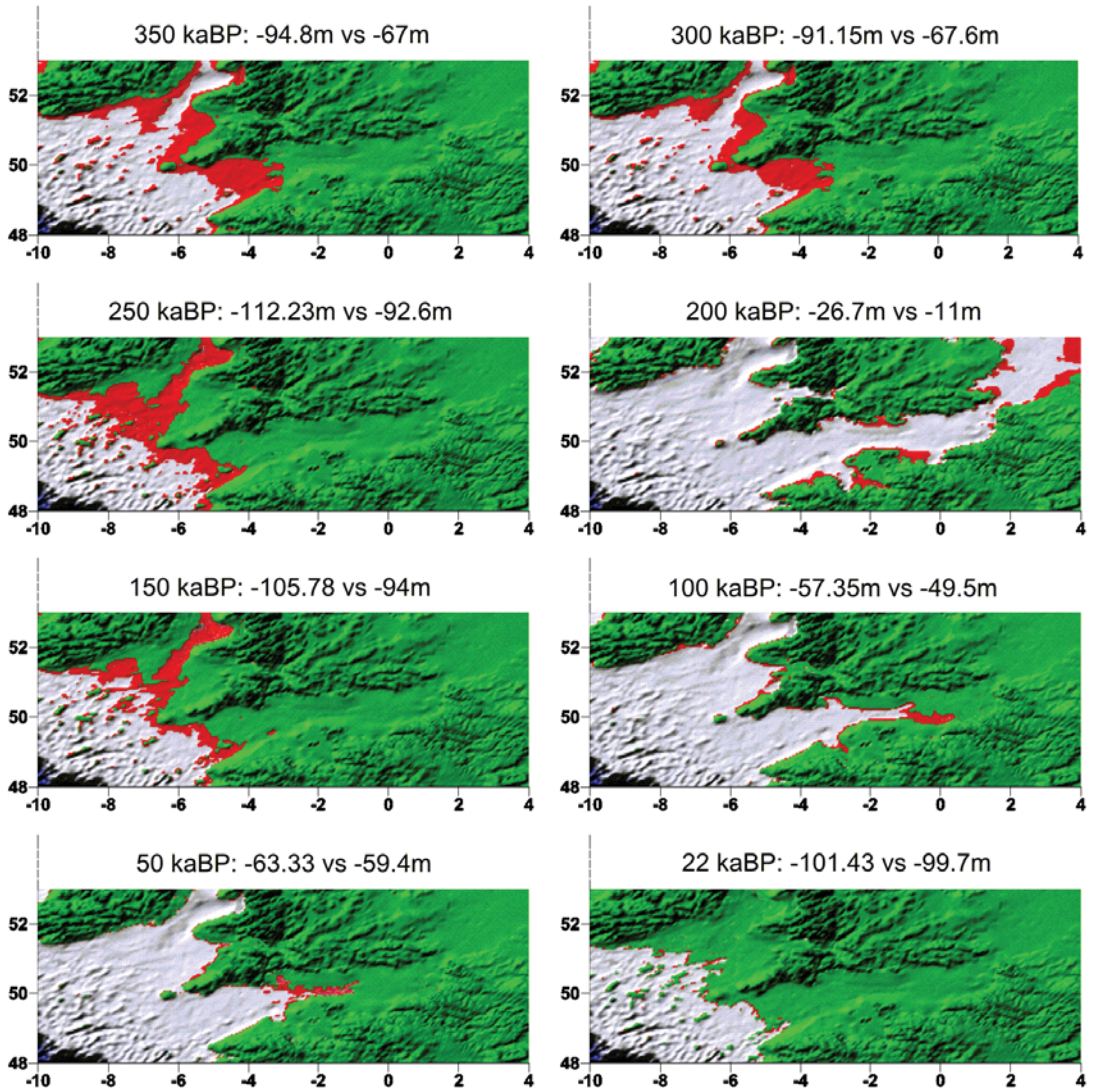


Figure 51. Palaeo-geographic reconstructions compared to highlight shoreline variability resulting from long tern tectonic uplift. Sea level data is from Siddall et al (2003). Red shoreline has is not corrected for uplift, green shoreline is corrected for an uplift of 0.0785 $mmyr^{-1}$ (Maddy et al, 2001).

For these reconstructions, only the English Channel and southern Bight of the North Sea have been digitised as the uplift correction used in this exercise has only been calculated for southern England (Figure 51). It has thus been assumed to be only regionally valid. The overall pattern is that the degree of shoreline difference is greatest the further back in time one goes. This is unsurprising as the magnitude of uplift and its long term constant nature are such that older areas have been uplifted further. However, even within the last 50,000 years differences of up to c. 180 km are possible, resulting from the English Channel taking the form of an embayment at the lower shoreline (-63 m in this case) and extending almost as far as the Isle of Wight with the upper shoreline (-59.4 m: Figure 51). Very large differences were also notable in the Celtic Sea, with shorelines characterised by the presence or absence of a large channel extending up into the Irish Sea at 250 and 150 ka BP.

2.5.5.5 GIA models versus glacio-eustatic curves

The GIA-based reconstructions were constructed by digitising and incorporating relative sea level isobases obtained from Lambeck & Purcell (2001) into the base map of North West European shelf bathymetry (Figure 52).

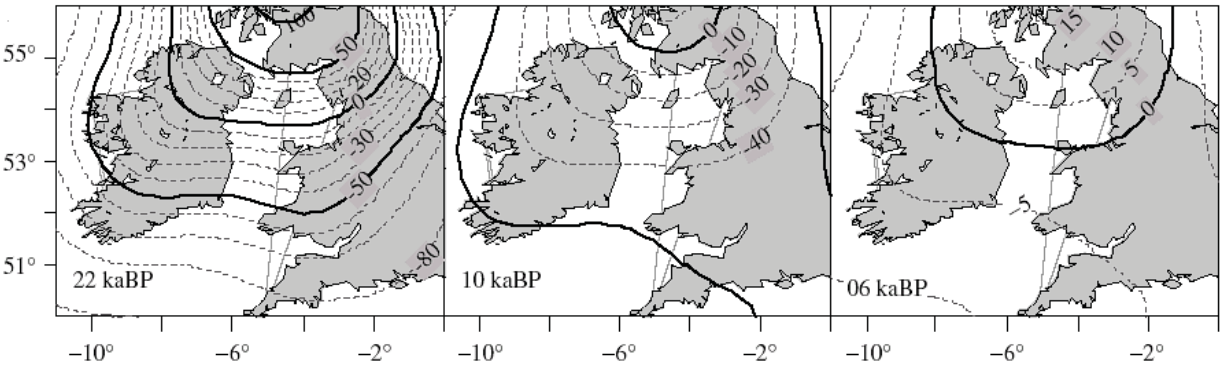


Figure 52. Predicted relative sea level change for the Irish sea region in response to contributions from the melting of the British, North-West European, American and Antarctic ice sheets (from Lambeck & Purcell, 2001)

