Chapter 2 Pleistocene environmental change in Britain

2.1 Introduction

This chapter introduces the terminology used throughout the thesis. An overview of Pleistocene environmental change and chronostratigraphy is presented and current understanding of the British Pleistocene stratigraphic framework is assessed; this sets out the thesis context and shapes the methods and interpretation applied in this thesis. The geomorphological and sedimentary record of the British Pleistocene is summarised and assessed, with a focus on Pleistocene features of the coastal lowlands of the Bristol Channel/Severn Estuary region (the Gower Peninsula Gwent Levels, Somerset Levels and Moors and coastal lowlands) in order to provide a regional context for the Pleistocene evolution of the Gordano Valley.

2.2 Terminology

Throughout this thesis the shorthand form recommended by the IUPAC-IUGS Task Group (2006) is used for years before present: ka – thousand years; Ma – million years. The age of events is defined using four different notations:

1. Radiometric dates are quoted in conventional (uncalibrated) radiocarbon years before present (BP) with the present taken as 1950 calendar years AD (e.g. ^{14}C ka BP).

2. Calibrated radiocarbon years BP, whereby radiocarbon years have been converted to a calibrated calendar timescale (e.g. Cal ka BP).

3. Where reference is made to Greenland ice-core records, direct dating of horizons within the North Greenland Ice Core Project (NGRIP) ice-core records derived by counting annual ice layers down from the surface, ages are expressed in ka GICC05 b2k (Greenland Ice Core Chronology 2005 ice-core years before 2000 AD).

4. Calendar ages based on other methods are expressed as ka.

The term proxy record, or proxy evidence, is used to refer to inferential evidence that is based on an indirect measure of former climates or environments (Lowe & Walker 1997, Walker 2005).

The Last Glacial Maximum (LGM) refers to the maximum extent of the British-Irish Ice Sheet (BIIS) about 27-21 ka (Chiverrell & Thomas 2010, Clarke *et al*. 2010), determined from stratigraphy and radiometric dating.

2.2.1 The Quaternary Period

The definition of Quaternary terms used in this thesis is set out in Table 2.1. The Quaternary Period comprises two epochs within the Cenozoic Era; the Holocene Epoch, comprising the present warm interval, and the Pleistocene Epoch (Gibbard *et al*. 2010). The start of the Quaternary/Pleistocene is formalised at 2.588 million years from a Global Stratotype Section Point (GSSP) at Monte San Nicola in southern Italy (Gibbard *et al*. 2010). The Pleistocene Epoch is further divided into glacial (cold) and interglacial (temperate) stages which are subdivided into stadial (cold) and interstadial (temperate) sub-stages (Lowe & Walker 1997).

Table 2.1: Definitions of Quaternary terms used in this thesis (sources: Gibbard & van Kolfschoten 2004, Walker *et al*. 2009, Chiverrell & Thomas 2010, Clark *et al*. 2010, Gibbard *et al*. 2010)

Quaternary Period	The most recent 2.588 Ma of Earth history
Pleistocene Epoch:	The geological time period from 2.588 Ma to the start of the Holocene Epoch 11.703 ka GICC05 b2k
Early Pleistocene	2.588-0.78 Ma
Middle Pleistocene	0.78-0.126 Ma
Late Pleistocene	0.126-0.01 Ma
Holocene Epoch	The present post-glacial epoch, beginning 11.703 ka GICC05 b2k and extending to the present day
Glacial stage	A cold phase during which there were major expansions of ice sheets and glaciers
Interglacial stage	A temperate interval during which temperatures were as high, or higher, than those experienced today
Stadial	A short cold sub-stage during a temperate stage when local ice advances occur
Interstadial	A relatively short period of climatic improvement during a cold stage, when temperatures may not have reached those of the present day
LGM	The maximum extent of glaciation in Britain, about 27-21 ka

The Holocene Epoch is generally regarded as having begun 10000 radiocarbon years before 1950 AD or 11703 ka GICC05 b2k ago (Lowe & Walker 1997, Walker *et al*.

2009, Cohen & Gibbard 2010, Gibbard *et al*. 2010). Its boundary has been defined as a Global Stratotype Section and Point (GSSP) in the North-GRIP ice core of the Greenland Ice-Core Project (Walker *et al*. 2009).

2.3 Overview of Pleistocene environmental change

The dominant characteristic of the Pleistocene is the global expression of major climatic and environmental oscillations, which in the mid-latitudes of the Northern Hemisphere takes the form of repeated extensive glaciations interspersed with temperate stages (Bowen & Gibbard 2007). These repeated changes have resulted in a highly complex record of sediments, relict landforms, polygenetic landscapes and biological remains (Bowen 1978, Barry 1997, Bowen & Gibbard 2007) from which it is possible to interpret and reconstruct the environmental conditions pertaining at particular intervals of the Pleistocene (Lowe & Walker 1997, Gibbard *et al.* 2005). This paradigm of Pleistocene climatic variations forms an integrating concept for understanding how the Earth evolved under contemporary conditions (Gibbard *et al.* 2005, Bowen & Gibbard 2007) and the presence of alternating warm and cold episodes, with periodic changes in global ice volume coupled with sea-level fluctuation, is fundamental to this research and is returned to later in this thesis.

Pleistocene sediment sequences have long been divided on the basis of inferred climatic changes (Bowen 1978, Gibbard & West 2000). Although mainly applied to the Middle and Late Pleistocene, climate change-based classification is central to Pleistocene subdivision (Gibbard & van Kolfschoten 2004). The terms ‘interglacial’ and ‘interstadial’ are customarily used to define the characteristics of non-glacial climate conditions and ‘glacial’ and ‘stadial’ to define cold climate conditions, although the periods between interglacial events are known to have been characterised by cold rather than glacial climates (Gibbard & van Kolfschoten 2004); the terms ‘cold stage’ and ‘temperate stage’ are now more frequently adopted (Gibbard & van Kolfschoten 2004) although the terms are used interchangeably in this thesis.

2.4 Chronostratigraphy of the Pleistocene Epoch

Chronostratigraphy is the classification of the stratigraphic record in units that correspond to intervals of geological time. It is inferential, being based indirectly on the characteristics of the sediment record (Lowe & Walker 1997). Geochronology is the division of the stratigraphic record on the basis of time and represents the time interval itself (Lowe & Walker 1997). Stratigraphic succession, in which vertical and lateral relationships are used to reconstruct a depositional sequence, rests on the assumption of superposition (that upper strata in sedimentary sequences postdate those which they overlie unless the sequence has been altered by intense folding, faulting or other disturbance, for example, reworking of sediment by marine and/or fluvial activity) and correlation (Lowe & Walker 1997, Gibbard & West 2000). All are integral to the reconstruction of former environments (Lowe & Walker 1997, Evans & Benn 2004).

2.4.1 Marine oxygen isotope sequences

Marine oxygen isotope chronostratigraphy relies on the presumption that sedimentation has been continuous throughout the Pleistocene (Bridgland 1994) and that sea level is lowered as continental ice builds up in response to global cooling. Ice is depleted in $\delta^{18}O$ (versus $\delta^{16}O$) relative to ocean water and this is reflected in the oxygen isotope composition of the carbonate shells of benthic foraminifera (Schreve & Candy 2010). Events differentiated in isotopic sequences are termed Marine Isotope Stages (MIS) (Shackleton 2006) and these provide a universal means of subdividing the Quaternary (McMillan *et al.* 2005). Stages are numbered from the present day (MIS 1) backwards in time; even numbers are assigned to cold stages, odd numbers to temperate stages (Figure 2.1). Sub-stages are indicated by either lower case letters or a decimal system; for example MIS 5 is divided into warm sub-stages 5a, 5c and 5e or 5.1, 5.3 and 5.5 and cold sub-stages 5b and 5d or 5.2 and 5.4 (Gibbard & van Kolfschoten 2004). Isotope studies from ocean bottom sediments have indicated over 100 Quaternary isotopic stages (Gibbard & van Kolfschoten 2004, Walker 2005, Cohen & Gibbard 2010).

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Figure 2.1: Age of Marine Isotope Stages. Even numbers represent cold stages, odd numbers represent temperate stages (source: Cohen & Gibbard 2010)

Although the marine isotope record provides a global stratigraphic framework for establishing the timing of glacial and interglacial episodes to which continental stratigraphies are increasingly tied, the marine oxygen isotope curve is not a direct indicator of temperature oscillation (Ehlers & Gibbard 2003, Schreve & Candy 2010). For temperature, reliance is placed on ice-core sequences which have revolutionised the way that the patterns and rates of global climate change are understood (Gibbard & van Kolfschoten 2004).

2.4.2 Antarctic and Greenland ice-core records

An independent record of later Pleistocene climatic conditions has been derived from ice cores from deuterium (δD) and oxygen (δO) isotopes in ice cores, as an indication of surface temperature change in Greenland and Antarctic ice sheets (Dansgaard *et al.* 1993, EPICA community members 2004). Variation in temperature dependent oxygen isotope ratios ($\delta^{18}O$/ $\delta^{16}O$) is determined from atmospheric gas bubbles in the ice, whilst deuterium excess ($d = \delta D - 8\ \delta^{18}O$) is determined from the ice itself, (Johnsen *et al.* 1992, Petit *et al.* 1997), and hence gives a direct indication of surface palaeotemperature (EPICA community members 2004). The Greenland and Antarctic ice-core records complement that of the marine sediment record whilst providing finer resolution of climate events, although beyond 800 ka reliance has to be placed on the marine record (Gibbard & West 2000, Hoek & Bohncke 2001, Ehlers & Gibbard 2003). Resolution may be annual or better for the Greenland ice-cores (Steffensen *et al.* 2008, Orombelli *et al.* 2010) compared to roughly 1000 years for foraminiferal $\delta^{18}O$ data from marine sediment cores (Berger 2008, Svensson *et al.* 2008).