As the information in this article was based on an uncalibrated C^{14} timescale, it was calibrated using the online CalPal program (provided by the University of Cologne Radiocarbon Laboratory http://www.calpal-online.de/index.html) to allow comparison with the calibrated sea level curves used in this exercise. The shorelines created from the GIA model were then compared with the sea level data from Lambeck et al (2002b) (both pre- and post-LGM in this instance) as this was assumed to provide a reasonable estimate of global glacio-eustatic change (Table 10 and Figure 53).

Date (uncal ka BP)	Calibrated date (ka BP)	Sea level altitude (Lambeck et al, 2002b)	Maximum shoreline difference	Minimum shoreline difference
22	25.6	-146 m	560.8 km	33.7 km
10	11.5	-62 m	289 km	7.1 km
6	6.8	-8 m	19.1 km	0.1 km

Table 10: Shoreline differences in kilometres between the GIA reconstruction of Lambeck & Purcell (2001) and the eustatic curve of Lambeck et al (2002b).

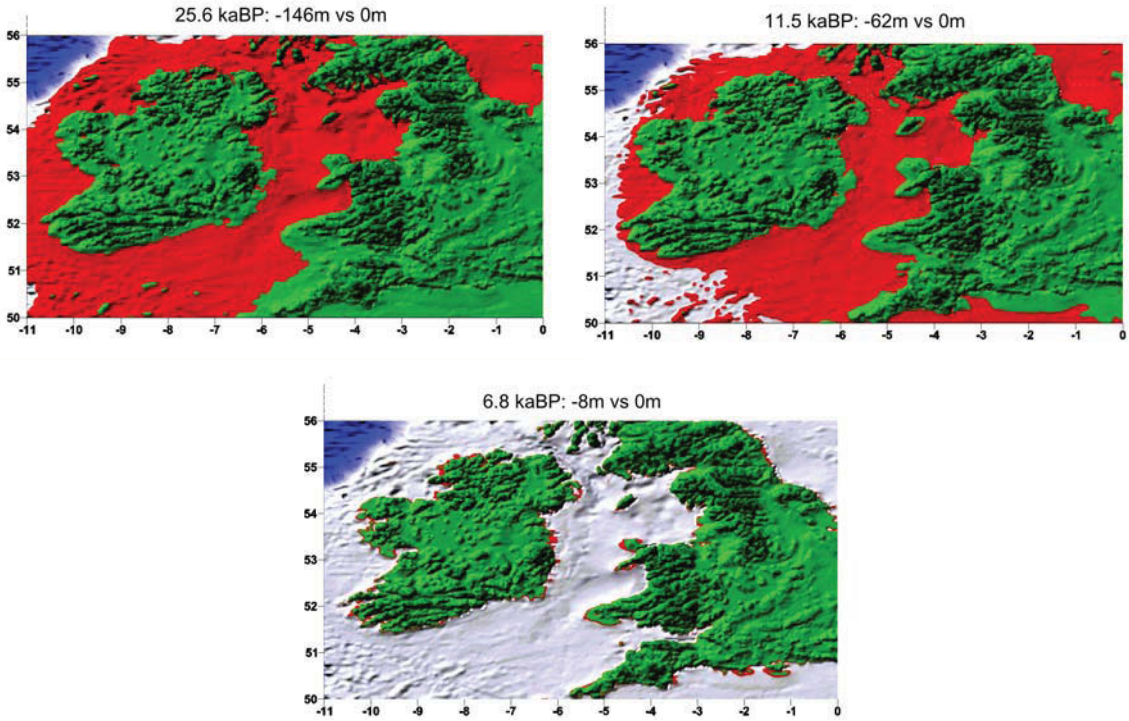


Figure 53. Comparison of palaeo-geographic reconstructions created with GIA models and global glacio-eustatic curves. Green shoreline represents the GIA model-based reconstruction of Lambeck & Purcell (2001). Red shoreline represents glacio-eustatic curve of Lambeck et al (2002b). The positions of the British and Irish ice sheets have been omitted for the sake of clarity.

It is clear from this that the use of global-glacio eustatic curves as the sole basis for a palaeo-geographic reconstruction will lead to significant errors in areas under the influence of major continental ice sheets (Figure 53). The errors are most pronounced when the sheets are large and accreting or melting. This is because the changing distribution of ice mass can cause the crust to uplift or downwarp significantly over a timescale of several thousand years. The errors are such that differences in shoreline position are not simply a case of movement of several kilometres in one direction. Rather, an entirely different picture of the palaeo-landscape is possible depending on whether isostasy has been taken into account. Note in Figure 53 that current islands of Britain and Ireland are part of a single landmass until the early Holocene, if the glacio-eustatic based reconstruction is to believed. In reality, isostatic factors would have meant that they were separated far earlier. Overall, the differences are greatest when the ice sheets are at their greatest extent (25.6 ka BP in this instance). However, there still are significant differences when the ice sheets have largely disappeared. This is the result of the lag between the isostatic response and ice sheet growth and decay (11.5 ka BP). Finally, several thousand years after the ice load has disappeared, on-going isostatic adjustments have now slowed down to the point where closer correlations to glacio-eustatic sea level change can be noticed (6.8 ka BP).