The European Project for Ice Coring in Antarctica (EPICA) Dome C (EDC) core spans a period of 800 ka (Jouzel *et al.* 2007, Masson-Delmotte *et al.* 2010), whilst those of the Greenland Ice Core Project (GRIP), North Greenland Ice Core Project (NGRIP) and Greenland Ice Sheet Project2 (GISP2) provide a sequence that extends back to the Last Interglacial (MIS 5e; NGRIP members 2004) and have been used to provide a chronology covering the past 60 ka based on annual layer counting (Svensson *et al*. 2008). Figure 2.2 shows Antarctic temperature change over the past 800 ka, the close agreement of ice core and marine core records and comparison with the NGRIP ice core record, including Dansgaard-Oeschger events (section 2.4.3) (Jouzel *et al*. 2007).

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Figure 2.2: Antarctic temperature change based on changes in δD over approximately the past 800 000 years (red curve) showing close agreement with MIS (blue figures) and comparison with NGRIP ice-core record (green curve & figures), including Dansgaard-Oeschger Events (Jouzel *et al*. 2007, Figure 2, p 794)

These sequences have revolutionised the understanding of rates and patterns of global climate change (Ehlers & Gibbard 2003). Annual time-resolution from the Greenland δO records form the basis of the GICC05 Greenland ice core chronology, the most precise available for the North Atlantic region (Hoek & Bohncke 2001, Dreyfus *et al.* 2007, Hoek *et al.* 2008).

2.4.3 Dansgaard-Oeschger and Heinrich events

From the study of Greenland ice-cores evidence has been found, within the longer-term glacial-interglacial cycle, of apparently cyclical high-frequency, abrupt, warm-cold climate oscillations, known as Dansgaard-Oeschger (D/O) events (Stocker & Schilt 2008, Svensson *et al.* 2008, Siddall *et al.* 2010). These appear in groups, sometimes referred to as Bond cycles, with a periodicity of around 1500 years. D/O events start with an abrupt warming of Greenland by 5-10°C over a few decades or less, followed by a gradual cooling over a period of up to a thousand years and often end with an abrupt final reduction of temperature back to cold (stadial) conditions. Seventeen D/O events, correlated with Greenland Interstadials (GI) dated back to 60 ka GICC05 b2k, are recorded in the Greenland ice-cores (Svensson *et al.* 2008), such that MIS 3 is a period of millennial-scale variability (Bond *et al.* 1993, 1997, Ganopolski & Rahmstorf 2001, Bohncke *et al.* 2008, Siddall *et al.* 2010). D/O events are centred on the North Atlantic, especially regions with strong atmospheric response to changes, although corresponding, but smaller changes are also recorded in the Southern Hemisphere (Ganopolski & Rahmstorf 2001, Jouzel *et al.* 2007, Siddall *et al.* 2010). The same abrupt short-term climatic fluctuations are also displayed in North Atlantic marine sediments (Bond *et al.* 1993, 1997, Bohncke *et al.* 2008) and the continental record (Siddall *et al.* 2010). D/O groups culminate in a cold Heinrich event, during which massive ice-rafted debris is deposited in the North Atlantic (Siddall *et al.* 2010).

Heinrich events principally involve surging of the North American Laurentide Ice Sheet, resulting in iceberg discharge and large inputs of fresh water into the North Atlantic sufficient to slow-down or interrupt thermohaline circulation, causing a cold snap (Ganopolski & Rahmstorf 2001, Lambeck *et al.* 2002b). They occur in the cold, stadial phase of some D/O cycles, and have an irregular spacing of several thousand years (Ganopolski & Rahmstorf 2001). Terrestrial geomorphological evidence (moraines and subaqueous glacial outwash) from around the northern margins of the Irish Sea Basin indicating calving ice-sheet margins has been correlated with the Heinrich 1 event (McCabe & Clark 1998, McCabe *et al.* 1998, Bowen *et al.* 2002, McCabe *et al.* 2005).

2.4.4 Correlation of terrestrial, marine and ice-core records

The chronology of major episodes of global change over the Quaternary is provided by inflections in the oxygen isotope curve of the marine isotope record, tuned to the Milankovitch orbital forcing cycles which dictate the timing of major glacial and interglacial stages as well as the occurrence of shorter sub-stage climate oscillations (Walker 2005, Candy & Schreve 2007). However, there are a number of limitations affecting the interpretation of the marine oxygen isotope record. Variable, generally low, sedimentation rates in the deep oceans mean that stratigraphic resolution is low. This may be exacerbated by sediment mixing on the sea floor as a result of bioturbation by benthic organisms, turbidity currents or removal of sediments through scouring by ocean bottom currents resulting in depositional hiatuses which further complicate the isotopic signal. In deep oceans, the carbonate benthic foraminifera tests that carry the isotopic record can dissolve as they settle through the water column. The critical depth (carbonate compensation depth, CCD) is between 3 and 5 km; oxygen isotope studies therefore tend to be confined to water that does not exceed the CCD (Lowe & Walker 1997, Walker 2005). Other sources of error arise through incorrect correlation of stages or stage boundaries between different isotopic profiles, or because of the complexities of certain isotopic stages (e.g. MIS 5 with five sub-stages). Consistent interpretation of different isotopic profiles is not easily achieved, especially where stratigraphic resolution is poor (Walker 2005).

Nonetheless, stable isotope records from undisturbed marine sediment cores and continuous ice cores recovered from sites in Greenland and Antarctica now provide a temporal framework for Pleistocene environmental change which can be read as a record of glacial and interglacial oscillations and provides an independent scheme which terrestrial sequences are increasingly compared and correlated against (Johnsen *et al.* 1992, Lowe & Walker 1997, Campbell *et al.* 1998, NGRIP members 2004, EPICA community members 2004). This has lead to terrestrially-based terms being supplanted by the chronostratigraphical subdivision of marine-core sequences (Schreve & Thomas 2001), which in turn has revolutionised palaeoenvironmental interpretations so that increasingly detailed palaeoenvironmental interpretations can now be attempted, leading to improved understanding of terrestrial response to climatic variations (Hoek *et al.* 2008, Schreve & Candy 2010).

By comparison the continental evidence is incomplete and fragmentary (Gibbard & van Kolfschoten 2004), providing records with insufficient evidence to match the numerous oscillations that have occurred throughout Pleistocene time demonstrated by marine oxygen isotope stages (Gibbard & West 2000, Ehlers & Gibbard 2003). Whilst climatic changes of the Late Pleistocene are relatively well known, events prior to this can only be broadly reconstructed, impeding correlation of terrestrial stratigraphy with marine oxygen isotope stratigraphy (Ehlers & Gibbard 2003). However, although difficult, correlation of fragmentary, local, high-resolution terrestrial records with the continuous, yet lower resolution marine isotope record is necessary because terrestrial sequences record local and regional climatic conditions whilst the marine record indicates the global situation (Gibbard & van Kolfschoten 2004).

A consequence of the development of high-resolution stratigraphy is the need for precise recognition and timing of correlative depositional events on land. Previously it was generally assumed that terrestrial boundaries identified using a variety of proxies were precisely coeval with those seen in ocean sediments. However, the recognition of climatic events from sediments is not straightforward. There are ambiguities, and characteristic sedimentary structures and textures, including periglacial structures, soil development and fossil assemblages must be relied on if the problems of discontinuous sedimentation, sedimentary rate changes and lateral variability are to be overcome and sequences unravelled (Ehlers & Gibbard 2003, Gibbard & van Kolfschoten 2004). Additionally, these changes may be time-transgressive, or indicate a higher degree of climate sensitivity than those of marine sediments. Changes in ocean currents, sea level, wind directions and patterns and tectonics may further complicate local responses to global temperature change reflected in coastal regions (Tzedakis *et al*. 1997, Sanchez-Goñi *et al*. 1999, Gibbard & van Kolfschoten 2004).

The chronostratigraphy and complexity of the isotopic record during the Pleistocene Epoch has been a subject of intense discussion and the desire to correlate changes in the sediment record across local areas, regions and globally has resulted in much debate. The search for terrestrial evidence of multiple climatic oscillations coupled with the move towards correlation with marine isotope stages and, in the North Atlantic region, with Greenland ice core stages, has resulted in increasingly sophisticated high-resolution stratigraphy, particularly over the last glacial-interglacial transition, such as that achieved at Llanilid, South Wales by Walker *et al*. (2003). However, limitations in the precision and accuracy of routinely used dating and correlation methods have proved a challenge to precise correlation (Hoek *et al*. 2008), and inhibited the temporal resolution of individual climatic episodes (Lowe *et al*. 2008). As the next section shows, the usefulness of larger-scale divisions of the Pleistocene has diminished (Gibbard 2003) as recognition of the detail represented in ocean sediment and ice-core sequences challenged stratigraphers to identify and correlate ever smaller-scale events and address the implications of high-resolution sequences.

2.5 British Pleistocene stratigraphic framework

Past attempts to subdivide the British stratigraphic record into a coherent scheme have proved difficult because the terrestrial stratigraphic record is fragmentary (Ballantyne & Harris 1994, Benn & Evans 1998, Lewis 2005). Consequently interpretation of this record, with discontinuities and unconformities, spatial variations in sediment type and thickness, limited lateral continuation of deposits and poorly fossiliferous deposits with little dateable material seen in sections and boreholes, is difficult. Additionally, the time-transgressive nature of Pleistocene stratigraphic boundaries has made inter-site correlation difficult (Bowen 1978, Jones & Keen 1993, Walker *et al*.1999, Evans 2005, McMillan *et al*. 2005). It has not yet been possible to construct a complete stratigraphy (Lee *et al*. 2006), and there are often contradictory interpretations of the same evidence, discussed later in this section.

2.5.1 The 'Named Stages' classification model

The main problems have revolved around the use of the 'Named Stages' classification model (summarised in Table 2.2) which recognises a total of sixteen major climatic oscillations or stages in the British Quaternary. This temporal resolution is wholly inadequate at representing Pleistocene climatic complexity when compared to the marine oxygen isotope record. Furthermore, attempts to rigidly adhere to the 'Named Stages' model often resulted in pigeon-holing new discoveries, perpetuating an illusion of precision and confirming the original model (Bowen 1978). A further problem lies in gaps in the stratigraphic record in comparison with the marine record (see Table 2.2). British Early/Middle Pleistocene cold stages lack precise dates (Jones & Keen 1993), and there are also gaps in the record of the most recent, Devensian (MIS 5d-2), cold stage; in Britain only four Devensian (MIS 5d-2) interstadials have been identified (Chelford, Brimpton, Upton Warren and Windermere) whereas Greenland ice core records indicate 17 interstadials during the same period (Ehlers *et al*. 1991, Svensson *et al*. 2008).

A major problem with this classification is the lack of suitable dating techniques which resulted in a stratigraphic classification model based on inferences from the identification of temperate stages defined by pollen assemblage zones, with glacial stages fitted into this framework based on the absence of fossil pollen (West 1963, Bowen 1978, Ehlers *et al*. 1991, Bowen 1999b, McMillan *et al*. 2005, Lee *et al*. 2006, Gibbard *et al*. 2008, Rose 2009). It is now recognised that pollen assemblages may be replicated during different chronostratigraphic stages; conversely correlation of temperate deposits of the same age may be precluded because they display widely different characteristics as a result of geographical and geological differences, thus undermining the value of pollen assemblages as correlative tools (Bowen 1978, Bowen 1999b, Campbell & Bowen 1989, Ehlers *et al*. 1991, Rose *et al*. 2001, Thomas 2001, Lee *et al*. 2006, Roe *et al*. 2009). With the discovery of additional interglacial episodes from marine core evidence, and improved stratographic resolution indicating warm and cool sub-stages, the pollen-based system of dating and subdivision was recognised to be inadequate (Bowen *et al*. 1989, Bowen 1991). It is now accepted that individual 'Named Stages' are an approximation of the stratigraphic sequence and several 'Named Stages' identified in the British sedimentary record contain a number of separate cold and temperate stages which can now be correlated with individual marine isotope stages (Lowe & Walker 1997, Candy & Schreve 2007, Schreve & Candy 2010).

However, historical use of the 'Named Stages' has led to continuing efforts to force multiple MIS substages into 'Named Stages'. Consequently, correlation between sites is problematical and there has been little nationwide or even regional integration of evidence from temperate deposits into a wider stratigraphic context (Rose 2009). Nevertheless, 'Named Stages' have proved useful in developing a stratigraphy for the British Isles and providing an understanding of landscape development during the Pleistocene (Rose 2009). Therefore, wherever possible, in order to reflect earlier practice, to achieve finer temporal resolution/comparison and to compare the stratigraphy of the Gordano Valley with that of other UK sites, British Stage names will be used in conjunction with MIS and both are shown in Table 2.2.

Table 2.2: Quaternary stratigraphy of the British Isles (sources: Lowe & Walker 1997, McMillan *et al*. 2005, Bradwell *et al*. 2008, Walker *et al*. 2009, Cohen & Gibbard 2010)

Timescale Ma	Series	Subseries	British Stage			MIS
0.0117	HOLOCENE		HOLOCENE			1
			DEVENSIAN			
			Lateglacial stratigraphic subdivision in the British Isles			
			Age ^{14}C ka BP	Subdivision		
			11-10	Loch Lomond Stadial (Younger Dryas)		
		LATE	13-11	Lateglacial or Windermere Interstadial (Bølling/Allerød)		
			Devensian stratigraphic subdivision in the British Isles			
			Age ka	Subdivision	Notes	
			31-11.7	Late Devensian	Dimlington Stadial	2
			58-31	Middle Devensian	Upton Warren Interstadial Complex	3
0.116	PLEISTOCENE		116-58	Early Devensian	Chelford Interstadial	4 5d-5a
0.126			IPSWICHIAN			5e
0.300			WOLSTONIAN			10-6
0.430			HOXNIAN			11
0.480		MIDDLE	ANGLIAN			12
0.630-0.780			CROMERIAN			13
			BEESTONIAN			21
1.806			PASTONIAN			63
			PRE-PASTONIAN/BAVENTIAN			
		EARLY	BRAMERTONIAN/ANTIAN			
			THURNIAN			
			LUDHAMIAN			103
2.588			PRE-LUDHAMIAN			104

Devensian (MIS 5d-2) glaciation is traditionally divided into Early- (116-58 ka), Middle- (58–31 ka) and Late- (31–11.7 ka) (Table 2.2 inset, Lowe & Walker 1997, Bradwell *et al*. 2008). Three British interstadial sites may date to the Early Devensian, although the date of these interstadials is uncertain as they lie outside the range of the radiocarbon dating technique (Coope *et al*. 1997). Early radiocarbon dating of interstadial deposits at Chelford, Cheshire, suggested an age of *c*. 60 ka (Simpson & West 1958, Bowen 1999a), but TL dating has indicated an age of *c*. 100 ka, suggesting correlation with MIS 5c is more appropriate (Worsley 2005). Other probable Early Devensian interstadials are recorded at Cassington, Oxford, correlated with MIS 5a on the basis of biostratigraphical data (Maddy *et al*. 1998), and at Brimpton, Berkshire, where an Early Devensian age is inferred from the stratigraphy (Bryant *et al*. 1983, Worsley & Collins 1995).

Interstadial deposits of the 'Upton Warren Interstadial Complex' represent the only MIS 3 interstadial in the British record, and are recognised at many sites (Coope *et al*. 1997, van Huissteden *et al*. 2001, Maddy & Lewis 2005). However, there is uncertainty surrounding the precise date of this interstadial; radiocarbon dates cover the period from 40-25 ^{14}C ka BP (Coope & Brophy 1972, Coope *et al*. 1997, Worsley 2005), whilst Lewin & Gibbard (2010) have recently placed a date of *c*. 43 ka BP on the Upton Warren Interstadial climatic optimum. Van Huissteden *et al*. (2001) have suggested that since identification is based on radiocarbon dating and occurs close to the age range limit of the technique, the various dates represent different warm intervals within the stage. Alternatively, the radiocarbon dates may represent several different age results for the same event. Attempts to estimate its age by other means led Bowen *et al*. (1989) to make a case for the correlation of Upton Warren deposits with MIS 5a on the basis of amino acid racemization (AAR) geochronology, although this correlation is not generally accepted.

The end of the Devensian (13-10 ^{14}C ka BP) is generally referred to as the Late Devensian Lateglacial. In Britain, a twofold division is usual (Table 2.2 inset); the Windermere Interstadial (13-11 ^{14}C ka BP) is followed by cold conditions during the Loch Lomond Stadial (11-10 ka BP) (Ballantyne & Harris 1994). In northwest Europe four major sub-divisions of the pollen-based Lateglacial biostratigraphic record are recognised: the Bølling (13-12 ^{14}C ka BP) and Allerød (11.8-11 ^{14}C ka BP) Interstadials, separated by two short cold phases, the Older Dryas (12-11.8 ^{14}C ka BP) and Younger Dryas (11-10 ^{14}C ka BP) Stadials (Lowe & Walker 1997). However, the validity of this stratigraphical model has been called into question due to uncertainties in radiocarbon dating (Lowe *et al*. 1999). Furthermore, proxies other than pollen, for example coleoptera, often suggest a different pattern of climate change (Coope & Brophy 1972, Walker 1995, Lowe *et al*. 1999). Consequently, although use of the northwest European scheme continues, it is gradually falling from favour (Lowe *et al*. 1999).

2.5.2 Correlation of sediments with MIS

A recurring theme of Pleistocene stratigraphy is the lack of absolute dating control on the age of sediments (Bowen 1991, O'Cofaigh & Evans 2007), either through the absence of dateable deposits or the absence of a technology that will allow deposits to be dated. The development of dating techniques for Pleistocene events has therefore been crucial for the reconstruction of past climatic changes (West 1985). Until recently, radiocarbon dating was the principle means of dating available and was widely used for dating Late Pleistocene events. However, the technique is severely limited; it relies on the availability of suitable organic material, is limited to approximately 60,000 years and determination of precise dates is difficult (Walker *et al*. 1999, Lian & Roberts 2006).