2.5.5.6 Regional relative sea level curves versus global glacio-eustatic curves

In order to compare the actual ability of a eustatic curve to predict local sea level variation and ultimately shoreline position, the eustatic curve of Lambeck et al (2002b:Post-LGM) was compared with the local relative sea level curve of Waller & Long for the Solent region (Table 11 and Figure 54). A 2D bathymetric dataset provided by New Forest District Council from the mouth of the Lymington river was used to provide the stratigraphic time horizon.

Date (ka BP)	Sea level altitude (Waller & Long, 2003)	Sea level altitude (Lambeck et al, 2002b)	Maximum shoreline difference	Minimum shoreline difference
7	-9m	-9m	0m	0m
6	-5m	-2m	1125m	23m
5	-4m	0m	>1600m	34m

Table 11: Shoreline differences in metres between the GIA reconstruction of Lambeck et al (2002b) and the local relative sea level curve (Solent region) of Waller & Long (2003).

The reconstructions depicted in Figure 54 demonstrate that different shoreline position can result depending on whether regional relative sea level curves or global eustatic curves are used. Overall, the differences are the smallest out of all the data sources analysed so far, ranging from over 1.6 km (5 Ka BP) to no difference at al (7 Ka BP). These differences are the cumulative result of local isostatic, tectonic, geoidal, steric & sedimentary modifications to the global glacio-eustatic signal. It should be noted that the impact of these modifications will vary over time and space. For instance, in this area (West Solent) isostatic adjustment will have been less than that of an area close to the Last Glacial ice sheets. In addition, by this stage in time (i.e. mid-Holocene) the magnitude of the local isostatic response had probably decreased to the point where global glacio-eustasy dominated the local trend of relative sea level change.

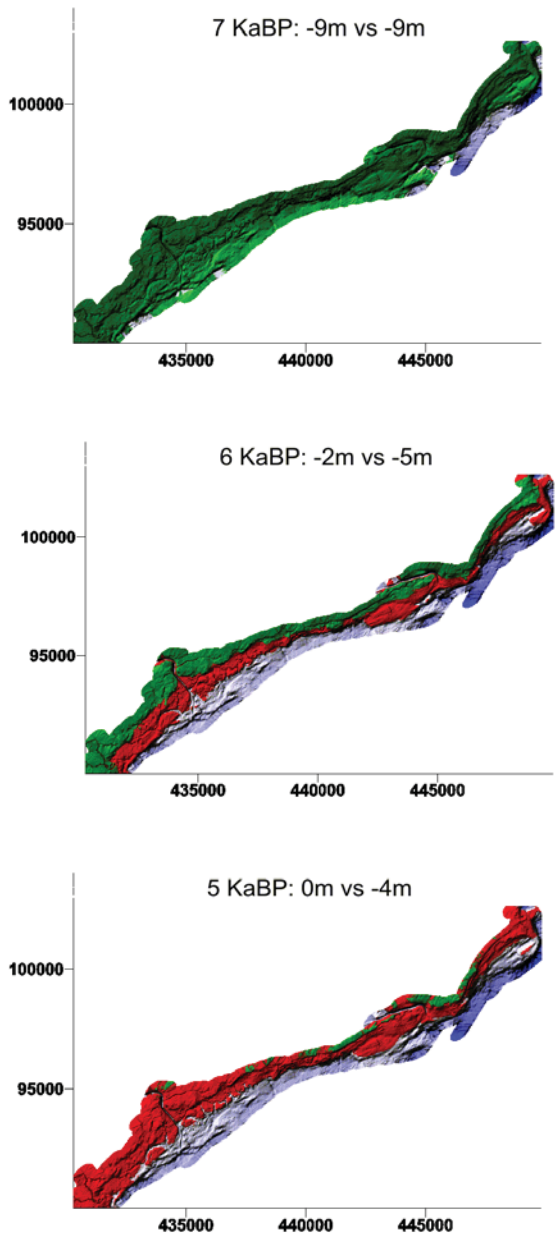


Figure 54. Comparison of palaeogeographic reconstructions created using regional relative sea level curves and global glacio-eustatic curves. Green shoreline represents glacio-eustatic curve of Lambeck et al (2002b:Post-LGM). Red shoreline represents the regional relative sea level curve of Long & Waller (2003).

2.5.5.7 Issues of Resolution

The impact of input data resolution on palaeo-geographic reconstructions is illustrated in the image below (Figure 55). In this instances, shorelines from a GIA model (Lambeck & Purcell, 2001) have been applied to a high resolution (ETOPO-2) and low resolution (ETOPO-5) bathymetric surface.

In these images it is possible to see distinct differences in palaeo-shoreline position. According to the ETOPO-5 reconstruction, there appears to be more land off south-east Ireland, west Wales and south-west England. In addition, the ETOPO-2 reconstruction has a number of islands in the Celtic Sea, which are not present in the ETOPO-5 reconstruction.

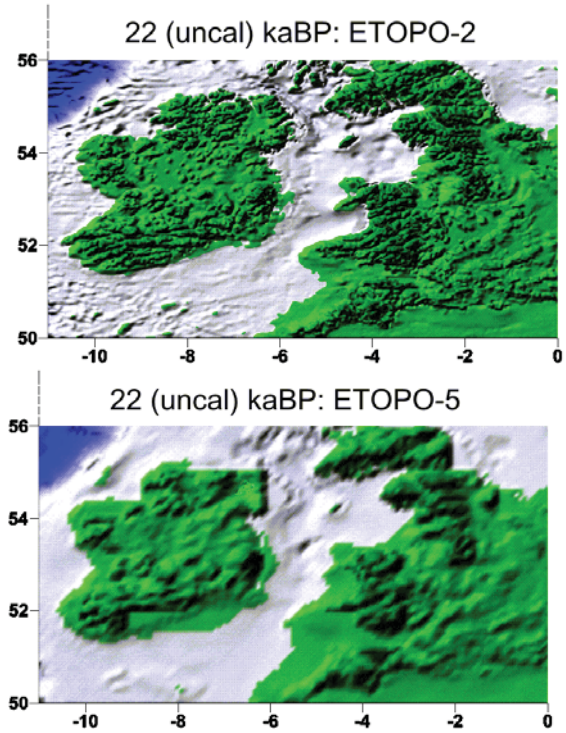


Figure 55. Comparison of palaeo-shoreline position variability resulting from the use of different resolution bathymetric and topographic data. Image on the left uses ETOPO-2 data, image on the right uses ETOPO-5 data. Shorelines are from Lambeck & Purcell (2001) – 22 (uncal) kaBP. For the sake of clarity, the British Ice Sheet has been removed.