Advances in dating technologies such as accelerator mass spectrometry (AMS), luminescence dating, both optically stimulated (OSL) and thermal (TL), electron spin-resonance (ESR), uranium-series (U-series), cosmogenic radionuclides (e.g. ^{36}Cl) and thermal ionisation mass spectrometry (TIMS) have allowed a wider range of material to be dated and AAR geochronology has further enabled dating through correlation with dated sediments. These advances are proving crucial for defining a framework for understanding Pleistocene environments, enabling extension of dateable deposits to the Middle Pleistocene, deposits of named stages to be firmly tied to marine isotope sequences and recognition of a highly diverse climatic structure (Schreve & Candy 2010). As a result, many sites are now being reassessed and reinterpreted, revealing a more complex terrestrial record of Pleistocene environmental change than was previously accepted (Schreve 2001a, Stuart & Lister 2001, Richards 2005). For example: the beach at Morston, Norfolk, long held to be Ipswichian (MIS 5e), was recently reassigned to the MIS 7-6 transition based on OSL dating (Hoare *et al*. 2009), whilst reinterpretation of the fossilferous deposits at West Runton, Norfolk, indicates that up to five distinct temperate stages were previously conflated into a single Cromerian (MIS 13) stage (Stuart & Lister 2010, Preece 2010), with the Cromerian stratotype deposits correlated by AAR with MIS 15 or 17 (Penkman *et al.* 2010). Mammalian and molluscan faunas from Pakefield and Corton, Suffolk, have also provided evidence for additional early Middle Pleistocene temperate stages, although there is conflict between mammalian evidence and the AAR geochronology of molluscs (Stuart & Lister 2001, Preece 2001). There is now increasing acceptance for three interglacials between the Anglian (MIS 12) and Ipswichian (MIS 5e) stages (Ballantyne & Harris 1994, Campbell *et al.* 1998, Schreve 2001a, Pawley *et al.* 2008) and recognition in the terrestrial record of sub-stages (Schreve 2001a and b); AAR geochronology and biostratigraphical evidence from Hoxne, Suffolk, recognises two post-Anglian (post-MIS 12) temperate sub-stages correlated with MIS 11c and 11a, with an intervening cold sub-stage correlated with MIS 11b (Ashton *et al.* 2008); U-series dating of tufa from Marsworth, Buckinghamshire, provides evidence for temperate sub-stages MIS 7e and 7c (Candy & Schreve 2007). Additionally, cosmogenic nuclide surface exposure dating, OSL and AMS radiocarbon dating and AAR correlation of continental margin marine cores has advanced the understanding of the BIIS at the LGM (Bowen *et al.* 2002, Evans & O'Cofaigh 2003, McCabe *et al.* 2005, O'Cofaigh & Evans 2007, Bateman *et al.* 2011).

Despite these developments there often remains uncertainty over the correlation of deposits to a particular stage or sub-stage. The timing of Early Pleistocene glacial activity can only be inferred from indirect evidence because glacial sequences are generally devoid of material suitable for dating using radiometric dating methods (Pawley *et al.* 2008), resulting in alternative interpretations for these early glaciations (Bowen *et al.* 1986). For example, in an interpretation of depositional sequences in East Anglia which challenged traditional biostratigraphically-based interpretations, Hamblin *et al.* (2005) and Lee *et al.* (2006) recognised evidence for three separate cold stages from which a MIS 16 glacial advance across most of Britain was inferred. However, this interpretation is now generally regarded as disproven (Pawley *et al.* 2008, Preece *et al.* 2009, Westaway 2010a); re-examination of the East Anglian sequences confirmed a MIS 12 age rather than the MIS 16 age, based on aminostratigraphy and faunal assemblages and the revision of stratigraphic relationships between glacigenic deposits assigned to this stage and deposits of the pre-Anglian Ingham river.

There is also ongoing debate surrounding differentiation of MIS 11 and MIS 9 deposits (Roe *et al.* 2009). Palynological evidence from Middle Pleistocene lake sequences in the Midlands has suggested that two interglacials have been conflated under the term

‘Hoxnian’, with sites formerly attributed to the Hoxnian interglacial including representatives correlated with MIS 11 and 9, the quality of the palynological signal being insufficient to differentiate between the two (Keen *et al*. 1997, Thomas 2001). Non-pollen lines of evidence have indicated that some ‘Hoxnian’ sites may be of different ages; for example, Cudmore Grove and Clacton have similar ‘Hoxnian’ pollen signals but their faunal records and sea-level history differ (Schreve 2001a and b, Roe *et al*. 2009).

Additionally, deposits formerly attributed to the Ipswichian (MIS 5e) stage may represent MIS 7 (Jones & Keen 1993). In the lower Thames region deposits at Aveley, Ilford and Crayford had been previously considered on palaeobotanical grounds to be Ipswichian (MIS 5e), whereas there is now a consensus, on the basis of mammalian and AAR geochronology, that most of these deposits belong to MIS 7 (Jones & Keen 1993, Bridgland 1994, Lowe & Walker 1997, Preece 1999). Conversely, mammalian fauna found in gravels at Gloucester are of MIS 5e age, contradicting the earlier interpretation of a MIS 7-6 age (Schreve 2009).

2.6 Extent of British Pleistocene deposits

Section 2.5 has discussed the fragmentary nature of Pleistocene sediments and the difficulties of assigning them to a particular cold or temperate stage. In this section, the known extent of Pleistocene sediments is discussed, and the implications of this are assessed.

2.6.1 Extent of Pleistocene glacial deposits

The distribution of glacigenic deposits is directly influenced by the extent and behaviour of Pleistocene ice sheets, particularly the most extensive (Anglian, MIS 12) and the most recent (Devensian, MIS 5d-2) (McMillan *et al*. 2005). However, glacial sequences in Britain have been the subject of considerable reinterpretation, leading to revisions of reconstructed Pleistocene ice limits (Ballantyne & Harris 1994, Campbell *et al.* 1998, Pawley *et al*. 2008).

No true Early Pleistocene glacial deposits are known in Britain, although Early Pleistocene cool climates have been inferred from fossil evidence in the shallow marine and intertidal sands of the East Anglian Crag deposits (West 1963, Jones & Keen 1993, Funnell

1996). This evidence appears to be confined to the Crag Basin of East Anglia, although gravels from Buchan, northeast Scotland, have also been interpreted as Early Pleistocene glacial deposits (Jones & Keen 1993).

The Kesgrave Formation and Bytham gravels of East Anglia have been used to infer pre-Anglian (pre-MIS 12) Middle Pleistocene glaciation. These contain erratic clasts from North Wales and the Midlands which are overlain by glacial deposits of Anglian (MIS 12) age (Bowen *et al*. 1986, Bowen 1991, Clark *et al*. 2004, Rose 2009, Parfitt *et al*. 2010, Rose 2010). The oldest glaciation for which there may be direct evidence (MIS 16) is recorded at Happisburgh, Norfolk (Clark *et al*. 2003, Lee *et al*. 2004, 2008, Hamlin *et al*. 2005, Rose 2009, Parfitt *et al*. 2010, Rose 2010) and possibly in Somerset (Gilbertson & Hawkins 1978b, Bowen 1991). Debate surrounding the age of the latter glaciation is discussed in more detail in section 2.8.

Anglian (MIS 12) glacigenic deposits are mainly found south of the Devensian (MIS 2) ice limit (Bowen 1973a and b, Bowen *et al*. 2002) and north of a line approximately east-west from the River Thames to the Bristol Channel (Figure 2.3), but also occur elsewhere in Britain although evidence in Scotland is tenuous (Bowen *et al*. 1986, Jones & Keen 1993, Gibbard *et al*. 2009). It has been inferred that during this glaciation, the course of the ancestral Thames was diverted south, the Bytham river in eastern England was obliterated and the course of the River Severn in western England was reorganised (Bridgland 1994, Maddy *et al*. 1995, Rose 2009).

There is controversy over the extent of a Wolstonian (MIS 10-6) glaciation, particularly in eastern England, where deposits formerly attributed to a Wolstonian (MIS 10-6) glaciation are now thought to represent the earlier Anglian (MIS 12) glaciation (Gibbard *et al*. 2009). The Wolstonian glaciation may be poorly represented in the Pleistocene record because it appears to have a similar extent to the later Devensian (MIS 2) glaciation (Clark *et al*. 2004). Till found in Leicestershire and Norfolk has been assigned to MIS 10 (Rose 2009), although the validity of this has been questioned by Pawley *et al*. (2008), who favour a MIS 12 age. However, till from a glaciation of this interval *is* found in northeast England (Bowen *et al*. 1986), whilst in eastern England gravels of the River Trent terrace archive provide evidence for glaciation within the MIS 10-6 period, probably MIS 8 (Westaway 2010a, White *et al*. 2010, Straw 2011). Till, and sands and gravels interpreted as marginal glacial provide evidence for a cold stage immediately preceding the Ipswichian (MIS 5e) stage, particularly in the Midlands, East Yorkshire, Lincolnshire and north

Norfolk (Bowen *et al*. 1986, Clark *et al*. 2004, Hamblin *et al*. 2005, Gibbard *et al*. 2008 and 2009, Rose 2009).

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Figure 2.3: Extent of Anglian (MIS 12) and Devensian (MIS 5d-2) glaciations in southern England and Wales (after: McMillan *et al*. 2005)

The most widespread evidence is for Devensian (MIS 2) glaciation. Recent work (e.g. Boulton & Hagdorn 2006, O'Cofaigh & Evans 2007, Scourse *et al*. 2009, Clark *et al*. 2010, McCarroll *et al*. 2010) using geographic information systems (GIS), digital elevation modelling (DEM), digital terrain models (DTM) and a database of published dates has shown that at its maximum extent, this ice lobe reached the Scillies Isles and the continental shelf off southwest Ireland. Of relevance for the Bristol Channel/Severn Estuary region, the LGM maximum extent of the BIIS in south Wales was attained approximately 23 ka (Clark *et al*. 2010).

2.6.2 Extent of Pleistocene temperate stage deposits

Palaeontological evidence for Early Pleistocene temperate climates is found in the shallow marine and intertidal sands of East Anglian Crag deposits (Jones & Keen 1993, Funnell 1996, Stuart & Lister 2001, Lee *et al*. 2006, Parfitt *et al*. 2010), at Netley, Surrey, Rothamsted, Hertfordshire and Bridlington, Yorkshire, in terrace deposits of the River

Thames and cave deposits in Westbury-sub-Mendip, Somerset (Jones & Keen 1993, Bridgland 1994).

Palaeontological evidence for Middle Pleistocene temperate environments, attributed to MIS 20-13, is found in terrestrial and shallow marine deposits on the East Anglian coast, mainly around Cromer, Norfolk (Lee *et al*. 2006, Rose 2009, Gibbard *et al*. 2010, Parfitt *et al*. 2010, Preece 2010, Stuart & Lister 2010). Inland the most spatially continuous evidence is from the Cromerian (MIS 13) Valley Farm and Barham Soils (Jones & Keen 1993, Walker 2005). Deposits of comparable age, based on their stratigraphic position below Anglian (MIS 12) deposits, are found in Berkshire, Somerset, the Isle of Wight, Sussex, Devon, County Durham and Warwickshire (Jones & Keen 1993, Bowen *et al*. 1989, Bowen 1999a, Campbell *et al*. 1999), in river terrace deposits of the Thames and Solent rivers and as raised beach deposits at Boxgrove on the Sussex coast (Bridgland 1994, 2000, 2010, Bates *et al*. 2003).

Hoxnian (MIS 11) deposits are more extensive, although they are found mainly in East Anglia, east and southeast England and the Midlands, with analogous deposits in Northeast Scotland and West Wales (Jones & Keen 1993). Hoxnian (MIS 11) deposits are also found amongst river terrace deposits of the Thames Valley and its environs (Bridgland 1994, 2000, 2010, Rowe *et al*. 1999, Schreve 2001a, Bridgland & Westaway 2008, Roe *et al*. 2009). Evidence of raised beaches associated with former sea-levels during this stage is found on the Sussex coast (Jones & Keen 1993, Bridgland 2000, Bates *et al*. 2003). However, correlation of peat in Shetland with this stage is not considered to be secure (Bowen *et al*. 1986).

Most evidence for Wolstonian (MIS 10-6) interglacial environments is found in eastern England, with scattered evidence from the Midlands, the south coast of England, the Channel Islands and in Wales (Jones & Keen 1993, Keen *et al*. 1997). Temperate climate fluvial deposits from several sites within the Thames system, river terraces of Fenland rivers and the Severn-Avon system have been correlated with MIS 9 and 7, and from the River Trent with MIS 7 (Jones & Keen 1993, Bridgland 1994, 2000, 2010, Bridgland & Westaway 2008, Bridgland *et al*. 2004, Maddy *et al*. 1995, Murton *et al*. 2001, Roe *et al*. 2009). Raised beaches associated with former sea-levels and correlated with MIS 9 or 7 are found in southwest England, south Wales, southern England and northeast England (Davies 1983, Campbell & Bowen 1989, Campbell *et al*. 1998, Bridgland 2000, Bates *et al*. 2003, Davies *et al.* 2009), whilst cave deposits in Wales have been correlated with MIS 7 (Collcutt 1984, Bowen *et al.* 1985, Campbell & Bowen 1989, Campbell *et al.* 1998).

Ipswichian (MIS 5e) fluvial and marine deposits are the most widespread of the Pleistocene interglacial deposits. The majority of sites are in East Anglia, but they are also found across England, in South Wales and Northern Scotland (Jones & Keen 1993), although attribution of sites in Scotland to this stage is not considered secure (Bowen *et al.* 1986). Deposits from this stage are also found within the Thames and Severn-Avon river systems (Bowen *et al.* 1989, Bridgland 1994, 2000, 2010, Bridgland & Westaway 2008, Bridgland *et al.* 2004, Maddy *et al.* 1995, Murton *et al.* 2001). In south Wales extensive raised beaches and beach deposits within caves are correlated with this stage (Bowen *et al.* 1985, Stringer *et al.* 1986, Campbell & Bowen 1989); raised beaches correlated with this stage are also found in southwest and southern England (Campbell *et al.* 1998, Bridgland 2000, Bates *et al.* 2003).

2.7 Geomorphological record of British Pleistocene environmental change

Landform evidence for former cold and temperate environments provides important indications of palaeoenvironment and has long been recognised in the terrestrial stratigraphic record (Lowe & Walker 1997). Former glacial episodes are represented by glacial deposits, such as lateral, terminal, end and retreat moraines, outwash gravels and sandar, kame terraces, boulder spreads and drift limit moraines, eskers, nunataks, meltwater channels, tunnel valleys and ice-dammed lakes (Lowe & Walker 1997, Bowen *et al.* 2002, Walker 2005) all of which have been used to establish the extent of ice cover (Clark *et al.* 2004, Evans *et al.* 2005, Clark *et al.* 2010). Further evidence of ice, ice movement and flow geometry can be inferred from striations and crescentic gouges on stones and rock surfaces, stoss-and-lee landforms such as roche moutonées, drumlins and erratic dispersal paths (Lowe & Walker 1997, Clark *et al.* 2004, Evans *et al.* 2005, Clark *et al.* 2010). Deglaciation is characterised by terminal outwash gravels (Bowen *et al.* 2002) and kettle-holes, formed when ice blocks melt out in hollows (Gray 1991).

Glaciation was responsible for major drainage rearrangement, modifying pre-existing valleys to form glacial troughs and, where main valleys were glacier-deepened more than their tributary valleys, hanging valleys (Bennet & Glasser 1996, Lowe & Walker 1997, Benn & Evans 1998, Ahnert 1998). Drainage in the lower Severn Valley to the east of the main Severn axis was reversed to form the present southwest draining catchment of the Warwickshire Avon (Maddy & Lewis 2005) and the Early Pleistocene Thames formerly flowed to the north of London until this route was blocked by ice (Bridgland 1994, Rose *et al.* 2001). Moreover, many drainage basins not directly affected by glacier ice lay close enough to the ice margin to be indirectly influenced by glaciofluvial activity, reflected in increased discharge and sediment load (Lowe & Walker 1997).

In many river valleys complex terrace sequences formed in response to hydrological changes as a result of cyclic climate fluctuation (Lowe & Walker 1997, Maddy 1997, Bridgland 2000, Westaway *et al.* 2002, Bridgland *et al.* 2004), with episodes of incision concentrated during cold-warm transitions interspersed with aggradation of fluvially deposited sands and gravels during warm-cold transitions and within the cold stages themselves (Lowe & Walker 1997, van Huissteden *et al.* 2001, Maddy & Lewis 2005, Bridgland & Westaway 2008). Generally, cold stage fluvial systems tend to be high-energy, multiple channel (braided) whilst temperate stage rivers are low-energy, meandering and single channel (Maddy *et al.* 1998). The stratigraphy of the terraced sediments appears to support the timing of terrace aggradation and incision being principally driven by sediment/discharge changes in response to climate change (Bridgland 1994, Maddy & Lewis 2005). However, because river terraces are most frequently developed in unconsolidated sediments they are easily destroyed by subsequent fluvial action and consequently only preserved as fragments (Lowe & Walker 1997, Lewin & Macklin 2003); preservation of interglacial fluvial sediment sequences appears to be particularly fragmentary (Gibbard & Lewin 2002).

Most Pleistocene sedimentary evidence comes from the cold, periglacial, climates of land lying immediately beyond that directly affected by glacier ice (Campbell *et al.* 1999). Periglacial conditions prevailed immediately before the onset of glacial periods and in areas marginal to ice sheets (French & Karte 1988, Ballantyne & Harris 1994, Lowe & Walker 1997). Periglacial environments are characterised by conditions of severe winter ground cooling, continuous and discontinuous permafrost, aridity and frost weathering cycles (Eyles & Paul 1983, Ballantyne & Harris 1994, Lowe & Walker 1997, Campbell *et al.* 1999, Huggett 2003, French 2007). Sediments affected by periglacial action may reflect not only the breakdown of rock by cold-climate processes but also the re-working and redistribution of pre-existing drift deposits (Lowe & Walker 1997). Sparse vegetation cover facilitated accelerated wind erosion of widely exposed unconsolidated sediments resulting in deposits of loess, coversand and sand dunes. There was widespread mass wasting, fluvial erosion, accelerated solifluction, the formation of head deposits and the development of solifluction sheets, lobes and terraces (Eyles & Paul 1983, French & Karte 1988, Ballantyne & Harris 1994, Lowe & Walker 1997). Valley sides, especially in smaller valley systems, are often mantled with periglacial solifluction deposits which have had a smoothing and rounding effect on the topography (French & Karte 1988, Ballantyne & Harris 1994).