2.5.5.8 Topographic time horizon variability

The area chosen in order to look at the impact of picking an alternative topographic time horizon to the seabed bathymetry was located between 2 degrees East and 2 degrees West, and between 50 and 51 degrees North, an area encompassing the Dover Straits to the Isle of Wight (Figure 56).

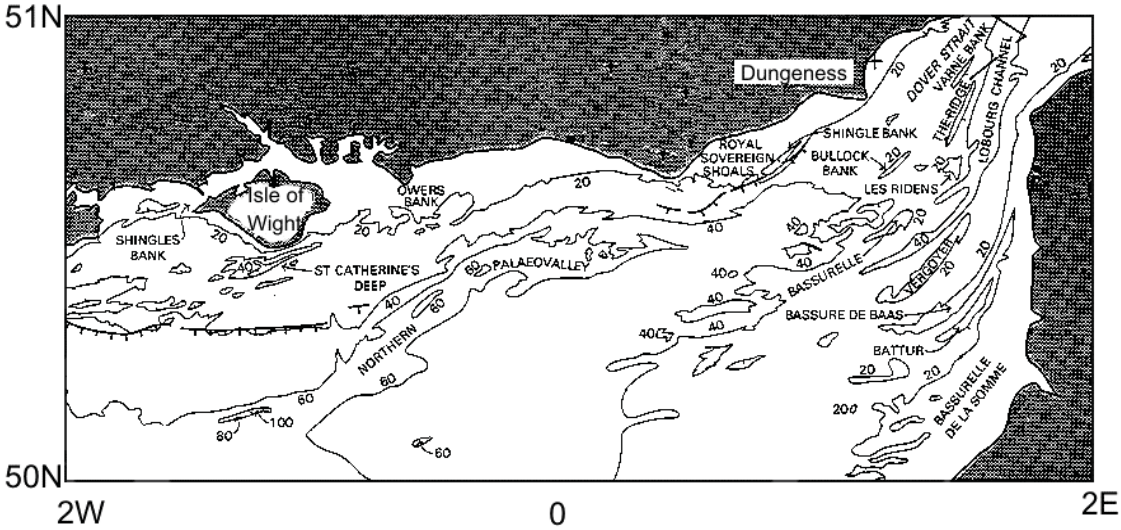


Figure 56. The present day bathymetry of the Wight-Dungeness area. Depths are in metres below sea level (modified from Hamblin et al, 1992).

The surfaces compared were the modern bathymetric surface (Figure 56) and the depth to the base of the Quaternary sequence (rockhead contours: Figure 57). This latter surface was obtained by combining bathymetric depths with measurements of sediment infill. The information was digitised from the British Geological Survey's 1:250000 series maps of seabed sediments and Quaternary geology.

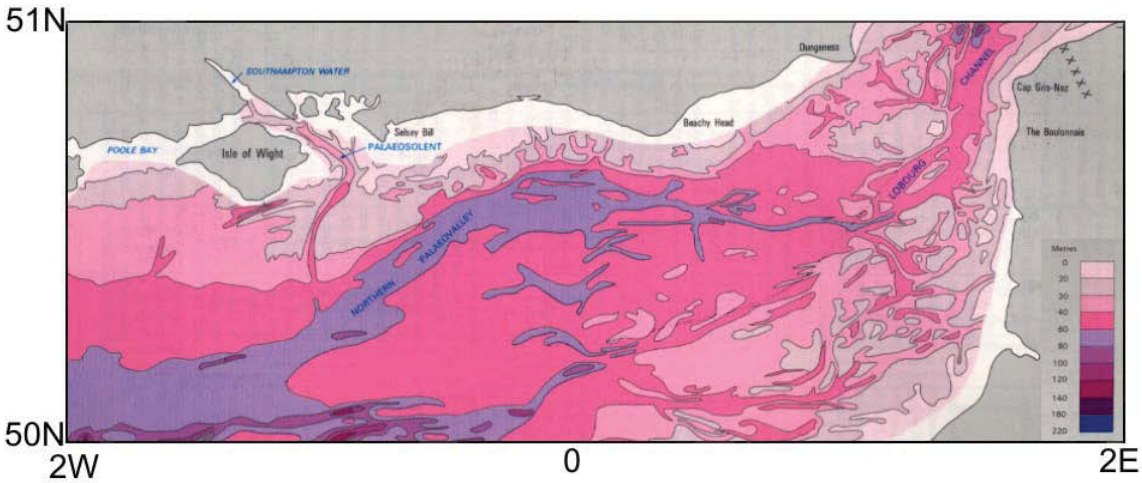


Figure 57. Depth (in metres) to base of the Quaternary – bedrock - in the Wight-Dungeness area (modified from Hamblin et al, 1992).

For ease of analysis the area was split in two down the 0° line of longitude. The area to its west will be known as the Wight area, and the area to its right will be known as the Dungeness area (Figure 58).