The formation of features unique to periglacial environments provides evidence for the former extent of the periglacial zone. These include indicators of permafrost such as pingo scars and thermokarst depressions, rock glaciers and protalus ramparts, whilst ice and sand wedges, cryoturbations, tundra polygons and large-scale sorted and non-sorted patterns are all features of the periglacial zone (Boardman 1987, Ballantyne & Harris 1994, Lowe & Walker 1997, Gurney 1998, Murton & Kolstrup 2003). However, periglacial features formed prior to the last cold stage are unlikely to have survived without some modification and are probably unrecognisable due to sediment redistribution (French 1976).

The stratigraphic record may also contain biological evidence, such as pollen or vertebrate remains, which are indicative of cold environments (Walker 2005). However, it is temperate stages and sub-stages that are primarily reflected in the fossil record (pollen, plant macrofossils, fossil insect remains etc.), or in biogenic sediments that accumulated in lakes and ponds during warmer climatic conditions (Walker 2005).

Coastal landforms, such as raised beaches and other littoral deposits that now stand above present sea level provide evidence for relative sea levels during the Pleistocene and may also contain fossils that provide data for detailed reconstruction of relative sea-levels (Lowe & Walker 1997). Dating deposits can be difficult, especially where fossils are absent, although recent technological developments (e.g. U-series, luminescence methods, ESR and AAR) have allowed age discrimination of the more recent beach deposits (Davies 1983, Bowen *et al*. 1985, Bates *et al*. 2003). Raised-beach deposits are traditionally thought to have formed under temperate conditions and represent high relative sea-level events (Bowen *et al.* 1985, Jones & Keen 1993, Allen 2001a, McCarroll 2002, Bates *et al*. 2003). This view was challenged by Eyles & McCabe (1989) who suggested that raised-beach deposits relate instead to complex patterns of glacio-isostatic subsidence and a rapid rise in sea level. However, this model for the LGM has been rejected for many sites on the south eastern shores of the Irish Sea Basin (Scourse & Furze 2001, Hambrey *et al.* 2001, O'Cofaigh & Evans 2001, 2007), as well as on theoretical grounds (Lambeck & Purcell 2001, Peltier 2002, Peltier & Fairbanks 2006) and consensus is now that deposition on the eastern margin of the Irish Sea is primarily of terrestrial origin (Thomas & Chiverell 2007). The conventional view of raised beaches as evidence for variations in sea level has also recently been challenged by Bridgland & Schreve (2009) and Westaway (2008, 2010b) who have argued that when their altitudes are related to current knowledge of global ice volumes from marine and ice cores, raised beaches can be explained by tectonic uplift.

Because of the effects of erosion the Pleistocene terrestrial stratigraphic record is highly fragmented and long, continuous records are rarely preserved (Walker 2005). Although evidence of the lateral extent of former Pleistocene environments may be determined on the basis of geomorphology, the incompleteness of British terrestrial Pleistocene records poses a problem for field mapping and interpretation (Lowe & Walker 1997, Benn & Evans 1998). Successive ice sheets have covered broadly the same areas, except at their margins, so much of the evidence of earlier glaciations has been destroyed by subsequent advances (Benn & Evans 1998). Many Pleistocene landforms, frequently developed in unconsolidated sediments, have been modified to some extent by slope processes or subsequent fluvial action and periglacial activity during and after regional deglaciation. In some cases evidence has been destroyed by meltwater activity during deglaciation or by postglacial erosion. Some environments may show no distinct geomorphological expression, whilst some evidence of former lower sea-levels now lies beyond the present coastline.

However, improved dating techniques and higher resolution stratigraphies provide a better understanding of when strata were deposited (Ehlers *et al.* 1991) and the chronology of terrestrial glacial-interglacial stratigraphies is increasingly being referred to the common time scale of the marine isotope record (Walker *et al.* 2001, Kukla *et al.* 2002, Gibbard & van Kolfschoten 2004). Even so, much of the evidence remains undated, correlations between individual MIS and the British terrestrial record are uncertain and problems in the designation of stratigraphic units and the development of an appropriate stratigraphic nomenclature remain (Lowe & Walker 1997, Walker *et al.* 1999, Clark *et al.* 2004, McMillan *et al.* 2005). The available evidence is of critical importance in reconstructing British Pleistocene landscapes; the evidence found within the Gordano Valley is discussed in Chapter 3.

2.8 Pleistocene environments in the Bristol Channel/Severn Estuary region

The lowland Pleistocene environments of the Bristol Channel/Severn Estuary region (boxed areas in Figure 2.4) provide a regional context for the Pleistocene evolution of the Gordano Valley. These coastal lowlands have been repeatedly subjected to the influence of both climate and sea-level changes (Walker *et al.* 2001); they therefore provide a valuable, albeit incomplete, sedimentary archive of evidence of ice advance and retreat, periglacial conditions and sea-level fluctuations (Lowe & Walker 1997).

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Figure 2.4: Location of lowland Pleistocene sites in the Bristol Channel/Severn Estuary region

Repetitive and widespread lithostratigraphical sequences containing dateable marker horizons in the superficial deposits of southwest England and south Wales are readily correlated (Bowen 1973a, 1999a and b). The application of AAR technology to mollusc fossils obtained from the widespread deposits of raised beaches and marine, intertidal and fluvial sediments has been a significant development in determining the age of pre-Devensian (pre-MIS 5d) events in the region and has been fundamental in assigning terrestrial sequences to the marine isotope framework (Campbell *et al.* 1998). However, difficulties still remain with assigning precise ages to older events. With the exception of AAR, advances in dating have been slow to be implemented in the Bristol Channel/Severn Estuary region.

A summary of key Pleistocene deposits and features in the Bristol Channel/Severn Estuary region, discussed in this section, is provided in Table 2.3. This provides a reference for environmental changes in the Gordano Valley.

2.8.1 Coastal landforms and deposits

Pleistocene variations in sea level have had considerable influence on the coasts of the Bristol Channel/Severn Estuary region. Gravels, sands and raised beaches of the Gower peninsula, Gwent Levels and coastal margins adjacent to the Gordano Valley (Table 2.3) have long been considered to provide evidence of former higher sea levels (Gilbertson & Hawkins 1977, Campbell & Bowen 1989, Briggs *et al*. 1991, Hunt 1998a and b) or sea levels approximately the same as at present (Allen 2000a).

Raised-beach deposits are widespread and are regarded as key sites for the establishment of Pleistocene chronology, vital to the interpretation of the overlying deposits (Bowen 1973a and b, Campbell *et al.* 1998). Many of these raised beaches lie on shore platforms (Allen 2001a), and are overlain by head or till. Whilst raised beaches provide a useful marker horizon that has been correlated across the region (Bowen 1999a), their age has long been the focus of debate (Campbell & Bowen 1989). Faunal evidence suggested raised beaches were temperate-climate deposits, probably of Ipswichian (MIS 5e) age and considered to pre-date Devensian (MIS 2) glaciation in the area (Campbell & Bowen 1989). However, lack of an agreed chronology (Kidson 1971) allowed controversies to develop, and because the precise age of raised beaches was an integral part of the interpretation and chronology of the sequences (Scourse 1991), this led to numerous disputes. Eventually techniques were developed that allowed the chronostratigraphic position of raised beaches to be determined, and regional correlations to be made (Campbell & Bowen 1989). AAR geochronology revealed the presence of raised beaches of more than one age (Andrews *et al.* 1979, Campbell & Bowen 1989); most have been correlated with the Ipswichian (MIS 5e) stage, a limited number have been correlated with MIS 7 or 9 (Bowen *et al.* 1986, Bowen 2005). Low Ham, Somerset, is the only site currently identified in Britain that demonstrates Devensian (MIS 5a) high sea-levels (Hunt 1998b, Hunt & Bowen 2006a).

Table 2.3: Summary of key Pleistocene deposits and landforms in the Bristol Channel/Severn Estuary region (sources: Andrews *et al*. 1984, Campbell & Bowen 1989, Campbell *et al*. 1998, 1999, Allen & Rippon 1997, Allen 2000a, 2001a and b, Lewis & Richards 2005, Findlay & Catt 2006, Hunt & Haslett 2006, Westaway 2010b). Uncertainties of evidence or age are signified ?