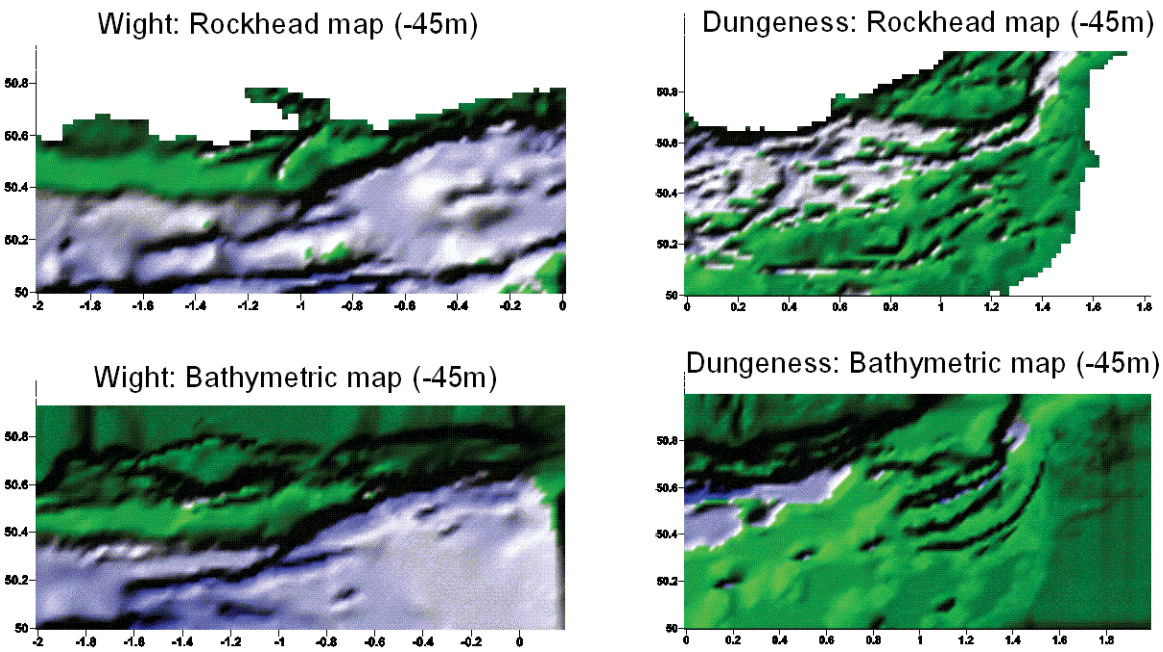
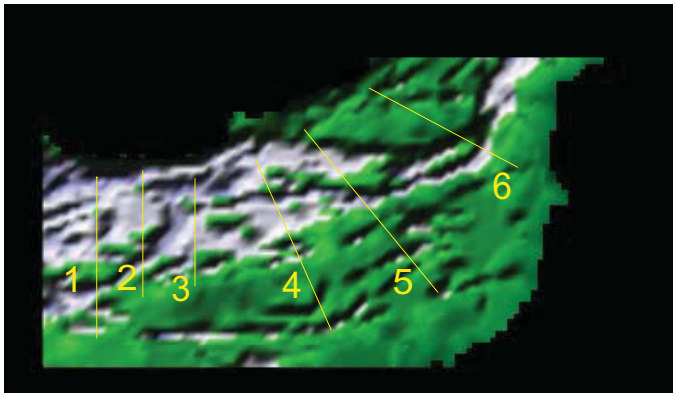


Figure 58. Digital maps of bathymetric and rockhead surfaces for the Wight and Dungeness regions. Sea levels have been nominally placed at –45m. X and Y axes show latitude and longitude in decimal degrees.

A visual inspection of these reconstructions would indicate that some differences in shoreline position are apparent, most obviously in the Dungeness region, where they may be of the order of several kilometres to tens of kilometres. Differences do occur in the Wight area, but are more subtle and are less obvious from these reconstructions. Consequently 6 transects running from north to south were taken across each of these two areas, for both the bathymetric surface and the rockhead contours. These have been compared in Figures 59 and 60 and provide a better indication of the degree of difference between the two surfaces.



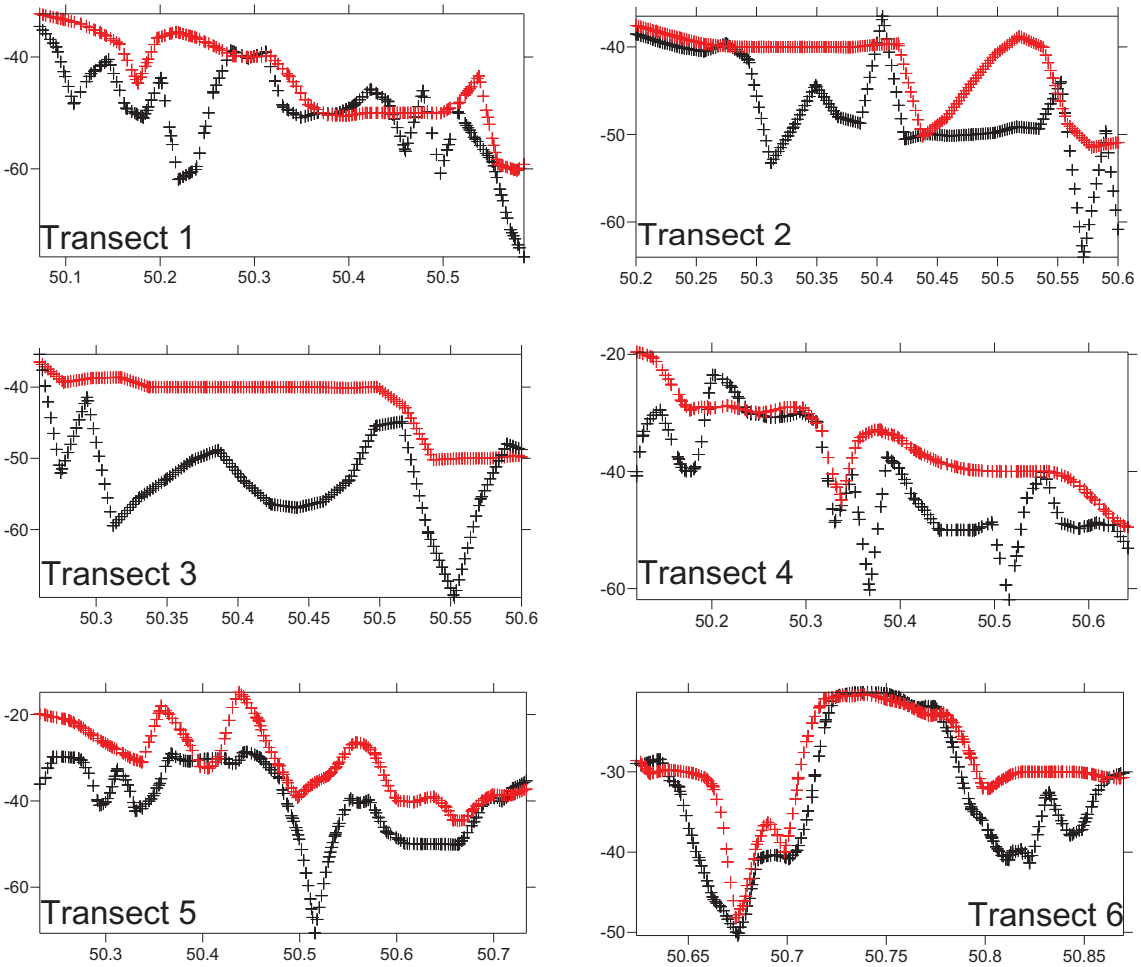


Figure 59. Comparison of rockhead and bathymetric surfaces along selected transects in the Dungeness region. The transect locations are provided by the digital bathymetric surface at the top of the figure. On the graphs, red lines indicate the modern bathymetric surface and black lines the rockhead surface. Y axis depicts metres below present sea level. X axis shows latitude in decimal degrees.