Timescale (ka)	MIS	Gower	Gwent	Somerset
c. 116 - 11.7	5d-2	Glaciofluvial outwash, head, coversands/loess, gravels	Head, cryoturbation, ice wedge casts, gravels	Head, coversands, fluvial gravels, alluvial fans
c. 125	5e	Raised beach	Raised beach, intertidal/marine sands & fluvial gravels	Raised beach, shore platform, intertidal sands & muds, palaeosol
c. 150	6	Cave earth ?Moraine		Head , fluvial gravels, alluvial fan
c. 200	7	Raised beach, marine sands		Marine sands, fluvial terraces, pedogenesis
c. 250	8	?Moraine		Braided rivers
c. 320	9		Gravels	?Fluvial gravel ?Estuarine silts & sands
c. 400	11			Fluvial gravels ?Estuarine silts & sands
c. 450	12	Moraine		?Till, glaciofluvial outwash
c. 600	15			?Estuarine silts & sands
c. 700	>15			?Till, glaciofluvial outwash

The Gwent Levels on the northern (Welsh) shore of the Severn Estuary between Cardiff and the River Wye provide evidence for an Ipswichian (MIS 5e) sea-level that was approximately the same as today. The present inner margin of the Gwent Levels records an extensive abandoned shoreline that contains littoral fauna of MIS 5e (Allen 2001b). Along the inner margin of the levels, extending 4 km or more inland, and around the flanks of Gold Cliff island is a discontinuous development of littoral shelly sands and gravels which mark an abandoned coast analogous to the raised beaches of Gower (Welch & Trotter 1961, Bell *et al.* 2000, Allen 2000a, 2001a and b, 2002, Allen & Haslett 2002). The flanks of Gold Cliff are covered by a calcite-cemented Ipswichian (MIS 5e) raised beach (Davies 2002, Bell *et al*. 2003) which grades laterally into locally shelly sands and gravels of probable shoreface origin. The deposits chiefly outcrop as a debris-covered ridge that curves away from the cliff (Allen 2000a). The uppermost 1-2 m of the raised beach deposit, widely exposed on the floor of a deep valley to the northwest of the ridge, mainly consist of yellowish, lightly cemented, well-laminated shelly sandstone. These deposits have yielded temperate water molluscan and foraminiferal assemblages closely resembling that of modern fauna in the Severn Estuary, indicating an interglacial episode (Haslett 1997, Allen 2000a) correlated by AAR with MIS 5e (Allen 2001b).

Correlatives in both age and character are found elsewhere on the Gwent Levels. Fossiliferous beach deposits at Llanwern, comprising sands and gravels with abundant temperate intertidal and rocky shore molluscs, suggest a beach close to an intertidal lagoon. The deposits have been correlated by AAR geochronology with the Ipswichian (MIS 5e) interglacial (Andrews *et al*. 1984, Allen 2001b). Similar deposits are found in the same context at nearby Bishton and Whitson. At Llandevenny, between Llanwern and Bishton, sands and gravels with a littoral-estuarine fauna are interbedded between Triassic bedrock and head deposits (Allen & Rippon 1997, Allen 2000a). Their temperate mollusc and foraminiferal assemblages resemble those of Gold Cliff and indicate littoral conditions (Allen 2001b). Stretching westwards along the inland margin of Caldicot Level concealed between Triassic bedrock and Holocene deposits are locally shelly littoral sands and gravels correlated with MIS 5e (Allen 2001b), whilst gravel at Magor Pill suggests the presence of MIS 5e beach or shallow intertidal deposits as described for Llanwern and Caldicot Level (Allen & Rippon (1997).

Evidence of higher sea levels is found on the Somerset coast where wide areas were flooded during an Ipswichian (MIS 5e) interglacial sea-level high stand (Allen 2002) and provides a possible analogue for the Gordano Valley. Estuarine sands and silts and marine and intertidal sands at nearby Kenn are thought to indicate more than one marine incursion (Gilbertson & Hawkins 1978a, Hunt 1998a, Bell *et al.* 2000, Allen 2001a), whilst the upper of two raised beaches at Swallow Cliff has been correlated with MIS 5e (Hunt 1998b, Campbell *et al*. 1999). In a recent reassessment of sea-level evidence Westaway (2010b) identified relative sea-level change in coastal north Somerset as the product of both eustatic and isostatic change. This allowed Westaway (2010b) to construct a chronology for raised beaches based on their inferred rate of isostatic uplift which largely confirmed the AAR geochronology.

The Burtle Beds of the Somerset Levels and Moors are also considered to be of marine provenance (Welch 1948, Findlay 1965, Kidson & Haynes 1972, Kidson *et al.* 1978, Bell *et al.* 2000, Hunt & Haslett 2006), although there has been much debate concerning their marine or glacial nature. An inferred marine origin for the sediments of the Burtle Beds (Welch 1948, Findlay 1965, Kidson & Haynes 1972, Kidson *et al.* 1974, Kidson & Heyworth 1976, Kidson *et al.* 1978, Bell *et al.* 2000, Hunt & Haslett 2006) was challenged by Hawkins & Kellaway (1973) who argued that they were glacial in origin. Such differences in interpretation of deposits have implications for determination of depositional environments of Pleistocene sediments in the Gordano Valley, where there are potentially glacial, marine and/or other environments, and secure attribution to a single depositional environment may not be possible.

2.8.2 Glacial landforms and deposits

Evidence for multiple glaciations is found on the Welsh coast of the Bristol Channel/Severn Estuary region, although this is fragmentary (Campbell & Bowen 1989, Evans 2005). Consequently, there is much uncertainty and continued debate about their type, extent and age (Harrison & Keen 2005). In particular, difficulties in dating glacial sediments, despite advances in technology, has placed reliance on the dating of underlying or overlying deposits in order to place a time frame on the glacial sediments (Harrison & Keen 2005). For example, the Paviland moraine, Gower, (Figure 2.5), has been variously attributed to glaciation during MIS 6, 8, or 12 (Bowen 1991, 1999b, 2005, Bowen *et al.* 1985, Campbell & Bowen 1989, Hiemstra *et al*. 2009).

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Figure 2.5: Pre-Devensian (MIS 6, 8 or 12) and Devensian (MIS 2) glacial limits on the Gower Peninsula (Campbell 1984, cited in Campbell & Bowen 1989, Bowen *et al.* 1985, Campbell & Bowen 1989)

Devensian (MIS 2) ice is assumed to have been confined to the north of the Severn Estuary (Jones & Keen 1993, Clark *et al.* 2004), all deposits in Somerset interpreted as glacigenic having been attributed to earlier (pre-MIS 2) glaciations (Bowen 1973a, Hunt 1998b, Evans *et al*. 2005). In keeping with the generally accepted age of the most southerly extent of ice when the deposits were first described, the age of the Somerset glaciation was originally assumed to be Wolstonian (MIS 6) (Hawkins 1972, Gilbertson 1974, Gilbertson & Hawkins 1978a) or earlier (Gilbertson & Hawkins 1978a, Andrews *et al*. 1984). The deposits were later assigned to either a MIS 10 or Anglian (MIS 12) glaciation (Hawkins 1977, Bowen *et al.* 1986, Jones & Keen 1993, Kellaway & Welch 1993, Keen 2001, Harrison & Keen 2005) and some have since been considered to be MIS 14 or 16 (Bowen 1991, Hunt 1998a). Deposits interpreted as till, possibly pre-MIS 15, are recorded at Kenn 4 km south west of the Gordano Valley, and may extend as far south as Greylake Quarry on the Somerset Moors (Hawkins & Kellaway 1971, Gilbertson & Hawkins 1978a and b, Hunt 2006a and e). Gravels interpreted as till and glacial outwash on the margins of the Gordano Valley (described in Chapter 3) have been correlated with this glaciation (Campbell *et al.* 1998, Hunt 1998a, Bowen 1999b, Campbell *et al.* 1999). Its age is uncertain, but it antedates MIS 5e and covered most of south Wales and undefined areas farther south (Bowen 1973a, 2005). At Kenn, gravel interpreted as till is overlain by younger interglacial deposits. Amino acid analysis of *Corbicula fluminalis* shells from these deposits indicates MIS 15 deposition (Andrews *et al*. 1984, Bowen *et al*. 1989), suggesting that the glacial deposits are MIS 16 or older. However, Hunt (2006h) advised that ratios obtained from *Corbicula fluminalis* are problematical, a view supported by Penkman *et al*. (2007). In their recent reappraisal of aminostratigraphy of the southern part of the North Sea Basin, Meijer and Cleveringa (2009) considered the AAR results for Kenn Pier and Yew Tree Farm to be aberrantly high; similarly high ratios have been reported for *Corbicula fluminalis* from Purfleet, Essex, which is assigned to MIS 9 on the basis of mammal biostratigraphy (Schreve 2001a). Meijer & Cleveringa (2009) suggest sampling was from the shell umbo, in which case the ratios are consistent with a MIS 9 age. Furthermore, the presence of *Corbicula fluminalis* is inconsistent with MIS 15 age as it is only known from pre-MIS 19, MIS 11, 9 and 7 deposits (Meijer & Preece 2000, Keen 2001, Meijer & Cleveringa 2009*)*. This indicates that the Kenn deposits are more likely to be MIS 11 or 9, which would place the glacial deposits in MIS 10 or 12 (Keen 2001, Harrison & Keen 2005, Westaway 2010b).

It has been suggested that at the height of this glaciation ice advanced eastwards up the Bristol Channel, affecting both sides of the Bristol Channel and impinging on the Somerset coast (Figure 2.6A) (Campbell & Bowen 1989, Ballantyne & Harris 1994), whilst ice derived from Wales may have blocked the Severn Estuary (Gilbertson 1974, Green 1992). Stephens (1970) suggested that this combination of ice, pressing southwards against the coast, may have formed a pro-glacial lake in lowland Somerset, the limits of which were controlled by the Bristol Channel ice front and the surrounding high ground. However, there is no unequivocal evidence for glaciation of the Mendips or south Somerset; consequently this scenario has been dismissed (Hunt *et al*. 1984, Farrant & Smart 1997, Hunt 1998b).

Despite there being little evidence for the limits of a pre-MIS 15 glaciation (Harrison & Keen 2005), Gilbertson & Hawkins (1978b) were able to infer its extent and direction of ice flow, and this is illustrated in Figure 2.6B. Their direction of ice flow agrees roughly with evidence for a glaciation of uncertain age, usually correlated with either Anglian (MIS 12) or Wolstonian (MIS 6) stages, found on the northern plateau of Lundy Island where there are extensive scatters of pebbles of erratic lithologies at 107 m above Ordnance Datum Newlyn (OD; the standard mean sea-level datum for Britain) and where west-north-west to east-south-east ice movement across the island has been inferred from ice moulded granite (Bowen 1973b, Harrison & Keen 2005).

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Figure 2.6: A. Inferred extent of Anglian glaciation in the Bristol Channel/Severn Estuary (dashed line) showing directions of ice movement (after: Campbell & Bowen 1989). B. Inferred extent and direction of ice flow of pre-MIS 15 glaciation of Somerset lowlands (Gilbertson & Hawkins 1978b). C. Inferred pre-MIS 15 ice movement across Somerset (Hawkins 1972). Arrows indicate assumed direction of ice flow

An earlier scenario, proposed by Hawkins (1972) and supported by Colbourne *et al.* (1974), whereby ice penetrated into Somerset from the south (Figure 2.6C) seems implausible, particularly if it is accepted that deposits from the Gordano Valley, interpreted as glaciofluvial and discussed in more detail in Chapter 3, indicate that Welsh ice crossed the Somerset coast from the north. Additionally, it is counter-intuitive to accept that ice moved towards the north from the lowlands of Somerset, although Rose (2009) has recently proposed a similar ice flow direction of west-southwest to east-northeast during Anglian (MIS 12) glaciation of northern East Anglia.

In Gwent, till and morainic deposits on the southwestern part of Wentlooge Level have been attributed to Devensian (MIS 2) glaciation (Allen 2000a), whilst the glacial drifts of north Gower contain a sequence of multiple till deposits dated to the Late Devensian stage, providing firm evidence of Devensian (MIS 2) ice encroachment (Campbell *et al.* 1982, Bowen *et al.* 1986, Campbell & Bowen 1989). Reappraisal of the key site of Rotherslade (Figure 2.5) has suggested that the long accepted idea of *in situ* basal till at the site is erroneous, and that the LGM ice limit should be repositioned eastward (Hiemstra *et al*. 2009).

2.8.3 Periglacial landforms and deposits

Much of the Bristol Channel/Severn Estuary region lies outside of the direct influence of the Devensian (MIS 2) ice sheet, but their proximity resulted in the widespread development of periglacial features roughly coincident in time with the advance and retreat of the ice sheet (French 1976, Campbell *et al.* 1998). These are characterised by solifluction deposits, breccias, aeolian sands, loess and alluvial gravels which occur widely across the region (Greenly 1922, Green & Welch 1965, Findlay 1965, Kidson 1971, Smith 1975, Gilbertson & Hawkins 1978a, 1983, Pounder & Macklin 1985, Macklin & Hunt 1988, Campbell & Bowen 1989, Hunt 1998b, Bell *et al.* 2000, Allen 2001a and b, Hunt 2006a, g and l).

Head and alluvial fan gravels are an important component of Pleistocene deposits in the region (Hunt & Haslett 2006); a Devensian (MIS 2) age is usually assumed for head deposits, although evidence of earlier cold stage (possibly MIS 6) slope deposits is found in Somerset (Hunt 2006l). On the Gwent Levels, head at Gold Cliff occurs as a discontinuous blanket that mantles the raised beach deposit and displays involutions, the sequence at Magor Pill is capped by head and on Caldicot Level head has been affected by periglacial processes. Head also commonly caps valley and sheet gravels (Allen & Rippon 1997, Allen 2000a, 2001b). In Somerset, extensive spreads of gravel occur opposite the gorges of Burrington, Churchill, Cheddar, Winscombe, Draycott, Wookey and Wells (Findlay 1965) and spread out fanwise across the floodplain beneath alluvium (Palmer 1931, 1934). Only the gravels of Burrington Combe and Wookey have been attributed to alluvial fan deposition; the remainder are considered to be head (Pounder & Macklin 1985, Macklin & Hunt 1988, Green 1992, Kellaway & Welch 1993), although Farrant & Smart (1997) have suggested that many of the Mendip 'head' deposits are probably alluvial fans. Findlay & Catt (2006) have identified two distinct periods of alluvial fan deposition at Burrington

Combe; one has been correlated with MIS 6 (Campbell *et al*. 1999) and pre-dates the interglacial Burrington Palaeosol, the younger period of deposition was attributed to Devensian (MIS 2) (Findlay & Catt 2006).

Cryoturbation features in drift deposits and ice-wedge casts in bedrock mark the extreme cold experienced regionally. On the Gwent Levels, intertidally exposed gravels with deeply entrenched surfaces capped by head are found at Sudbrook Point and display periglacial features including possible ice-wedge casts (Allen 2001b) and Triassic mudrocks and sandstones on the lower foreshore at Magor Pill are cut by periglacial ice-wedge casts (Allen & Rippon 1997).

2.8.4 Fluvial landforms and deposits

There appears to be little sedimentary evidence of Pleistocene fluvial conditions recorded for Gower. However, the bedrock surface of the Gwent Levels is dissected by pre-Ipswichian (pre-MIS 5e) valleys, created by glacial, marine, periglacial and fluvial processes. The valleys are plugged by head mantled gravels and some sands with a generally entrenched top. The entrenched valley-fill gravels were assigned by Allen (2001b) to the Devensian (MIS 5d-2) or Ipswichian (MIS 5e) stages; the valleys themselves probably antedate this stage (Allen 2001b).

In the Somerset lowlands, there is evidence of flashy, ephemeral, sand and gravel bedded cold stage valley floor streams. These resulted in substantial thicknesses of sands and gravel which interdigitate with, and pass laterally into, valley side head deposits (Campbell *et al.* 1998). Periods of aggradation alternated with periods of incision. Interglacial fluvial sedimentation was from small meandering streams and, although seldom preserved, deposits from these are also recorded in Somerset in MIS 9 and MIS 7 fluvial gravel stratigraphies (Hunt *et al*. 1984, Hunt 1998b, 2006k, Hunt & Bowen 2006b). Around the margins of the upland areas of Mendip, Somerset, in locations possibly analogous to Gordano Valley locations, large alluvial fans have been attributed to Devensian (MIS 5d-2) and/or earlier fluvial activity (Findlay 1965, Pounder & Macklin 1985, Macklin & Hunt 1988, Findlay & Catt 2006) and cold stage braided river sedimentation is correlated with MIS 8 (Hunt 1998b, 2006l). Chronological control on these deposits is provided by AAR geochronology of their molluscan fauna (Hunt *et al*. 1984, Hunt 1998b).

Fluvial karstic-system incision may reflect progressive lowering of the water table in response to lowering of base levels (Lowe & Walker 1997, Waltham *et al*. 1997, Lewin & Gibbard 2010) which for the Mendip Hills approximates to sea-level (Smith 1975). More recently, Westaway (2008, 2010b) advocated tectonic uplift in the Mendip Hills around Cheddar as a driver for fluvial karstic-system incision, and suggested uplift of ~150 m since the Early Pleistocene, ~100 m since the Anglian (MIS12) glaciation and ~14 m since Ipswichian (MIS 5e) time (Westaway 2010b). Probable MIS 7 and MIS 5e fluvial and intertidal deposits at Weston in Gordano in the Gordano Valley, which lie at ~14 m OD, have been linked to this uplift (Westaway 2010b).

2.8.5 Palaeontological evidence

The Bristol Channel/Severn Estuary region contains fossiliferous sequences of marine interglacial and interstadial deposits and important fossilferous non-marine sites (Hunt 1998b). In particular, caves of the Gower Peninsula, the Mendip area of Somerset and north Somerset have revealed both temperate and cold stage mammal fauna, some of which has been dated by OSL and TIMS and correlated by AAR geochronology (Day 1866, Reynolds 1907, 1934, Stringer *et al*. 1986, Campbell & Bowen 1989, Gilmour *et al*. 2007). Cold taxa molluscan assemblages, fossil pollen and vertebrate fauna, variously attributed to MIS 8, 6 and 5/4 have been recorded from across the region (Hunt 1998b, Currant *et al*. 2006, Hunt 2006a and g). Regional examples of temperate assemblages include marginal-marine fauna attributed to MIS 9, 7, 5e, 5a and possibly MIS 15 (Gilbertson & Hawkins 1978a, Haslett 1997, Hunt 1998a and b, 2006c, h & k, Bell *et al*. 2000, Allen 2001b, Hunt & Bowen 2006a).

Pleistocene environmental change in Britain and the framework into which this thesis fits has been introduced. A synopsis of evidence for Pleistocene environmental change in the Bristol Channel/Severn Estuary region has been presented, providing a regional context for the Pleistocene evolution of the Gordano Valley for which a detailed review is presented in Chapter 3.