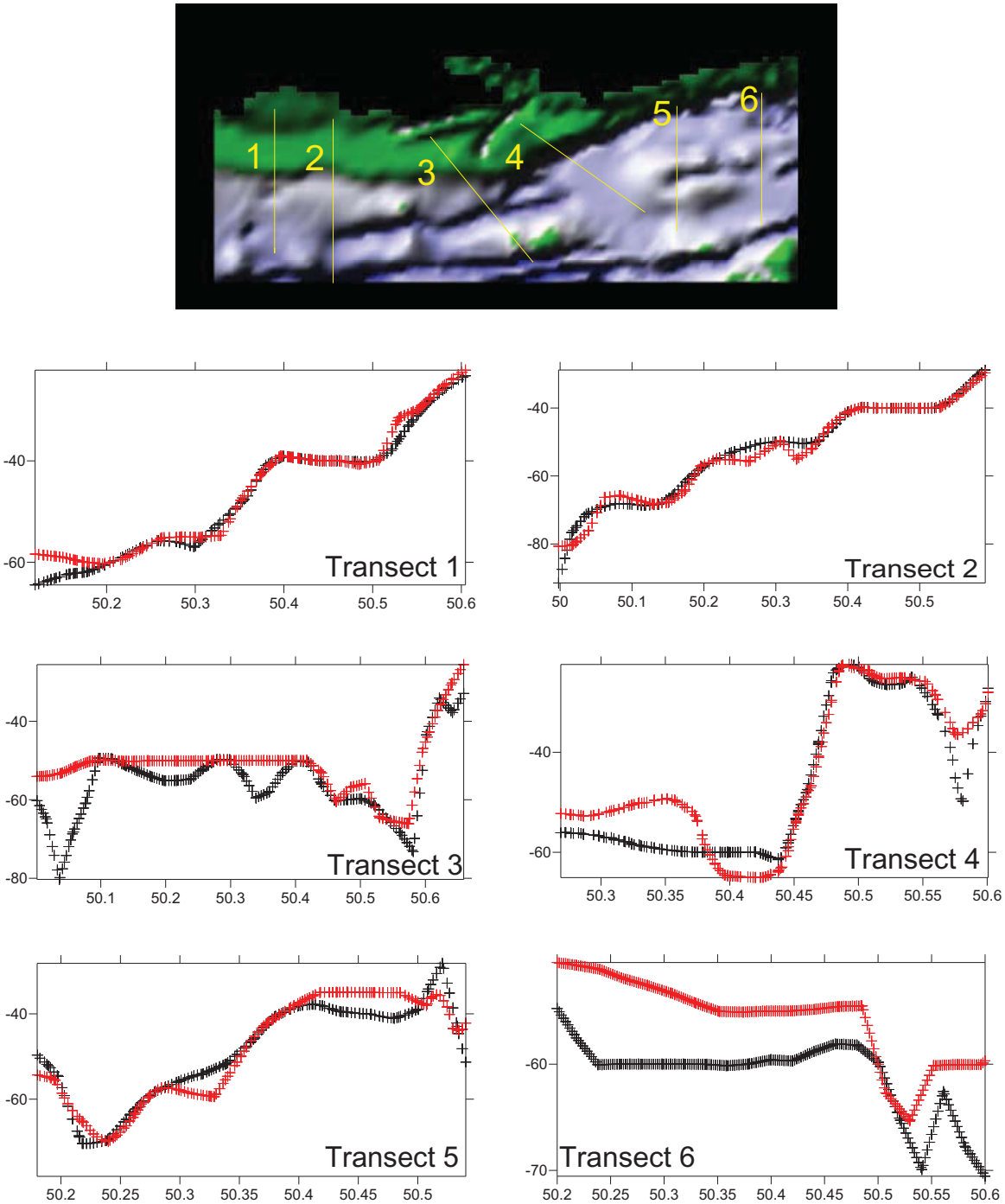


Figure 60. Comparison of rockhead and bathymetric surfaces along selected transects in the Wight region. The transect locations are provided by the digital bathymetric surface at the tope of the figure. On the graphs, red lines indicate the modern bathymetric surface and black lines the rockhead surface. Y axis depicts metres below present sea level. X axis shows latitude in decimal degrees.

It is apparent from Figures 59 and 60 that the difference between the modern bathymetric surface and the bedrock horizon can vary significantly over a relatively restricted region (in this case c.432 km by c.108 km). The vertical differences between the surfaces can range from greater than 20m (Dungeness Transect 3) to no appreciable differences on this scale (Wight Transect 1). This can lead to variations in the actual topography of the reconstructed landscape. In this instance, it is apparent

that a number of palaeo-valleys appear to be infilled (e.g. Wight Transect 3, Dungeness Transect 1) while the close correlation between the surface most likely relates to the transgressive erosion of the former landsurface by the post-LGM sea level rise (Reynaud et al, 2002). Overall, differences appear to be more significant in the Dungeness region than the Wight region, where only 1 transect (number 6) revealed consistent significant differences between the two surfaces along most of the transect. The use of different surfaces also resulted in the creation of different shoreline configurations (Figures 61 and 62 and Tables 12 and 13). These palaeo-shoreline maps were produced by overlaying 3 contour maps of the Wight and Dungeness regions; the modern bathymetric surface, the rockhead surface and an intermediate surface created by calculating the mid-point altitude between the two other surfaces for each nodal point. Mean sea-level was dropped by -25 m, -35 m and – 45 m respectively.

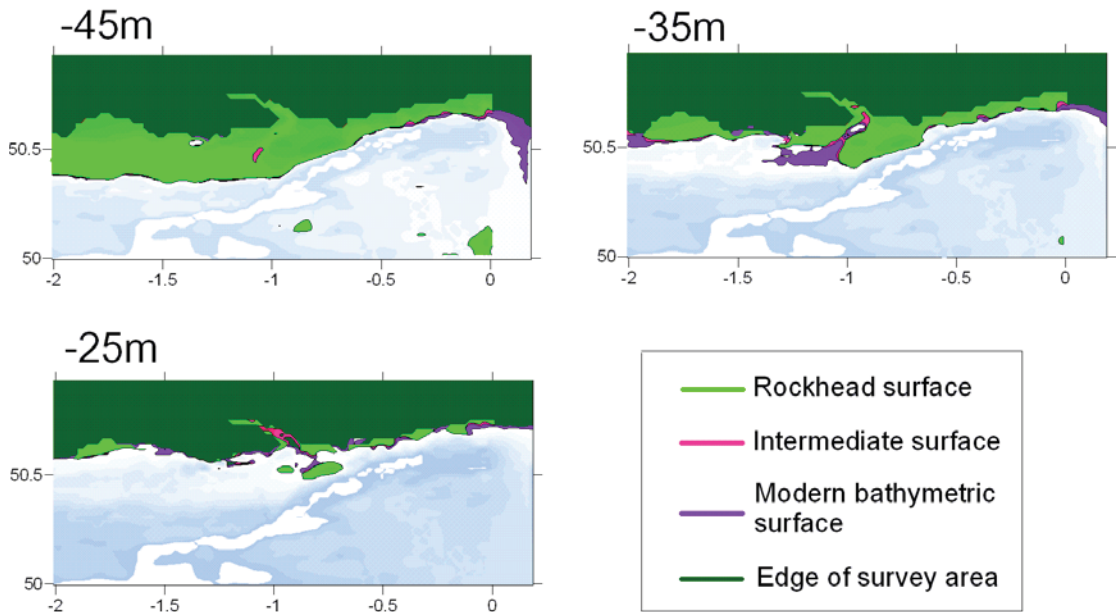


Figure 61. Overlaid shorelines produced by different bathymetric surfaces for the Wight region. The edge of the survey area indicates the limit of the known rockhead contours, and hence for this exercise the limits of the data. To some extent it approximates the present shoreline

Depth (mbsl)	Maximum shoreline distance (rockhead to intermediate)	Maximum shoreline distance (intermediate to bathymetry)	Maximum shoreline distance (rockhead to bathymetry)	Minimum shoreline distance (rockhead to intermediate)	Minimum shoreline distance (intermediate to bathymetry)	Minimum shoreline distance (rockhead to bathymetry)
-45	1.4 km	1.2 km	2.6 km	-	-	< 0.1km
-35	18.2 km	18.2 km	36.4 km	-	-	< 0.1km
-25	1.1 km	3.4 km	4.5 km	-	-	< 0.1km

Table 12: Shoreline differences in metres between three topographic time horizons (bathymetry, bedrock and a mid-point surface) for the Wight region. MSL has been nominally dropped by -25 m, -35 m and – 45 m respectively.

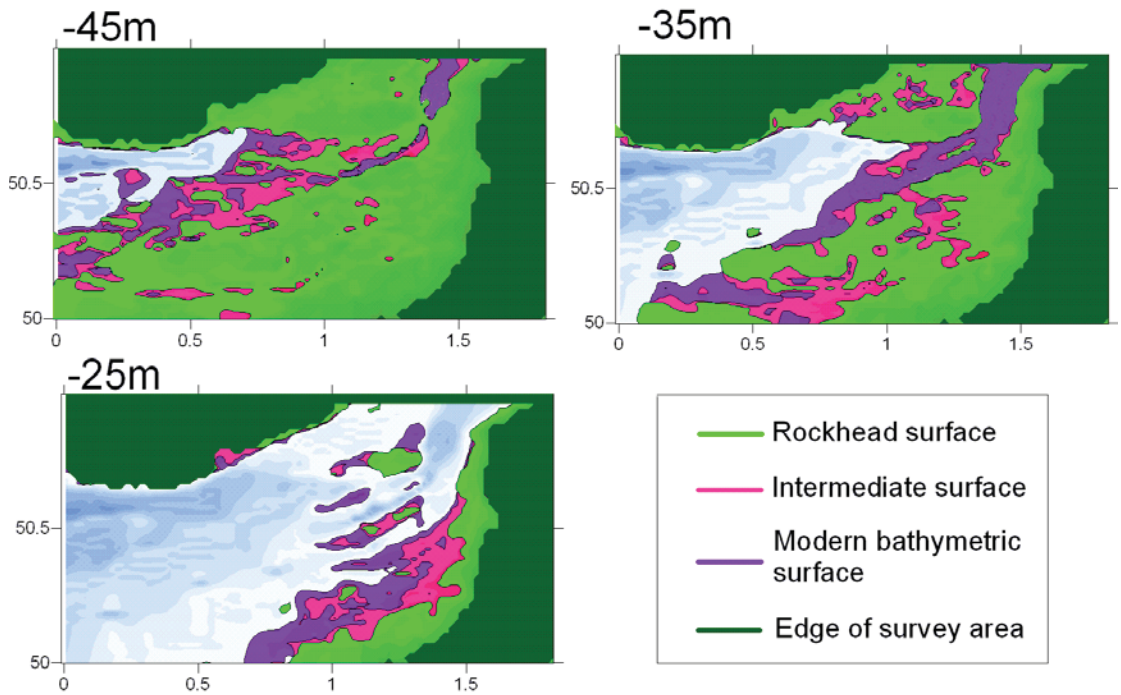


Figure 62. Overlaid shorelines produced by different bathymetric surfaces for the Dungeness region. The edge of the survey area indicates the limit of the known rockhead contours, and hence for this exercise the limits of the data and represents the present shoreline.

Depth (mbsl)	Maximum shoreline distance (rockhead to intermediate)	Maximum shoreline distance (intermediate to bathymetry)	Maximum shoreline distance (rockhead to bathymetry)	Minimum shoreline distance (rockhead to intermediate)	Minimum shoreline distance (intermediate to bathymetry)	Minimum shoreline distance (rockhead to bathymetry)
-45	19.6 km	20 km	39.6 km	0.1 km	-	0.1km
-35	37.2 km	25.9 km	63.1 km	-	-	< 0.1km
-25	13.4 km	17.9 km	31.3 km	-	-	< 0.1km

Table 13: Shoreline differences in metres between three topographic time horizons (bathymetry, bedrock and a mid-point surface) for the Dungeness region. MSL has been nominally dropped by -25 m, -35 m and – 45 m respectively.

It is clear from the figures that shoreline configurations can vary radically (e.g. Dungeness) or relatively little (e.g. Wight). Maximum and minimum shoreline position differences ranged from over 60 km to less than 100 metres. In fact, in most instances the minimum distances were so small as to be almost negligible given the scale of the area, in these instances they were assumed to be less than 100m as the scale does not really permit any measurements smaller than this. The implications of these differences are most marked in the -35 m metre comparison for the Dungeness area where we have connectivity between the current European mainland and Britain if the modern bathymetry is used whereas the island status of Britain occurs for both the intermediate and the rockhead horizons.

2.6 Discussion

The creation and analysis of these maps raises a number of relevant issues.

- Significant variability in shoreline position can result even when the difference between two curves, or the error margins within a curve appears small (e.g. several metres). The effect of this largely depends on the gradient of the surface being transgressed. On a continental scale, the variations do not appear particularly significant. However, if the actual differences in shoreline position are quantified, they can range from 1 to over 100 km. In terms of the discussion of large-scale migration or habitation patterns, or indeed creating a reasonably accurate sense of palaeo-geographic space, this magnitude of error is not particularly significant. Conversely, if these reconstructions are used for potential site prospection, these reconstructions are simply not accurate enough.

- A great deal of variability can result from the use of different curves, or data obtained from different sources as demonstrated by the comparison of GIA model, local relative sea level curves and global glacio-eustatic curves. A greater and more widespread understanding of the advantages and disadvantages of each particular type of sea level data is necessary.

- Error margins tend to increase the further back in time one goes. Note the comparison of shoreline differences in section 2.5.5.2. While the +/- 2.5 m error margin of Bard et al (1990) resulted in differences of up to 120 km, those of Rohling (up to 30m error) could create differences that were almost four times as great. Even the highest resolution pre-LGM glacio-eustatic curve currently available (Siddall, 2003) has a +/- 12m error margin. The implications of this are that for the earlier periods of prehistory, shoreline reconstruction are likely to be less accurate and rather more difficult than the later periods. Most of the error margins of this magnitude can be glossed over when discussing large scale issues, though some of the more extreme results should be taken into consideration.

- The use of lower resolution data can lead to significant differences in shoreline configuration and position. Researchers should bear this in mind and use data of appropriate resolution to their work. Ideally the data of the highest resolution should be used. If this is not available, then lower resolution data can be used, but only if its limitations and applicability are taken into account. ETOPO-2 and ETOPO-5 data, for example is far too coarse to examining anything smaller than a global or regional scale (i.e. tens of kilometres or more). It should also be remembered that that data on a single map may have been drawn at several different scales, and researchers should take this into account when using it.

- Regional relative sea level curves and GIA models provide the closest approximations to the situation in the past. However, they too suffer from their own limitations such as a limited range prior to the LGM (GIA models and regional curves) and poor large scale coverage (regional curves).

- Even in regions regarded as tectonically stable such as southern England, very slow long term trends of uplift and subsidence can create major differences in palaeo-shoreline reconstruction depending on whether these are taken into account. Although their effects are relatively small for more recent periods, they are more significant the further back one goes in the past. Accounting for them is therefore very important when considering the development of palaeo-geography on scales of hundreds of thousands of years.

- Although modern bathymetry can correlate to surfaces relating to earlier periods, in many instance there may be a significant difference (up to c. > 20 m) between them. This can lead to inaccurate representations of shoreline positions (up to 60 km difference) and past topography can be markedly mis-interpreted. The bedrock horizon represents a minimum value that could be used in reconstruction. However, modern bathymetry does not represent a maximum value as processes of erosion may have reduced its height over time.

Most of these above issues are rarely questioned when reconstructions of past landscapes are made. Maps tend to be created and presented as faithful representations despite the fact that the complexities of palaeo-geographic change have been glossed over, and assumptions have been made about the nature of the input data. The result is that reconstructions often depart significantly from each other (Section 2.1).

This is not entirely the fault of their authors. In many instances there is often no alternative but to use certain types of data, such as glacio-eustatic curves, in the absence of other categories of evidence. This is not to say that all previous work should be discarded, and regarded as a totally unrepresentative image of the past. Rather, by developing some measure of recognition of the error margins involved in existing representations their use can still be a worthwhile exercise.

Future work should also consider the nature of the input data before simply applying sea level data uncritically. For instance, GIA models have become increasingly popular in recent years (e.g. Coles, 1998; Flemming, 2002), but rather than simply adopting them wholesale, it is essential that the issues surrounding them are known, such as, whether they apply in all instances (e.g. McCabe, 1997), or whether the ANU models are more accurate than the University of Toronto's (Lambeck et al, 2003) or vice versa. In addition, these reconstructions are only available for regions that have been studied by the GIA modellers. Constructing such models independently tends to be beyond the resources of most archaeological organisations.

In any case, improving existing reconstructions of submerged landscapes is essential if archaeologists wish to take their study to the next level - that of actually locating and excavating evidence from the sea bed. Given the expensive and time consuming nature of underwater survey, an accurate picture of the past landscape and possibly the location of probable sites will be necessary, to make the prospect of systematic underwater work a reality. The study of these areas on a smaller scale (i.e. not simply looking at broad processes of colonization or migration) will also require a far more detailed knowledge of the landscape than is currently available. This may in turn require detailed information on local sea level changes and accompanying landscape evolution. This will be discussed further in Section 4, while the above proposals will be reviewed at the end of the entire document, and modified if necessary on the basis of the results of the following Themes.